Studies in Maine Geology

Volume 3: Igneous and Metamorphic Geology

Edited by Robert D. Tucker and Robert G. Marvinney

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DEPARTMENT OF CONSERVATION

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1989

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Cover photograph:

Several concordant Mesozoic dikes in the Kittery Formation as exposed south of Bald Head Cliff near York Beach, Maine. See "Mesozoic dikes of southern coastal Maine: an historical perspective and update of research, 1838 - 1988" by M. T. Swanson, this volume. Photograph by M. T. Swanson.

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Published under Appropriation 4501-2300

Studies in Maine Geology

Papers to commemorate the 150th anniversary of
C. T. Jackson's reports on the geology of Maine

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C. T. Jackson's reports on the geology of Maine

Charles Thomas Jackson (1805-1880)

Foreword

During the period 1836-1839, Charles Thomas Jackson conducted the first comprehensive, government-funded survey of the geology of the state of Maine. As part of our commemoration of the 150th anniversary of that monumental undertaking, the Maine Geological Survey is publishing *Studies in Maine Geology*, a series of 6 volumes which cover a broad spectrum of geological investigations. The response to our initial call for papers was overwhelming not only in number, but also in the quality of contributions, which would have been well received by professional journals. I appreciate the efforts of each contributor in ensuring the success of these volumes.

Many issues have challenged the geologic community since the time of that first survey. Jackson was charged with the seemingly insurmountable task of surveying the geology of the entire state over a period of 3 field seasons without the aid of accurate maps or modern equipment. One of the main objectives of his work was to assess the potential for mineral deposits, coal, and building materials in this unexplored territory. In the latter part of the 19th century the geological issues of the state concerned sources of dimension stone and pegmatite gemstones. During this period, the gold and silver rushes accelerated the pace of geologic exploration in Maine, causing a short-lived metal-mining boom. In the first half of the 20th century and with the advent of two world wars, the need for "strategic minerals" once again stimulated geologic investigation. The economic downturn in the domestic minerals industry of the past few decades has reversed and interest in exploration and extractive commodities is once again rapidly gaining ground in the state.

Today, a growing environmental awareness has led to a concern for protecting our resources, both now and in the future. Geologic information provides the basis for a multitude of decisions aimed at confronting the complex problems of modern society. Primary issues are those of ground water resources and contamination, coastal development and shoreline protection, and the disposal of nuclear wastes in geological repositories.

To meet the changing and expanding needs of the state over the past century and a half, the Maine Geological Survey has been under the auspices of many different government departments, culminating in its current position within the Department of Conservation. Since Jackson's time, the survey has expanded from essentially a one-man agency to include full bedrock and surficial geology, marine geology, hydrogeology, and cartographic divisions. In spite of the changes in issues and manpower, one of the primary objectives of the Maine Geological Survey has remained unchanged since Jackson's time: to provide the public with the highest quality information about the geology of the state of Maine. These volumes meet the challenge of that objective.

Walter A. Anderson
Director and State Geologist
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DEPARTMENT OF CONSERVATION
Augusta, Maine

Acknowledgments

Studies in Maine Geology began as a project to commemorate 150 years of geologic investigation in Maine. The original intent was to assemble a volume of papers which summarized our current understanding of all aspects of the geology of Maine. As word of the project spread, however, the volume grew into a six-volume set centered around three broad subject categories: structure and stratigraphy, igneous and metamorphic geology, and Quaternary geology.

The publication of a project of this size involves the time and goodwill of a great many people. The editors thank Marc Loiselle who initiated the project and contacted the authors. The illustrations and manuscript assembly were done by the Cartography and Information Services Division of the Maine Geological Survey. Additional editorial assistance was provided by Steve Dickson, Woody Thompson, Tom Weddle, and John Williams. Particular thanks are given to Ben Wilson and John Poisson who drafted many of the illustrations and to Cathy Stultz and Cheryl Fiore who typed and corrected the manuscripts. We also wish to thank the geologists who willingly undertook the task of reviewing the manuscripts.

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Maine Geological Survey
Studies in Maine Geology: Volume 3
1989

Metamorphism in Maine: An Overview

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ABSTRACT

The metamorphic events that have affected rocks in Maine range in age from Precambrian to Carboniferous and possibly Permian. The tectonic settings and pressure-temperature conditions of these events are variable. Most of the high-grade events involved low-pressure, high-temperature conditions. Recrystallization was post-tectonic, and the heat was convected in by intruding magma. For these high-grade events the pressures range from 1-2 kbar in the contact aureoles of the north-central part of the state to as much as 4 kbar in the regional metamorphism of the southwestern third. In addition, two small areas of high-grade Barrovian metamorphism have been identified in southern Maine.

Broad-scale, low-grade metamorphism affects the entire northern two-thirds of Maine. By analogy with the pressures determined in contact aureoles, much of this low-grade metamorphism must have formed at 1-2 kbar of pressure. In addition, low-grade metamorphism having possible affinities with subduction zone and ocean floor types of metamorphism may be present in a northeast-trending belt extending across the north-central part of the state.

In a chronological context, the following events have been recognized. (1) High-grade Precambrian metamorphism occurs in the Chain Lakes massif and on a few islands at the north end of Penobscot Bay. In both cases, the rocks may be allochthonous such that the metamorphism occurred elsewhere. (2) Cambrian metamorphism is difficult to demonstrate unequivocally. Low-grade effects may have occurred throughout much of the northern part of the state, and some may have involved subduction zone tectonics and sea-floor metamorphism. (3) Low-grade Ordovician metamorphism also may have affected much of the northern part of Maine, but as with the Cambrian metamorphism, unequivocal demonstration is difficult. Elsewhere in the state it appears that low-grade Ordovician metamorphism occurred in the area to the northeast of Penobscot Bay and on some islands in the Bay. Possibly some high-grade Ordovician metamorphism occurred in a belt along the Norumbega fault zone in the south-central part of Maine. (4) Siluro-Devonian metamorphism affected virtually all of the state. In the north and east it involved low-grade regional metamorphism and narrow, high-grade contact aureoles around plutons. In the southern third of Maine the Siluro-Devonian metamorphism was regional and attained high grades. Nonetheless, the cause of the high temperatures involved heat convected in by intrusion of sheet-like granitic plutons. (5) High-grade Carboniferous metamorphism was superimposed on earlier high-grade Siluro-Devonian metamorphism in the southern third of the state but its exact areal extent is yet to be established. As with the Siluro-Devonian high-grade metamorphism, intruding granitic plutons provided the heat required for recrystallization. In addition, some low-grade Carboniferous metamorphism occurred in the northeast part of Maine. (6) Possibly some Permian metamorphism occurred in the south-central part of Maine, but this suggestion is still highly tentative.

The above described chronology of metamorphism in Maine applies most rigorously to the area west of the Norumbega fault zone and east of the Turtle Head fault zone. In the intervening strip, our understanding of the metamorphic events is much more tenuous. However, it is likely that this strip was affected at least by the same Siluro-Devonian events and history as described above.

Probably only the Precambrian, Siluro-Devonian, and Carboniferous metamorphic events are sufficiently well understood that they can be readily integrated into models aimed at a plate tectonic synthesis of the northern Appalachians.

INTRODUCTION

Metamorphism is an intrinsic aspect of orogeny whether considered in classical terms or in the context of more recent plate tectonic models. Hence, understanding an orogeny requires consideration of its associated metamorphism. It is now well recognized that metamorphic style is intimately related to the tectonic style of an orogen and also provides information on an orogen's thermal evolution.

For example, the relationship between subduction zone tectonics and the occurrence of blueschist style metamorphism is now well known (see reviews by Ernst, 1975, and Miyashiro et al., 1982). There have also been recent attempts to use differences in metamorphic style and in deformation/recrystallization relationships for distinguishing accreted terranes (see Zen, 1985). And, as long ago as the papers by Zwart (1967a,b), it was recognized that extensive areas of low-P, high-T metamorphism were intimately associated with extensive granitic batholiths. More recently, thermal modeling approaches (Lux et al., 1986) have led to suggestions that much low-P, high-T regional metamorphism is a direct product of heat *convected* in by emplacement of the batholiths. In the absence of the plutons, only greenschist facies recrystallization would have occurred in the given region.

Orogenesis and plutonism in Maine involves a long, complex history ranging from Precambrian to at least Carboniferous times. In some cases the metamorphism has been intimately interrelated with other facets of the orogenic activity. This is mainly so only for the Siluro-Devonian (Acadian) regional metamorphisms of the southwestern portions of the state and the Siluro-Devonian, purely contact metamorphism around the scattered plutons that intrude the low-grade rocks throughout much of the northern and eastern parts of Maine. Indeed, references to metamorphism in Maine are generally with respect to these metamorphisms because they include the most pronounced recrystallizations that have affected the rocks of the state. Reflecting this, the metamorphic map that accompanies the new Bedrock Geologic Map of Maine (Guidotti, 1985) is essentially a map of Siluro-Devonian metamorphisms. In addition to this map, these metamorphisms in central and western Maine have been reviewed recently by Holdaway et al. (1982), Guidotti et al. (1983), and Holdaway et al. (1986).

Several other metamorphic events have also affected rocks in various parts of Maine, but information on these events is sparse and scattered throughout the literature. Hence, a special effort is made herein to provide descriptions of these other metamorphic events. In the context of the extensive literature about the Siluro-Devonian metamorphisms, the review given below will be relatively brief and will emphasize ideas developed since the reviews of Holdaway et al. (1982) and Guidotti et al. (1983). Because these events are the strongest and cover so much of the state, they must be considered first inasmuch as the other metamorphisms are detected by "looking through" the Siluro-Devonian metamorphisms or looking at effects superimposed on them. These other metamorphisms will be treated chronologically with the exception of the metamorphism affecting the east-central portion of Maine. The timing of the metamorphism(s) there is not well known, and so it is not feasible to place it into a specific chronological position.

The overall aim for each event is to provide metamorphic information that will be useful to workers synthesizing the orogenic evolution of the Northern Appalachians. The emphasis herein will be on "factual or observational aspects" rather than on synthesis. It is realized that continued work will change even the "factual or observational" as presented below. However, it is hoped that the emphasis given will extend at least somewhat the half-life for which the material presented has some validity or worth to other workers.

SILURO-DEVONIAN METAMORPHISM

As shown on Figure 1, Siluro-Devonian metamorphism involves regionally developed recrystallization with grades ranging from sub-greenschist facies to well up in the upper amphibolite facies. It also involves simple contact effects around the Siluro-Devonian plutons that intrude the greenschist and sub-greenschist facies rocks which underlie the central, eastern, and northern portions of the state. These contact effects appear to be straightforward responses to very localized heat sources emplaced at high structural levels. This is indicated by the development of narrow hornfels zones around the plutons, the common occurrence of andalusite and cordierite-bearing parageneses, and the relative scarcity of staurolite- and garnet-bearing assemblages, (e.g. Moore, 1960; Harwood and Larson, 1969; Tewhey, 1975). Both the regional and contact metamorphic effects appear to be manifestations of the thermal aspects of the Acadian orogeny.

High-grade Siluro-Devonian metamorphism is recorded in four northeast-plunging lobes (shown on Fig. 1) that form the northern terminus of an extensive Acadian high-grade terrane extending for several hundred miles to the southwest (Thompson and Norton, 1968). The three westernmost lobes lie mainly in the Kearsarge-central Maine synclinorium, but also extend across the Bronson Hill anticlinorium in New Hampshire, and the Lobster Mountain and Boundary Mountains anticlinoria

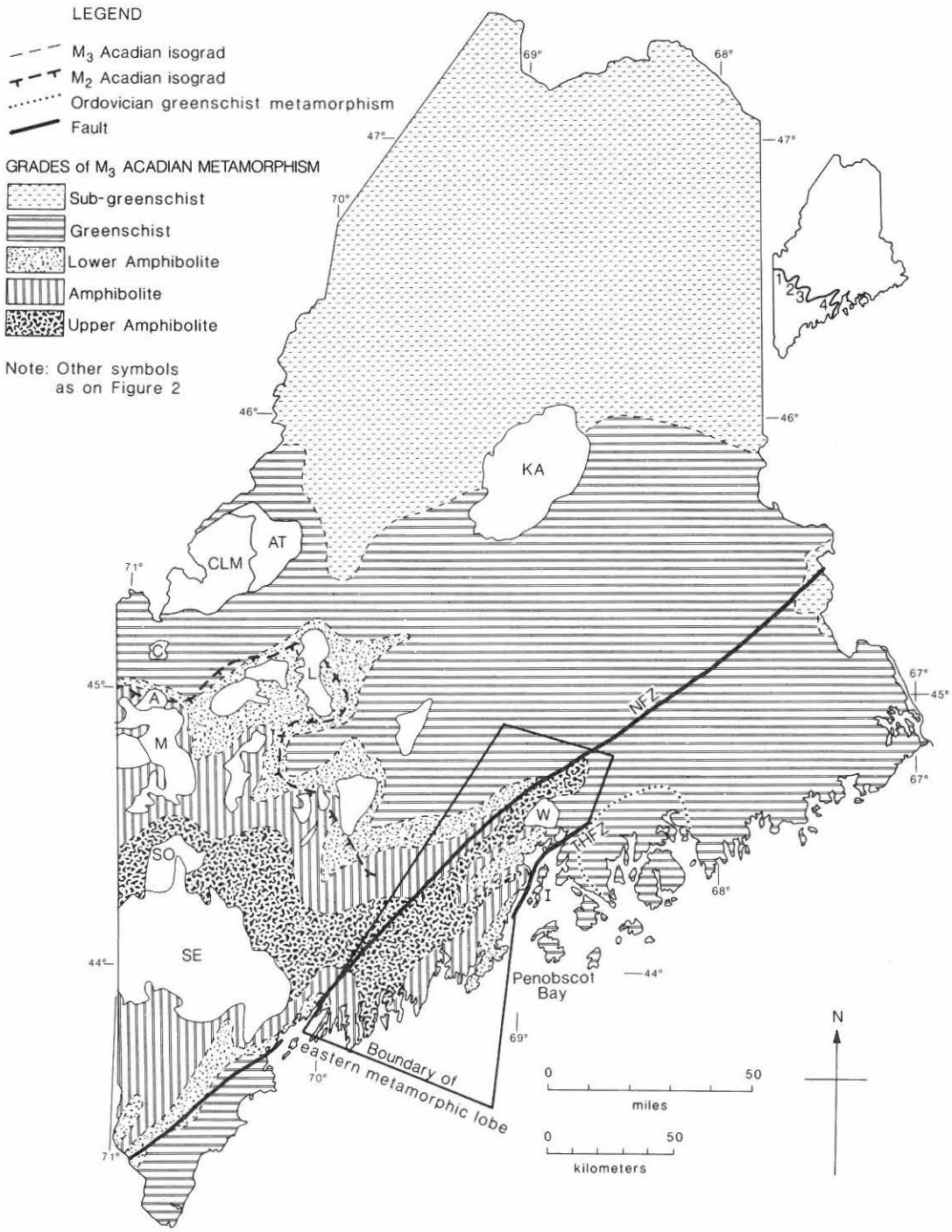

Figure 1. Summary of Acadian metamorphism and coastal area Ordovician metamorphism as modified from the state metamorphic map compiled by Guidotti (1985). High-grade side (AFM biotite + andalusite compatiblilty) of M_2 Acadian isograd shown by hachures. Inset map shows four metamorphic lobes discussed in text.

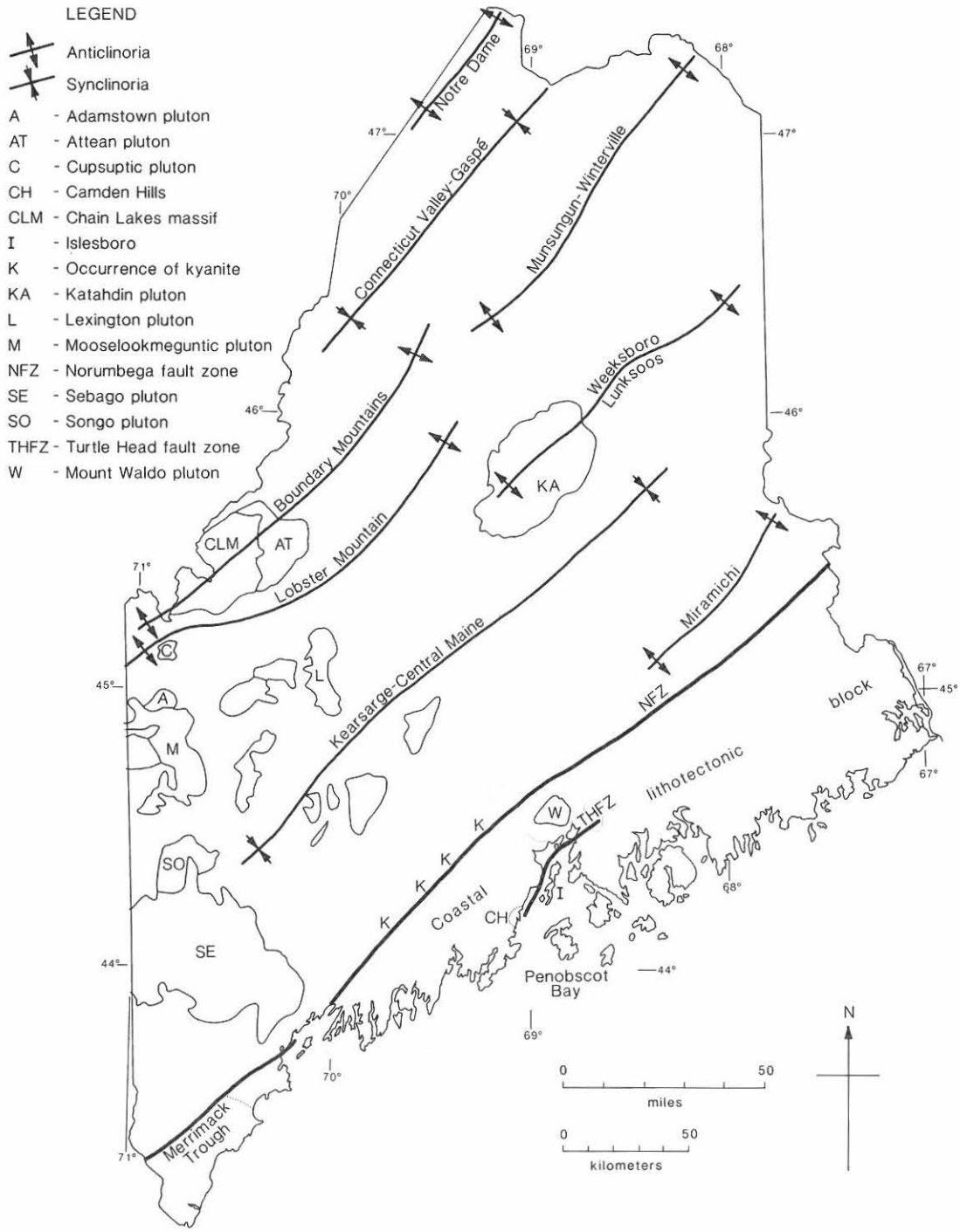

Figure 2. Location of plutons, major structures, mineralogic and other geologic and geographic features mentioned in the text. Based on tectonic map of Maine, Osberg et al. (1985).

(Fig. 2). The easternmost metamorphic lobe occurs largely in the coastal lithotectonic block (see Figs. 1, 2). Although it is shown as resulting from Siluro-Devonian metamorphism, this timing is quite problematic and will be discussed in detail below.

From Holdaway et al. (1982) and Guidotti et al. (1983), the following generalizations can be made about the three western lobes:

(a) The Siluro-Devonian regional metamorphism involves several largely post-tectonic recrystallization events that overlap in space and time. Due to these overlaps, one commonly finds areas in which an earlier event is overprinted by a later event. This results in prograded rocks in some areas and retrograded rocks in others. A textural result of such overlaps is the development of numerous pseudomorphs (prograde and retrograde; see Guidotti, 1970).

(b) In general, it appears that two dominant events (M_2 and M_3) can be defined. Both involve grades high enough to establish the AFM join connecting Al-silicate (andalusite or sillimanite) and biotite (i.e., amphibolite facies). Hence, they are relatively low-P, andalusite/sillimanite types of metamorphism. They were preceded by an earlier event (M_1) which attained only greenschist facies and may have been syntectonic.

(c) M_3 probably occurred at slightly higher P than M_2 (0.5 to 1 kbar increase). Holdaway et al. (1982) and Guidotti et al. (1983) suggested that the non-tectonic increase in P was possibly due to extrusion of a few km of volcanic rocks. (See below for an alternative suggestion.)

(d) Relationships between prograde and retrograde minerals suggest that the rocks at a given locality cooled to ambient T between M_2 and M_3.

(e) The metamorphic grades of the western lobes are clearly related spatially with the distribution of Devonian plutons (see Fig. 1). In most places the isogradic surfaces appear to have gentle dips, thereby forming the broad, NE-plunging lobes.

(f) In some cases the Siluro-Devonian metamorphism has strongly recrystallized earlier plutons. One example involves the Ordovician Adamstown granite (Fig. 2). Due to the thermal effects of the Mooselookmeguntic pluton, some parts of the Adamstown pluton are in the greenschist facies and other parts in amphibolite facies, (Green and Guidotti, 1968; and Guidotti, 1977). Another striking example is the Siluro-Devonian Plumbago pluton, a small mafic-ultramafic layered complex which occurs just southeast of the Mooselookmeguntic pluton. It has been completely recrystallized to amphibolite facies (Moody, 1974).

More recent work suggests the following additional generalizations or modifications of our understanding of the Siluro-Devonian regional metamorphism of Maine.

(a) Isotopic studies by Hayward and Gaudette (1984), Aleinikoff (1984), and Aleinikoff et al. (1985), and isotopic plus petrologic work by Lux and Guidotti (1985) strongly suggest that the highest grade rocks in the western lobes (upper amphibolite facies, Fig. 1) are due to the thermal effects of the Carboniferous Sebago pluton (see below).

(b) Lux et al. (1986) have developed quantitative models showing that the regional Siluro-Devonian recrystallization is actually deep level (12-16 km) contact metamorphism. Its large areal extent results from the gently dipping, tabular nature of the granitic plutons (the heat sources). Thus, the high-grade metamorphism is directly due to *convective* heat transfer. In the absence of the plutons, the metamorphic grade would not have exceeded greenschist facies. In this sense, the *geologic* significance of this metamorphism decreases because the important question becomes the source of heat to produce the magmas at greater depths.

(c) In the context of plutons serving as heat sources for the regional metamorphism, Lux and De Yoreo (1987) have suggested an alterative means for the above-mentioned, non-tectonic P increase between M_2 and M_3. Specifically, emplacement of sheet-like plutons above the now exposed rocks would cause an increase in P just as readily as would extrusion of a blanket of volcanic rocks. Of course, both may have occurred.

(d) $^{40}Ar/^{39}Ar$ work on the northern parts of the Mooselookmeguntic pluton (Lux and Guidotti, 1985; Lux et al., 1986) strongly support (verify?) the above-mentioned cooling to ambient T between M_2 and M_3. Hornblende cooling ages there have been found to be essentially identical with the crystallization age of the pluton reported by Moench and Zartman (1976) (corrected for new decay constants). Hence, the pluton (i.e. the heat source for M_3) must have intruded cold M_2 metamorphic rocks, and then, along with the M_3 metamorphic rocks, cooled below 500°C closure T of hornblende in a matter of a very few million years.

(e) As shown by Lux et al. (1986), the metamorphism associated with the sheet-like Siluro-Devonian plutons was nearly isobaric. In terms of diagrammatic PT space, the metamorphism occurred essentially as a geologically instantaneous thermal spike resulting in prograde reactions. De Yoreo et al. (this volume) present a similar model for Carboniferous metamorphism associated with the Sebago batholith.

Given such a thermal history, the recrystallization history is relatively simple and closely approaches equilibrium. In this respect, such low-P, high-T metamorphism contrasts strongly with recrystallization along the P-T-time (P-T-t) loops commonly believed to occur during deeper- level Barrovian metamorphism in which convective heat transfer is presumably negligible (e.g. see England and Thompson, 1984).

(f) The extensive area of low-grade rocks to the northeast of the high-grade lobes may be a manifestation of the same early heating event (M_1) that produced only greenschist facies rocks in the area now affected by the higher grades of M_2 and M_3. This remains an open question.

On the other hand, it now appears that the transition from the high-grade, regional Siluro-Devonian metamorphism northeastward to the low-grade areas containing plutons surrounded by narrow contact aureoles involves a change in pluton geometry. As earlier stated, the plutons associated with the high-grade areas occur as abundant, gently dipping sheets. In

contrast, the plutons intruding the low-grade rocks to the northeast seem to be more nearly equant in shape (e.g. Moore, 1960; Tewhey, 1975; Hodge et al., 1982; and Carnese, 1981). Lux and De Yoreo (1987) argue that from southwest to northeast, the rocks now exposed at the earth's surface are from successively higher structural levels. Presumably, in conjunction with the resultant systematically varying ambient T and P, this would influence the shapes attained by the intruding plutons, and hence, the nature of the resultant contact metamorphic effects.

In a few places the transition of the plutonic and metamorphic style is quite abrupt along a southwest to northeast traverse. For example, at the northern end of the Mooselookmeguntic pluton, the contact of the pluton locally steepens and coincidentally the dip of the regional metamorphic isograds becomes steeper. In map view the isograds are more closely spaced (Fig. 1). The pluton there is probably not a gently dipping sheet extending outward beneath the metasedimentary rocks. Nonetheless, the metamorphic rocks do not become hornfelsic but retain a strong foliation and have assemblages which include staurolite, garnet, and sillimanite.

In contrast, only 15 km northward is the granite of the Cupsuptic pluton which intrudes the same strata as the Mooselookmeguntic pluton, but in an area of regional greenschist facies metamorphism. A very prominent, narrow contact hornfels has developed and involves assemblages containing andalusite, sillimanite, and abundant cordierite (Harwood and Larson, 1969). Garnet and staurolite seem to be absent. Following the suggestions of Lux and De Yoreo (1987), one can speculate that the Cupsuptic pluton represents a high-level apophysis of the Mooselookmeguntic or some other sheet-like pluton present at greater depths below the Cupsuptic area.

Finally, it should be noted that on Figure 1, the low-grade metamorphism in the northern part of the coastal lithotectonic block is shown as continuous with the extensive Siluro-Devonian, low-grade terrane covering much of northern Maine. The work of Ludman et al. (in press) tends to support such an interpretation. However, as noted below, some doubt exists about such an interpretation for several other areas in the coastal lithotectonic block. Another interesting implication from the work of Ludman et al. (in press) is that the erosion level of the plutons which intrude the low-grade portions of this block (at its northern end) may be somewhat deeper (equivalent to 3 kbar) than that of plutons intruding the low-grade portions of the Kearsarge-central Maine synclinorium (equivalent to 1 to 2 kbar).

Metamorphism of the Eastern Metamorphic Lobe

In the context of rocks at grades mainly higher than greenschist facies, the eastern metamorphic lobe is designated as the area covering much of the central Maine coast, Figure 1. Although this lobe crosses the Norumbega fault zone and covers part of the southeastern limb of the Kearsarge-central Maine synclinorium, it mainly covers portions of the coastal lithotectonic block, (see Figs. 1 and 2). As shown by Guidotti (1985), the northern end of the high-grade rocks of the lobe appears to be largely fault-bounded. However, this may in part reflect lack of detailed work to date. Another prominent fault boundary associated with the eastern metamorphic lobe is the Turtle Head fault zone (Figs. 1 and 2). This fault occurs as a north-northeast-trending structure in the western portion of Penobscot Bay (Stewart, 1974; Stewart and Wones, 1974). It appears to form a sharp boundary to the eastern metamorphic lobe (see further discussion below).

In comparison with the three Siluro-Devonian metamorphic lobes described above, relatively little petrologic study has been done on the eastern metamorphic lobe. The only studies dealing at least in part with the metamorphic aspects of the geology include Bickel (1971, 1974), Wones (1974, and pers. commun.), Stewart (1974), Stewart and Wones (1974), Hathaway (1969), Osberg and Guidotti (1974, and in prep.), Guidotti (1979), and Berry (1986). In addition, Stewart (1974) has described the metamorphic rocks to the east of the Turtle Head fault zone, and Stewart and Flohr (1984) have presented detailed petrologic and mineralogic data for the rocks exposed on the east side of Penobscot Bay (strictly speaking, not parts of the eastern metamorphic lobe as defined above). Excluding this latter study, most of the work cited above is only reconnaissance in its approach to the study of metamorphism, and none has been published that involves the kind of detailed work that characterizes much of the effort on the three western lobes.

Metamorphism in the eastern metamorphic lobe may involve recrystallization events ranging in age from Precambrian (treated separately in the next section) to as young as Permian. However, Figure 1 shows the lobe as Siluro-Devonian (Acadian) metamorphism -- mainly for the sake of simplicity rather than certainty. For example, in at least a general way the grades of metamorphism appear to be directly continuous with those in the western lobes. Moreover, in a descriptive way, several features of the metamorphism in the eastern metamorphic lobe are similar to those in the western lobes. (a) The range of grades is nearly identical and commonly involves static recrystallization. (b) Evidence for polymetamorphism is very strong, and in many cases results in the formation of various types of prograde and retrograde (depending upon location) pseudomorphs which appear to have formed in a static environment. (c) Most of the lobe involves andalusite/sillimanite pressure conditions. (d) At least some of the metamorphism affects rocks believed to be of Siluro-Devonian age. (e) At the northern end of the lobe, some Siluro-Devonian plutons have clear contact aureoles which are superimposed on the regional metamorphism. For example, Stewart (pers. commun., 1986) has observed staurolite replaced by cordierite in the contact aureole of the Devonian Waldo pluton (Fig. 1).

Because of these similarities, and in the absence of strongly contradictory evidence, it was most reasonable to designate the eastern metamorphic lobe on the state metamorphic map as part of the Siluro-Devonian regional metamorphism. Thus, on Fig-

ure 1 the high-grade rocks of the eastern metamorphic lobe are also shown as a broad easterly hook of the extensive high-grade terrane described by Thompson and Norton (1968) as well as forming another northeast-trending lobe in Maine. This broad easterly hook of high-grade rocks (up to upper amphibolite facies) that forms the core of the eastern metamorphic lobe covers hundreds of square miles and so is not a minor, incidental perturbation of the extensive terrane described by Thompson and Norton (1968).

Some of the metamorphism in the eastern metamorphic lobe affects pre-Silurian rocks (e.g. Boucot et al., 1972), but it has never been demonstrated that all of this metamorphism also affects Siluro-Devonian strata as well. Hence, some of the metamorphism in the lobe could be pre-Silurian in age. Moreover, in places there appear to be significant differences between the metamorphic style in the eastern metamorphic lobe and that in the three western lobes. These include the following points: (1) The pattern of isograds is much more elongate in map view. This may involve steep horizontal metamorphic gradients or the effects of later faulting (see below). However, it should be emphasized that the isograd patterns shown are largely based on reconnaissance work. (2) There is little evidence that the high-grade regional isograds have a spatial relationship with plutons. (3) In contrast with the three western lobes, there is evidence that isograds are offset by faults on both a local and regional scale. As shown by Guidotti (1985), this includes: (a) offsets by the Norumbega fault zone (Fig. 1), (b) a fault bounding the north end of the highest grade rocks in the lobe, and (c) the Turtle Head fault zone in Penobscot Bay which separates relatively high-grade rocks on the mainland from low-grade rocks a few kilometers offshore on Islesboro. On a more local scale, Bickel (1971, 1974, 1976) shows the St. George fault offsetting presumed Acadian age isograds in the east-central portion of the eastern metamorphic lobe. (4) Some kyanite-bearing parageneses occur in this lobe (see below). (5) Although most of the abundant pseudomorphs developed are post-tectonic, some are clearly not so as seen from their deformed shapes. Good examples of deformed pseudomorphs are common in the southeast portion of the lobe (in the Tenants Harbor 7-1/2' quadrangle, Guidotti, 1979).

With these comparisons and contrasts between the eastern metamorphic lobe and western lobes as background, several additional generalizations can be made about the former.

Polymetamorphism is extremely prominent in the eastern metamorphic lobe and is predominantly post-tectonic. However, the number of events, their timing, and their spatial distribution are largely unknown. In addition, there is little idea whether the polymetamorphism involves discrete, essentially isobaric events (as in the western lobes, Lux et al., 1986) or continuums along protracted P-T-t paths (England and Thompson, 1984).

The occurrence of kyanite in the eastern metamorphic lobe was first reported by Pankiwskyj (1976). Additional reconnaissance petrographic work by the writer (in conjunction with mapping by Newberg, 1985, 1986) suggests that the kyanite-bearing rocks are restricted to a narrow zone about 35 km long and lying just west of the Norumbega fault zone (Fig. 2). The kyanite occurs in a thin pelitic unit within the Cushing Formation (as designated by Osberg et al., 1985).

The paragenesis consists of kyanite + staurolite + biotite and appears to be the result of a pre-Acadian metamorphism. The kyanite is now much resorbed (to elongate, rounded anhedra) and the staurolite is partially replaced by cordierite. Abundant, fresh fibrolitic to coarse sillimanite is present in the same specimens. It would appear that an early kyanite + staurolite type of metamorphism has been overprinted by the Siluro-Devonian age Buchan-type metamorphism that is well known (Osberg, 1974) in the Augusta-Waterville area, about 10 km west of the belt of kyanite-bearing rocks. Although the age of the kyanite-staurolite metamorphism is presently unknown, it seems to be present only in the pre-Silurian Cushing Formation. In addition, H. Gaudette (pers. commun., 1985) has obtained an age of 495±42 Ma for this group of rocks (located by symbol K on Fig. 2). He interprets the age as a metamorphic age and if so, this would suggest that the kyanite parageneses are part of an early Ordovician metamorphism.

It appears, therefore, that at least on portions of the western margin of the eastern metamorphic lobe, the last high-grade metamorphism represents an easterly extension of the Siluro-Devonian metamorphism(s) that produced the three western lobes. However, because the eastern metamorphic lobe has so many overlapping polymetamorphic effects and so little work has been done on it to date, there is little basis (other than simplicity as noted above) for arguing very strongly how much of the metamorphism in the lobe can be equated with that which occurred in the three western lobes.

The recent work of Berry (1986) is the most suggestive in terms of arguing that the Siluro-Devonian metamorphism(s) that produced the three western lobes might have affected rocks on the eastern side of the lobe. Berry's study covered the Camden Hills area (Fig. 2) and lies just west of the Turtle Head fault zone. It deals with the Megunticook and Penobscot Formations of inferred Cambrian to Lower Ordovician age.

Berry (1986) notes the existence of some compositional segregational banding that might be a remnant of a pre-Acadian fabric (e.g. a recrystallized spaced cleavage). Considering the age of the strata involved, the event causing this banding could be as old as Lower Ordovician. However, Berry suggests that the strongest metamorphic events occurred between Late Silurian and Middle Devonian time (i.e. Acadian). This suggestion is based on the field relationships between the metamorphic events (see below) and plutonic rocks that are dated, albeit only by analogy with dated plutons outside his area.

Of particular interest is the fact that Berry recognized two high-grade (inferred to be Acadian) events that involved development of micaceous pseudomorphs after staurolite and andalusite (some much like in western Maine) and disequilibrium textures. The first event involved the formation of andalusite and staurolite throughout the Camden Hills. Berry

believes this metamorphism may have accompanied formation of the foliation -- i.e. is syntectonic. However, due to the pseudomorphing produced by the subsequent high-grade event, he allows that the andalusite and staurolite may have crystallized after the foliation formed, but still before the final deformations.

The second high-grade event, a post-tectonic recrystallization, reached sillimanite grade and caused much of the pseudomorphing of andalusite and staurolite. The nature of these events and their suggested ages are strikingly similar to those inferred for M_2 and M_3 in the three western lobes, and so lends support to the proposition that the Siluro-Devonian metamorphism(s) in the three western lobes extended completely across the eastern metamorphic lobe.

A third metamorphic event recognized by Berry involved extensive retrogression of rocks metamorphosed by the above described events. No time constraints exist for this event except that it be Acadian or younger. The intensity of this retrogression varies systematically across his map area, being greatest in the east and south. This suggested to Berry that the event was a regional-scale phenomenon rather than just local retrogression due to minor faults, etc. It may be of some significance that at least in Berry's map area the intensity of the retrogression is more pronounced in the direction of the Turtle Head fault zone. Conceivably this event could be related to low-grade metamorphism (described below) which affected rocks on the east side of this fault.

In support of Berry's suggestion of this retrograde event being regional in extent, it should be noted that similar late retrograde effects are quite prominent in much of the eastern metamorphic lobe. North of the Camden Hills, Bickel (1974) described a retrograde zone in the east-central part of the Belfast quadrangle. To the south of Berry's study area, Osberg and Guidotti (1974) described a retrograde metamorphism line in part of the Rockland, Thomaston, and Rockport quadrangles. Recent work by the author shows that extensive retrogression continues farther southward into the Tenants Harbor quadrangle also.

Although the rocks to the east of the Turtle Head fault zone on the islands and eastern shores of Penobscot Bay (Fig. 1) are not in the eastern metamorphic lobe as defined above, their history may eventually bear on how one interprets this lobe and so it is discussed in this section. Virtually all of the information on the metamorphic rocks east of the Turtle Head fault zone is based on the work of Pinette (1983) and Stewart (1974, and pers. commun., 1986). Work on the rocks immediately west of the fault zone involves that of Berry (1986), Osberg and Guidotti (1974), Stewart (op. cit.) and Bickel (1971, 1974, 1976).

East of the Turtle Head fault zone, some medium to high-grade Precambrian rocks are present on Islesboro (Fig. 1) beneath a cover sequence. They are discussed under Precambrian metamorphism in the next section. According to Osberg et al. (1985), the cover sequence may be as old as Precambrian. In addition to the cover sequence, the rocks in Penobscot Bay (east of the Turtle Head fault zone) include volcanic strata ranging in age from Precambrian (North Haven Formation) to Siluro-Devonian (Castine Formation). Only low-grade metamorphism has affected the cover sequence and volcanic strata. The work of Stewart (op. cit., and in progress) suggests the possibility that more than one low-grade metamorphism may have affected some of these rocks.

Specifically, sub-biotite zone metamorphism probably affected the cover sequence and North Haven Formation during pre-Middle Silurian time. Stewart suggests that the metamorphism may be of Cambro-Ordovician age like that suggested for the Ellsworth Schist (see later section on Ordovician metamorphism).

Obviously the metamorphism of the Siluro-Devonian Castine Formation must be younger than that discussed above. Stewart believes that the cover sequence was re-metamorphosed during a Siluro-Devonian event that also affected the Castine Formation. The suggestion is based on the development in the cover sequence of recrystallized phengite plates along a cleavage that appears to have formed during motion on the Turtle Head fault zone. Geological and radiometric dating arguments enable Stewart to show that the fault was active during the time period from 425 to 380 Ma. Stewart believes that the recrystallization of phengite in the cleavage coincides with the same metamorphism that affected the Castine Formation.

This second, low-grade metamorphism could be the same as the low-grade, Siluro-Devonian regional metamorphism Ludman et al. (in press) have described along strike in the coastal lithotectonic block near the Canadian border. However, in the Penobscot Bay area it could be post-Siluro-Devonian and could even be on a different terrane from that of the rest of Maine.

An important feature of the metamorphism to the east of the Turtle Head fault zone is that it appears that rocks of very different grade have been juxtaposed against rocks to the west of the fault in the eastern metamorphic lobe (Fig. 1). This disparity in grade across the fault varies along the fault such that, although rocks on the west are usually much higher grade, in a few places (e.g. at the north end of Penobscot Bay) the rocks on the west are lower grade than those on the east of the fault. Verification of these seeming relationships must await further detailed work on both sides of the fault.

Finally, it was noted at the beginning of this section that some of the metamorphism in the eastern metamorphic lobe could be as young as Carboniferous or Permian. This suggestion is highly speculative and is based on preliminary results (Lux, Guidotti, and Newberg, in prog.) of $^{40}Ar/^{39}Ar$ dating on the hornblendes from the amphibolites that occur in the Cushing Formation just west of the Norumbega fault zone (essentially in close association with the kyanite-bearing rocks, see Fig. 2). From north to south (Freedom to Richmond, Maine) the closure ages progressively decrease from about 370 Ma to about 280 Ma. The ages at the north end of the traverse could represent cooling through 500°C of the Siluro-Devonian metamorphism. The much younger ages to the south could involve a Permian event, but at this time it can only be considered as an interesting

possibility. As amphibolites are common throughout the eastern metamorphic lobe, in contrast with the three western lobes, it should be feasible to apply the $^{40}Ar/^{39}Ar$ technique to the metamorphic rocks as well as the igneous rocks.

PRECAMBRIAN METAMORPHISM IN MAINE

Only two Precambrian metamorphic events have been recognized with a fair degree of certainty in Maine. They involve the rocks in the Chain Lakes massif and the structurally lowest rocks on Islesboro (Fig. 2). Two other possible Precambrian recrystallizations exist, both in the eastern metamorphic lobe, but the evidence for them is more suggestive than direct. These would include some of the metamorphism which affects units such as the Cushing Formation (i.e., the kyanite parageneses described above) and the upper amphibolite facies metamorphism that affects the Passagassawakeag Gneiss (Bickel, 1974). Because of the lack of direct evidence for Precambrian recrystallization of these units, further discussion is focused on the two cases for which there is direct evidence of Precambrian metamorphism.

Chain Lakes Massif

As seen in Figure 2, the Chain Lakes massif involves a northeast-trending elliptical patch of rocks, approximately 50 x 30 km, on the Maine-Quebec border, just northeast of its intersection with the New Hampshire border. It is a suite of high-grade rocks that most typically has a peculiar diamictite texture. Bulk compositions include amphibolites, pelites, quartzites, etc., but most common is a fragmental, quartzo-feldspathic granofels.

The geologic and petrologic aspects of the Chain Lakes massif have been studied by a number of authors. Recent reports include Boone et al. (1970), Boudette and Boone (1976), Boudette (1982), Biedermen (1984), Cheatham (1985), and Cheatham et al. (1985). Biederman reviewed virtually all previous work on the massif and attempted to assess the metamorphism in some detail. His study involved a detailed cross section along State Highway 27 across the southwestern end of the massif.

Based upon Biederman's observations and those of previous workers, the metamorphic picture he suggested is fairly simple (however, most other geologic aspects of the massif remain quite enigmatic). It includes:

(1) In Precambrian time, the massif itself was metamorphosed to at least amphibolite facies and probably in part to as high as upper amphibolite facies (possibly separable by the K-feldspar + sillimanite isograd). Biederman presented arguments that in addition to quartz + 2 feldspars + sillimanite + biotite ± muscovite, cordierite may have formed during the high-grade metamorphism but has now been replaced by chlorite during a later, low-grade metamorphism. Boudette and Boone (1976) report quartz + 2 feldspars + sillimanite + muscovite + biotite ± cordierite ± garnet present in some of their specimens. However, from their descriptions it is not clear if garnet and cordierite coexisted in the same rock such that conditions ranging into the granulite facies could be inferred. Nonetheless, it is reasonable to infer that the Chain Lakes massif was affected by at least upper amphibolite facies conditions and according to Biederman, recrystallization during this event was responsible for the main foliation present. This foliation was broadly arched by later deformational events.

(2) These high-grade rocks were all retrograded to greenschist facies, but with textural and mineralogical evidence of the high-grade event persisting in virtually all specimens. Quoting from Biederman (1984, p. 86):

> The low-grade assemblage consists of: quartz + muscovite + chlorite ± epidote + calcite + opaques, which places the low-grade event in the lower greenschist facies of regional metamorphism. Texturally, the effects of the retrograde event are apparent by the alteration of the high-grade phases. Biotite is commonly pseudomorphed by a combination of chlorite and sagenitic rutile; fibrous sillimanite is replaced by fibrous muscovite; cordierite(?) is converted to a fine-grained mat of pale green mica; and plagioclase is altered to aggregates of fine-grained white mica (sericite) or mica, carbonate and/or epidote (sausserite). In addition, plagioclase sometimes shows considerable exsolution and the possible development of peristeritic texture. However, no low-grade foliation is apparent and the low-grade minerals generally lie in the plane of the high-grade foliation.

The work of Boudette (1982) suggests that the Boil Mountain ophiolite which occurs along the southern margin of the Chain Lakes massif has also been adjusted to the same greenschist facies event. He reports the typical assemblage as chlorite + albite + epidote + actinolite in the metavolcanic rocks of the ophiolite.

(3) The Cambro-Ordovician to Siluro-Devonian strata surrounding the Chain Lakes massif have been recrystallized to this same low-grade of metamorphism, and like the Boil Mountain ophiolite were unaffected by the high-grade event that occurred in the massif. Biederman discusses the possibility that the greenschist facies conditions in the pre-Siluro-Devonian rocks could be due to both Taconian and Acadian activity although he favors the latter alone. Discussion below will return to this question.

(4) Geologic evidence alone indicates that the high-grade metamorphism in the Chain Lakes massif is pre-Taconian. In particular, the massif is intruded by the Attean quartz monzonite (443±4 Ma, U-Th-Pb method, Lyons et al., 1986) but the pluton is not affected by the high-grade metamorphism.

The discussion presented by Biederman (1984) strongly suggested that the high-grade metamorphism of the Chain Lakes massif occurred during Precambrian time, although some ambiguity remained. Ambiguity also existed for the age of the event(s) that retrograded the Chain Lakes massif and prograded the surrounding Lower Paleozoic strata. Fortunately, the more recent work by Cheatham (1985) and Cheatham et al. (1985) has to some extent reduced the degree of these ambiguities. Their

Rb-Sr whole rock and Rb-Sr mineral analyses suggest that the high-grade metamorphism may have occurred at 770 Ma and the later low-grade event may have occurred at 405 Ma. This latter date would support Biederman's suggestion that the low-grade event belongs solely to the Acadian orogeny.

Cheatham (1985) and Cheatham et al. (1985) also presented strontium evolution lines and Nd model ages for the Chain Lakes massif. Their results suggest that the protolith of the massif was probably no older than ca. 1525 Ma, thereby putting a maximum age on its formation. In addition, this 1525 Ma age seems to agree well with the 1500 Ma zircon age reported by Naylor et al. (1973) for the Chain Lakes massif.

(5) Finally, drawing from previous works and Stewart (pers. commun., 1984), Biederman argues that the Chain Lakes massif is an allochthonous mass emplaced before or during the Taconic orogeny. Most important for our purposes is that the massif was probably emplaced as an already high-grade mass. Hence, the high-grade metamorphism occurred somewhere other than where the massif is now located.

Islesboro

Strongly recrystallized Precambrian rocks occur on Islesboro, the large island in the north-central portion of Penobscot Bay (Fig. 1). They have been discussed briefly by Stewart (1974) and by Brookins (1976). However, the account presented here is mainly based upon recent information provided by Stewart (pers. commun., 1986).

The rocks of concern occur as a small, oval-shaped patch on the northern portion of Islesboro and on Seven Hundred Acre Island a few miles to the south as an elongate, narrow zone near the center of the island and then continuing a short distance southward to Lime and Lasell Islands (Stewart, pers. commun., 1987). These crystalline rocks form the structural basement of the islands.

Pelitic bulk compositions involve the assemblage biotite + andalusite + garnet, and calcareous rocks have assemblages with minerals including hornblende, phlogopite, and vesuvianite(?). Hence, the metamorphic grade was approximately at middle amphibolite facies. The age dating of Brookins (1976) shows that this metamorphism is older than ca. 600 Ma, the age of a granitic pegmatite that cuts the metamorphic rocks. In addition, Brookins (1976) gives a "rough" Rb/Sr date for the metamorphic rocks at ca. 750 Ma ± 100 Ma.

These middle amphibolite facies assemblages have been extensively retrograded to greenschist facies (sub-biotite grade). It would appear that this retrogression reflects the effects of the low-grade metamorphism(s) that have affected the rocks in the cover sequence.

Summary

In Maine, only two areas of Precambrian rocks can be identified with much certainty. Metamorphism in the Chain Lakes massif may have occurred at some location quite distant from the present location of the rocks. Based on various suggestions in the literature (Zen, 1983; Zen et al., 1986), the same question can be raised with regard to the Precambrian metamorphic rocks on Islesboro. Thus, in both cases, the Precambrian metamorphism may bear on the question of suspect terranes in Maine.

CAMBRIAN METAMORPHISM

Cambrian metamorphism in Maine is not well established. The tentative aspect of the following discussion must therefore be kept in mind. To a great extent the question of Cambrian metamorphism in Maine arises from evidence that a deformational event occurred in Cambrian to Early Ordovician time such that folded Cambrian strata are unconformably overlain by Lower to Middle Ordovician strata. Evidence of such a deformational event has been noted in a number of localities in northern Maine and Quebec (e.g. Cooke, 1955; Riordan, 1957; Pavlides et al., 1964; Neuman, 1967; and Hall, 1970). Neuman (1967) named this event the Penobscot disturbance and more recently (Neuman and Max, 1986) called it the Penobscottian orogeny. The tectonic map of Maine (Osberg et al., 1985) shows the relevant unconformity throughout much of northern Maine.

Because regional metamorphism commonly accompanies orogenic activity, it is possible that some metamorphism accompanied the Penobscottian orogeny. However, Penobscottian metamorphism is difficult to document because the entire terrane has undergone later low-grade metamorphism (Fig. 1) during the Acadian orogeny.

Some evidence, however, suggests low-grade metamorphism during the Penobscottian. For example, Neuman (1967) and Hall (1970) indicate that the Cambrian strata were both folded and cleaved before being overlain by younger Ordovician beds. Moreover, the degree of folding and deformation is more intense in the rocks below the unconformity. The formation of a recognizable cleavage seems especially suggestive that low-grade metamorphism accompanied the Penobscottian orogeny.

An additional feature of possible Penobscottian metamorphism involves aspects differing from "conventional" views of broad-scale regional metamorphism. Specifically, field work in the past decade has demonstrated the existence of a number of mélange and ophiolitic units that form a northeast-trending belt, essentially coinciding with the Lobster Mountain, Munsungun-Winterville, and Boundary Mountains anticlinoria (Fig. 2) (see also Osberg et al., 1985). Recent work by Boone (1983, pers. commun., 1986, and in press) and by Pollock (Whittier and Pollock, 1986; Pollock, 1985, and pers. commun., 1987) has suggested the existence of the type of metamorphism associated with some aspects of subduction zones and possibly even seafloor metamorphism.

Boone (1983, and pers. commun., 1986) has described the Hurricane Mountain Formation as a mélange unit within the Lobster Mountain anticlinorium. The matrix of the mélange has a Middle Cambrian to Tremadocian(?) age. This formation is one of three polydeformed units that lies unconformably below Lower, Middle, and Upper Ordovician strata. The structural and stratigraphic similarities with the relations described above by Neuman (1967) and Hall (1970) are obvious.

An important additional point for this report is that Boone (1983, and pers. commun., 1986) has observed a variety of clasts in the Hurricane Mountain Formation including amphibolites and metagabbros with assemblages indicating markedly higher temperatures than those that produced the low-grade metamorphism of the matrix (see Boone, in press, for an extended treatment). Clearly the age of the clasts must predate the age of the matrix of the Hurricane Mountain Formation. If the clasts are only a little older than the matrix, they could be a reflection of some higher temperature metamorphism during the Penobscottian orogeny. It would be desirable to attempt $^{40}Ar/^{39}Ar$ dating on these amphibolite clasts.

The work of Pollock (op. cit.) involves mélange units in the vicinity of Caucomgomoc Lake, at the northern end of the Boundary Mountains anticlinorium. Pollock recognizes three "packages" of basaltic rocks. The youngest is Silurian (Wenlock) in age and has been metamorphosed weakly (pumpellyite + chlorite; i.e. sub-greenschist) but still retains a clear volcanic aspect. The older "packages" are of Cambrian(?) and/or Ordovician(?) age, have a distinctly different field appearance than the Silurian basalts, and are associated with mélange features. One, the Caucomgomoc Lake Formation, is non-schistose and metamorphosed to greenschist facies. Pollock tentatively interprets its metamorphism as an ocean-floor metamorphism (i.e. spilitization) which may or may not predate the Penobscottian orogeny. The Avery Brook Formation, the other older basalt package, is a thoroughly recrystallized, schistose metabasalt. Again, on a tentative basis, Pollock speculates that this package was actually subducted. Pollock, like Boone and Boudette (1985), believes the subduction was associated with the Penobscottian event.

In summary, for the area excluding the eastern metamorphic lobe, some evidence suggests the possibility of regional metamorphism accompanying the Penobscottian orogeny, but none of it is very definite. The question of whether any significant regional metamorphism accompanied the Penobscottian orogeny is of considerable importance for large-scale, plate tectonic reconstructions of the Appalachian-Caledonide belt as evident from the recent discussion by Neuman and Max (1986). Their work considers the interrelationships among the Penobscottian, Grampian, and Finnmarkian orogenies in the context of the larger Appalachian-Caledonide belt. The Grampian and Finnmarkian orogenies were accompanied by intense metamorphism. It would be highly desirable to have a clearer picture of any metamorphism associated with the Penobscottian orogeny so that additional comparisons or contrasts can be established with the other orogenic events. Finally, it is interesting to speculate as to whether the kyanite-grade metamorphism in the eastern metamorphic lobe might be a high-grade Penobscottian effect.

On a more local scale, the implications of the work by Boone (op. cit.) and Pollock (op. cit.) are extremely interesting and hold promise of providing information on aspects of apparently Penobscottian metamorphism which will be of considerable value in attempts to locate (spatially and temporally) subduction zones active during Early Paleozoic time.

ORDOVICIAN METAMORPHISM

Excluding the coastal lithotectonic belt (treated separately in the next section), discussion of Ordovician metamorphism in Maine largely centers around the extent to which the Taconian orogeny occurred in Maine. Some early studies imply the occurrence of a fairly pronounced Taconian deformation throughout much of northern Maine (see Pavlides et al. 1968; Roy and Mencher, 1976; and Roy, 1980). Indeed, megatectonic syntheses as recent as Zen et al. (1986) ascribe suturing and closing of the Iapetus Ocean to Taconian events. However, in the past few years there has been a downplaying of possible Taconian deformation in Maine, possibly due to greater emphasis on the Penobscottian event, e.g. Neuman and Max (1986). Elsewhere, as in Vermont, the Taconian orogeny was a major deformational event accompanied by intense metamorphism. Moreover, it was noted by Guidotti et al. (1983) that the low-grade metamorphism throughout the area lying to the west of the Connecticut Valley-Gaspé synclinorium in the Eastern Townships of Quebec is attributed to the Taconian orogeny. This would include strata only ca. 75 km northwest of western Maine.

Several general points do suggest some sort of orogenic activity in Maine during the latter part of the Ordovician Period. These include: (1) Osberg et al. (1985) show two lower Paleozoic (pre-Silurian) unconformities throughout large portions of northern Maine. (2) Intrusion of Highlandcroft plutons (Attean, 445±4 Ma, Adamstown, 453 Ma) occurred in Late Ordovician time, Lyons et al. (1986). Moreover, the cataclastically deformed Rockabema quartz diorite in north-central Maine is older than Early Silurian (late Llandovery) but intrudes Middle Ordovician rocks (Neuman, 1967). Other more specific points include: (1) Boone (1983, Figure 1) shows two pre-Silurian, Paleozoic unconformities in the Brassua Lake and Moosehead Lake quadrangles. (2) Harwood (1970, 1973) described folding of the pre-Silurian strata of western Maine before the Acadian orogeny. (3) Ludman (pers. commun., 1986) has observed foliations in Caradocian strata of the Miramichi anticlinorium (Fig. 2) near Danforth that formed during a pre-Silurian folding event.

Possibly the view expressed by Pavlides (pers. commun., 1987) is presently the best way of viewing the Taconian deformation as seen in Maine. That is, it was modest in intensity and

occurred in northern Maine to a variable extent through both space and time. In this respect it would contrast with the Acadian orogeny which appears to have involved intense deformation, essentially during a single time span throughout most of Maine.

In the context of the above discussion, the author will assume that some sort of deformation occurred at least in northern Maine during the latter part of Ordovician time such that it might be ascribed to a Taconian event. Even with this assumption it is difficult to identify associated metamorphic effects because all of the metamorphism (excluding some of the demonstrable Acadian or younger, high-grade metamorphism) which affects Ordovician strata is of low intensity and the rocks have also been affected by later, low-intensity Acadian metamorphism. This problem is similar to that mentioned in discussing possible Penobscottian metamorphism.

Aside from the expectation of some metamorphic effects to be associated with an orogenic event, the question becomes: do we have any direct observations suggesting Taconian metamorphism? As developed below, the answer appears to be yes, but with much uncertainty. Further, with the sole exception of the eastern metamorphic lobe, there is no evidence of any high-grade, regional Taconian metamorphism in Maine.

Some of the relevant, direct observations include the following:

(1) In discussing the prehnite + pumpellyite facies rocks of northern Maine which occur in both Ordovician and Silurian strata (in Aroostook County, essentially the same area as covered by Roy and Mencher, 1976), Richter and Roy (1974) note the occurrence of prehnite + pumpellyite-bearing lithic clasts in Silurian graywackes. They suggest that this could indicate possible Taconian metamorphism to this grade. Later on, these same authors (1976) appear to be more equivocal on this point, but still leave open the possibility of Taconian age, prehnite-pumpellyite facies metamorphism.

(2) Ludman (pers. commun., 1986) observed weakly developed pre-Silurian foliations in Caradocian strata. One might presume that development of a foliation (weakly aligned muscovite) requires at least low-grade metamorphism.

(3) A contact metamorphic aureole was produced by intrusion of the Adamstown granite (Guidotti, unpub. data). This aureole has subsequently been retrograded by the Acadian metamorphism associated with emplacement of the Mooselookmeguntic pluton. In addition, tabular chloritic clots occur sporadically in the Cambro-Ordovician strata for several kilometers north of the Adamstown granite (Fig. 1). These clots could be interpreted as pseudomorphs of porphyroblasts (e.g. biotite or chloritoid) formed by a pre-Silurian metamorphism.

In summary, for that portion of Maine lying to the west of the Norumbega fault zone, there appears to be abundant evidence of significant pre-Silurian deformation and plutonic activity. However, it is not at all straightforward whether to ascribe them to a Penobscottian or Taconian orogeny. Nonetheless, it seems that widespread, low-grade metamorphism is at least a possibility during Early Paleozoic (i.e. pre-Silurian) orogenic events. There is also the suggestion of some localized higher-grade metamorphism during the Penobscottian (e.g. the observations of Boone, op. cit.).

Ordovician Metamorphism in the Coastal Lithotectonic Belt

Here, the aim is to discuss another "Ordovician metamorphism candidate" which occurs in the coastal lithotectonic block, the Ellsworth Formation. The area of concern includes the region east-northeast of Penobscot Bay which is underlain by this unit (see Osberg et al. 1985; and area on Fig. 1 designated as Ordovician greenschist facies metamorphism). Part of this area was included in the report of Stewart and Wones (1974), and based upon their results, the whole area of Ellsworth Formation was shown by Guidotti (1985) as a possible Ordovician metamorphism. The most up-to-date information on the metamorphism of this unit is the work by D. B. Stewart (pers. commun., 1986).

The regional metamorphism of the Ellsworth Formation is everywhere at greenschist grade as shown by the assemblage albite + chlorite + phengitic-muscovite + quartz + Mn-rich ilmenite in metapelites. More mafic bulk compositions include actinolite + epidote + calcite.

The Ellsworth Formation is shown with a stratigraphic age of Precambrian to Ordovician (Osberg et al. 1985). This information alone does little to establish the age of the low-grade regional metamorphism(s) that may have affected the unit. Moreover, in many places it is either in fault contact with the Ordovician age Penobscot Formation or intruded by Devonian plutons.

Some information on the age of the metamorphism(s) of the Ellsworth Formation is afforded by its mineralogic and stratigraphic/structural relations with the volcanic rocks of the Siluro-Devonian Castine Formation. In places far removed from any contact metamorphic effects, these volcanic rocks lie unconformably over the Ellsworth Formation and are themselves metamorphosed to greenschist grade, essentially like the Ellsworth Formation. However, according to Stewart, the overlying Castine Formation contains much more evidence of relict minerals (e.g. augite) and local equilibrium is less well established than in the Ellsworth Formation (e.g. a single chlorite composition in a given Ellsworth specimen vs. a range of chlorite compositions in individual Castine specimens). On the basis of such observations one could infer that the Ellsworth Formation has been affected by an event preceding the later, low-grade metamorphism (see previous section on eastern metamorphic lobe) which affected the Castine as well as the older stratigraphic units.

However, there exists other, still more compelling evidence that the Ellsworth Formation was affected by a pre-Silurian, greenschist-grade regional metamorphism. Specifically, along the unconformity at the base of the Castine Formation at both Bagaduce (Stewart, 1974, p. 228, 236) and Ames Knob (Stewart, 1974, p. 236), there are strongly foliated greenschist-grade frag-

ments of the Ellsworth Formation. Also present along the unconformity at Ames Knob are Llandovery C4-5 fossils (Berry and Boucot, 1970, p. 116). These observations strongly support the suggestion that the Ellsworth Formation was regionally metamorphosed to low grade before Silurian time, but because the age of the formation is not well established, the age of the metamorphism is ambiguous except that it is pre-Silurian. On the other hand, Brookins (1976), has determined a Rb/Sr whole-rock isochron age of 510±15 Ma for the Ellsworth Formation based on 12 samples. According to Brookins, this age (late Cambrian, using the 1983 DNAG Geologic Time Scale) would be an approximate stratigraphic age. If one accepts this stratigraphic age for the Ellsworth Formation, the above described metamorphism would have been Ordovician. Until direct, contradictory evidence is avaliable, the writer continues to accept Ordovician as the best "tentative guess" for the age of the first low-grade metamorphism to effect the Ellsworth Formation (i.e. essentially the same age as the first event to affect the cover sequence on Islesboro).

POST-DEVONIAN METAMORPHISM

The studies of Faul et al. (1963), Zartman et al. (1970), and Dallmeyer and Van Breeman (1981) provided direct observational data of post-Devonian isotopic ages (metamorphic(?); Permian) in various parts of central and southern Maine. However, these studies largely focused on isotopic approaches and were not tied closely with petrologic and geologic aspects. The results were variously interpreted to indicate a Permian event to diachronous cooling following Devonian plutonism and metamorphism (see also Lux et al., in press).

Based on geologic and petrologic factors, Osberg (1968) and Guidotti (1970) had also made tentative suggestions of some post-Devonian regional metamorphism in Maine. However, as with the above cited isotopic studies, there was little in the way of truly convincing arguments. Hence, it was routinely assumed that all of the high-grade metamorphism in western and southern Maine was of Devonian age. The first really strong evidence for a post-Devonian regional metamorphism in Maine resulted from the isotopic age dating of the Sebago batholith (325 Ma) (Hayward and Gaudette, 1984; Aleinikoff, 1984). Noting that the highest grade metamorphism in western Maine is spatially related to the Sebago batholith, Aleinikoff et al. (1985) briefly speculated on the possibility of this metamorphism being the same age as the Sebago pluton.

Simultaneously, Lux and Guidotti (1985) presented a much more complete picture combining geologic and petrologic data as well as $^{40}Ar/^{39}Ar$ hornblende cooling ages for various rock types north of the Sebago batholith. Their results make a compelling argument that the extensive area of migmatitic, upper amphibolite facies rocks to the north of the batholith is due to heat imparted to the country rocks by the intruding Sebago batholith. More recently, Guidotti et al. (1986) have presented further details of the effects of this Carboniferous metamorphism on both the metasedimentary strata and Devonian plutons that occur to the north of the Sebago pluton.

Briefly, the metamorphic event was regionally extensive as shown merely by the distribution of upper amphibolite facies, migmatitic rocks. These rocks (corresponding to K-feldspar + sillimanite zone) were shown previously as due to Devonian metamorphism (Guidotti, 1985), but would now be reassigned to a younger age. Thus, the highest grades of the three western metamorphic lobes as shown on Figure 1 is Carboniferous in age rather than Devonian. This event involved a post-tectonic heating superimposed on rocks metamorphosed earlier during Devonian time. To the north this recrystallization resulted in readjustment of the grades attained during the M_3 event described earlier, and in the north-central part of the Rumford 15' quadrangle produced a retrogression of the grades previously attained.

Subsequently, J. J. De Yoreo (in Guidotti et al., 1986; and De Yoreo et al., this volume) has carried out thermal modelling of the heating effects expected from the Sebago pluton if it was intruded as a relatively thin sheet (Hodge et al. 1982) at about 15 km and with a gentle northerly dip. In the context of the indicated geometry, De Yoreo's modeling shows that the Sebago batholith would be an adequate heat source to produce the observed metamorphism. Clearly the nature of this Carboniferous metamorphism is quite similar to the Devonian metamorphism described earlier in that it involves post-tectonic recrystallization related to heat imparted by a gently dipping, sheet-like granitic pluton. As with the Devonian plutons, the intrusion took place in rocks which had essentially cooled to ambient temperature.

In Guidotti et al. (1986) the heating associated with the Sebago intrusion was considered to be a late effect of the Acadian orogeny. This was done because the thermal modeling of De Yoreo shows that the Sebago intrusive event is likely to be a later response to the deep-level melting that also produced the Devonian plutons subsequent to the crustal thickening produced during Acadian deformation. In this sense the metamorphic zonation shown by Guidotti (1985) for the three western lobes is all Acadian metamorphism, but includes recrystallization events separated by as much as 55 million years. In the context of the Sebago batholith being a late stage Acadian pluton, it has intruded earlier Acadian plutons such as the Songo granodiorite and thus produced a reheating effect. The possible consequences of such reheating were considered by Lux and Guidotti (1985) and Guidotti et al. (1986) (see also, Gibson and Lux, in press).

A further interesting aspect of the Carboniferous metamorphism in western Maine is that on the south side of the Sebago batholith, presumably its underside, kyanite-bearing assemblages have formed. As reported by Thomson (1986) and Thomson and Guidotti (1986, this volume), true Barrovian pressures appear to have prevailed during the recrystallization. The metamorphic grades, ranging to amphibolite facies, show a clear spatial relationship with the contact of the Sebago batholith. Moreover, the metamorphism involves mainly post-tectonic

recrystallization superimposed on earlier Devonian metamorphism.

In contrast with the metamorphism on the north (upper) side of the pluton (highly migmatitic, upper amphibolite facies), the grade on the south (lower) side only reached amphibolite facies and the rocks are not migmatitic. Presumably this difference reflects greater heat transfer upward due to escaping hot fluids moving upward, as the Sebago body cooled. Additionally, although some pegmatites are present close to the Sebago pluton along its southern margin, it is on the northern side that there is a spectacular development of pegmatites. Variable sized bodies of pegmatite occur in a broad zone extending out into the country rocks for about 30 km. This zone includes many famous pegmatite quarries that have yielded vast quantities of tourmaline, beryl, lepidolite, and other minerals. From the spatial and petrologic considerations discussed above, it is likely that the Sebago pluton might be the ultimate source for these pegmatites.

The areal extent of the Carboniferous recrystallization has not yet been determined except for the areas cited in the above discussion. Presumably the distribution of upper amphibolite facies-grade rocks, shown on Figure 1 in the three western metamorphic lobes, gives a rough indication of its potential areal extent in the north side of the pluton. Presently there is no clear geologic or petrographic evidence bearing on the age of metamorphism to the south and east of the Carboniferous metamorphism described by Thomson (1986) on the south side of the batholith. Presumably it is Siluro-Devonian in age and mainly part of the three western metamorphic lobes. However, as seen from Figure 1, at least some of it is involved in the problem of the ages and numbers of metamorphisms that have affected rocks in the coastal lithotectonic block.

Earlier in the discussion of the eastern metamorphic lobe, mention was made of some preliminary Permian hornblende $^{40}Ar/^{39}Ar$ closure ages in rocks near the boundary of the coastal lithotectonic block. Moreover, the Permian ages suggested by the work of Zartman et al. (1970) and Faul et al. (1963) (cited above) applies most directly to the portions of Maine to the south and east of the Sebago pluton. At present it is still an open question as to what these Permian ages represent.

From the preceding discussions it should be apparent that the areal extent of Carboniferous metamorphism in western and southern Maine and the existence of any even younger metamorphic events is still not fully established.

Finally, in this discussion of post-Devonian metamorphism, some discussion of the interesting study by Gottfried et al. (1984) is necessary. In the Mars Hill, Bridgewater, and Presque Isle quadrangles (Aroostook County) they have observed a number of tholeiitic and mafic-alkalic dikes which are weakly metamorphosed. They discuss color alteration index data for conodonts from adjacent metasediments which indicate temperatures compatible with the metamorphic minerals formed in the dikes. The minerals imply zeolite and low greenschist facies.

$^{40}Ar/^{39}Ar$ work on these dikes is compatible with their intrusion during Late Silurian to Middle Devonian times and subsequent weak metamorphism in the Pennsylvanian (ca. 300 Ma) which disturbed the isotopic systems. It is interesting that the dikes do not intrude the Middle Devonian Mapleton Sandstone which rests unconformably on the pre-Acadian strata which the dikes do intrude. The Mapleton Sandstone appears to be essentially unmetamorphosed but was deformed in post-Acadian time by the Maritime disturbance of Pennsylvanian age. Gottfried et al. (1984) suggest that this may relate to the time at which the dikes were metamorphosed.

The work of Gottfried et al. (1984) raises obvious questions about the assertion made earlier that the low-grade metamorphism throughout northern Maine is mainly of Devonian age. Only further isotopic studies like those in Gottfried et al. (1984) will resolve such questions. Until such work is done, caution seems in order when discussing the age of the low-grade metamorphism in northernmost Maine.

SUMMARY AND CONCLUSIONS

An overview has been provided of the metamorphic effects that can be discerned in the rocks now exposed in Maine. Clearly there is a wide range of metamorphic ages (Precambrian to Carboniferous and maybe even Permian) and also a range in metamorphic styles and conditions. The overview is highly uneven in terms of the extent of coverage given to the various metamorphic events. This unevenness results from the variable amounts of data available on the metamorphism in different parts of the state. The wide range in the extent to which the metamorphisms in Maine are understood or can be characterized is a crucial point for readers to perceive. It constrains the extent to which the metamorphic information can be integrated with large-scale geological models or syntheses of the northern Appalachians. Below, an attempt is made to appraise the status of work to date in order to give readers some idea of the extent to which the metamorphic information can (in the author's view) be usefully integrated into broad-scale, geologic syntheses.

Our understanding of the Precambrian metamorphism of the Chain Lakes massif, the Siluro-Devonian regional metamorphism in the three western lobes, and the contact metamorphism around the Siluro-Devonian plutons which intrude the extensive area of low-grade rocks in central, eastern, and northern Maine, probably documents adequately their geologic significance. Possibly the essential significance of the high-grade, regional Carboniferous metamorphism is also well enough known that it can be reliably integrated into broader geologic models. Fortunately, the Siluro-Devonian and Carboniferous metamorphism, (or in aggregate, Acadian metamorphism), covers a very large part of the high-grade metamorphism in Maine.

Further study of the metamorphism noted above will probably not provide much increase in terms of their geological significance. On the other hand, because much of the Siluro-Devonian and Carboniferous metamorphism occurred under nearly isobaric conditions and closely approached equilibrium, it provides an ideal setting to study the detailed nature of

metamorphic reactions and other mineralogically oriented problems.

In contrast, the other metamorphisms discussed in this overview are, to date, so poorly characterized that any "geologic synthesizers" would be well advised to use extreme caution in integrating the metamorphic information into their thinking. This caution seems especially appropriate for discussions that cover the area involving the eastern metamorphic lobe and coastal lithotectonic block. Unfortunately, these areas coincide with areas presently of extreme interest for geologic models aimed at trying to work out a plate tectonic synthesis for the easternmost portions of the northern Appalachians (e.g. Zen et al., 1986). For example, the areas noted are intimately involved with the question of what is, and where is Avalonia distributed along the east coast of North America (e.g. Stewart, 1986 and other papers in that symposium).

The metamorphic studies to date suggest a very complex history in the eastern metamorphic lobe as well as most of the rest of the coastal lithotectonic block. The results of these studies are such that there is much reason to believe that refinement of the metamorphic story there will pay significant dividends toward development of much more refined overall geologic syntheses of the coastal portions of the northern Appalachians -- especially in terms of questions involving plate-tectonic models. It is evident from the tentative statements made in the overview of Cambrian and Ordovician metamorphism that much remains to be done on them. This would apply especially with regard to metamorphism associated with the Penobscottian orogeny. As noted earlier, this might enable more refined interpretation and/or comparison with the Grampian and Finnmarkian orogenies in Europe. Moreover, the fact that mélange-type units appear to be involved in the Penobscottian suggests it would be well to be aware of metamorphic effects that might be associated with sea-floor and/or subduction zone types of metamorphism. Fortunately, work is well underway on various aspects of the metamorphism associated with the Penobscottian orogeny (e.g. Boone, op. cit., and Pollock, op. cit.).

Finally, in order to better relate the geologic events in Maine with those elsewhere in the northern Appalachians, it would be desirable to better establish the areal extent of the Carboniferous metamorphism and also to establish with some certainty whether or not any Permian metamorphic overprint has affected portions of Maine.

ACKNOWLEDGMENTS

The author is indebted to a host of colleagues who read and criticized all or portions of this work. These include: H. N. Berry, G. M. Boone, J. J. De Yoreo, H. E. Gaudette, B. A. Hall, A. Ludman, D. R. Lux, R. B. Neuman, L. Pavlides, S. G. Pollock, D. B. Stewart, and W. E. Trzcienski. Moreover, many of these same people generously made available to the writer unpublished manuscripts and data. Support has been provided via EPSCOR Grant ISP-8011448.

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Maine Geological Survey
Studies in Maine Geology: Volume 3
1989

A Thermal Model for Carboniferous Metamorphism near the Sebago Batholith in Western Maine

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ABSTRACT

In southwestern Maine, a Carboniferous-aged metamorphic event (M_C) produced widespread medium to upper amphibolite facies metamorphism. Although this region of high-grade metamorphism has a strong temporal and spatial correlation with the 330±5 Ma Sebago batholith, the effects of M_C extend some 30-50 km north of the batholith and the reason for that correlation is unclear.

In this paper we present geochronologic data which provide an estimate of the ambient temperature at the time of the intrusion of the Sebago batholith, outline the progression of cooling throughout the batholith from 245 Ma to 230 Ma, and delineate the northern limit of M_C. We also present a thermal model which suggests that M_C is a deep-level contact aureole of the batholith itself and that the pluton is likely to be a result of lower crustal melting in response to Acadian crustal thickening. We hypothesize that the Sebago batholith, or closely related granitic sheets, extend at least 30 to 50 km to the northeast and that the metamorphism to the north of the Sebago batholith was produced above the batholith while that to the south was produced below the batholith.

INTRODUCTION

In their classic paper on Paleozoic regional metamorphism in New England, Thompson and Norton (1968) said of the plutons: "Their distribution relative to the higher grade metamorphic zones shows a striking correlation ... A genetic relationship is thus highly likely, but it is by no means clear which is the primary and which is the secondary feature." They also noted that while contact aureoles are well-defined in northern Vermont and New Hampshire and in northeastern Maine, throughout most of the high-grade terrane distinct contact aureoles are lacking, and high-grade metamorphism can occur tens of kilometers away from the nearest igneous outcrop. This suggested to them that the granitoid plutons were as likely to have been a result of, as a heat source for, the regional metamorphism.

Since that time, more detailed petrologic studies have revealed a polymetamorphic history for this regional metamorphism, particularly in western Maine near the northern terminus of the New England metamorphic high. In this region at least three distinct Devonian metamorphisms (M_1, M_2, and M_3) have been described (Holdaway et al., 1982; Guidotti et al., 1983) which appear to correlate with periods of igneous activity (De Yoreo et al., 1989). Most recently, the existence of a Carboniferous-aged metamorphism (M_C), temporally and spatially connected with the Sebago batholith in western Maine, was documented by Lux and Guidotti (1985) and Aleinikoff et al. (1985). In addition, geochronologic studies have shown that at least some of the plutons in western Maine which are spatially correlative with regional metamorphic isograds were not melted in situ but emplaced from below (Lux et al., 1986). Combining these petrologic and geochronologic observations with geological (Moench and Zartman, 1976) and geophysical (Hodge et al., 1982; Nielson et al., 1976; and Carnese, 1983) evidence for the

geometry of the plutons in the high-grade terrane, Lux et al. (1986) concluded that, at least in western Maine, the high-grade metamorphism results from contact effects near sill-like igneous intrusions at intermediate crustal levels. They also concluded that the plutons are likely to be a result of crustal thickening produced by the Acadian orogeny. The purpose of this paper is to present the results of a thermal model for Carboniferous-aged metamorphism in western Maine which addresses the question raised by Thompson and Norton (1968) by demonstrating that under conditions of normal, mantle heat-flow, the Sebago batholith itself was a viable heat source for the surrounding regional metamorphism provided its lateral dimensions are sufficiently large.

GEOLOGIC AND PETROLOGIC CONSTRAINTS

The Sebago batholith lies in the Kearsarge-central Maine synclinorium in southwestern Maine with its long axis perpendicular to the structural grain of the New England Appalachians (Fig. 1 and 2). The geology and petrology to the north of the batholith have been studied in detail (Guidotti, 1965; Moench and Zartman, 1976; Creasy, 1979; Holdaway et al., 1982; Guidotti et al., 1983). At least 13 km of Ordovician, Silurian, and Devonian metasedimentary rocks are exposed in the region (Moench and Zartman, 1976). These strata were deposited on a shelf/slope and were subsequently deformed into upright, northeast-trending, isoclinal folds during the Acadian event at about 400 Ma. The rocks were then metamorphosed during a series of static events (M_1?, M_2, M_3, and M_C).

The metamorphic grade of the rocks associated with M_C ranges from medium amphibolite facies to upper amphibolite facies. The K-feldspar + sillimanite isograd approximately parallels the northern boundary of the Sebago batholith at a distance of 10 to 25 km (Fig. 1). A transect from Puzzle Mountain southward to the northern contact of the Sebago batholith passes through lower sillimanite, upper sillimanite, and K-feldspar + sillimanite zones (Fig. 2). Textural evidence demonstrates that the reactions involved in M_C were prograde at pressures less than the alumino-silicate triple point pressure of Holdaway (1971). Muscovite-almandine-biotite-sillimanite geobarometry gives metamorphic pressures of 3.8 ± 0.25 kbar (Holdaway et al., 1988). This is consistent with a petrogenetic grid approach based on the melting relations for pelitic rocks of Thompson and Algor (1977). A petrogenetic grid for these rocks is shown in Figure 3 with $P_{H2O} = P_{Total}$, and the inferred temperatures of metamorphism for the various isograds are listed in Table 1. The heavy line in Figure 3 indicates the metamorphic field gradient (array of maximum temperatures) for the rocks of M_C. Metamorphic grade increases towards the batholith in a nearly isobaric fashion and temperatures of metamorphism range from about 580°C in the lower sillimanite zone at the top of Puzzle Mountain to about 635°C in the K-feldspar + sillimanite zone near the Sebago batholith. Some of the rocks in the upper sillimanite zone are migmatitic. In the K-feldspar +

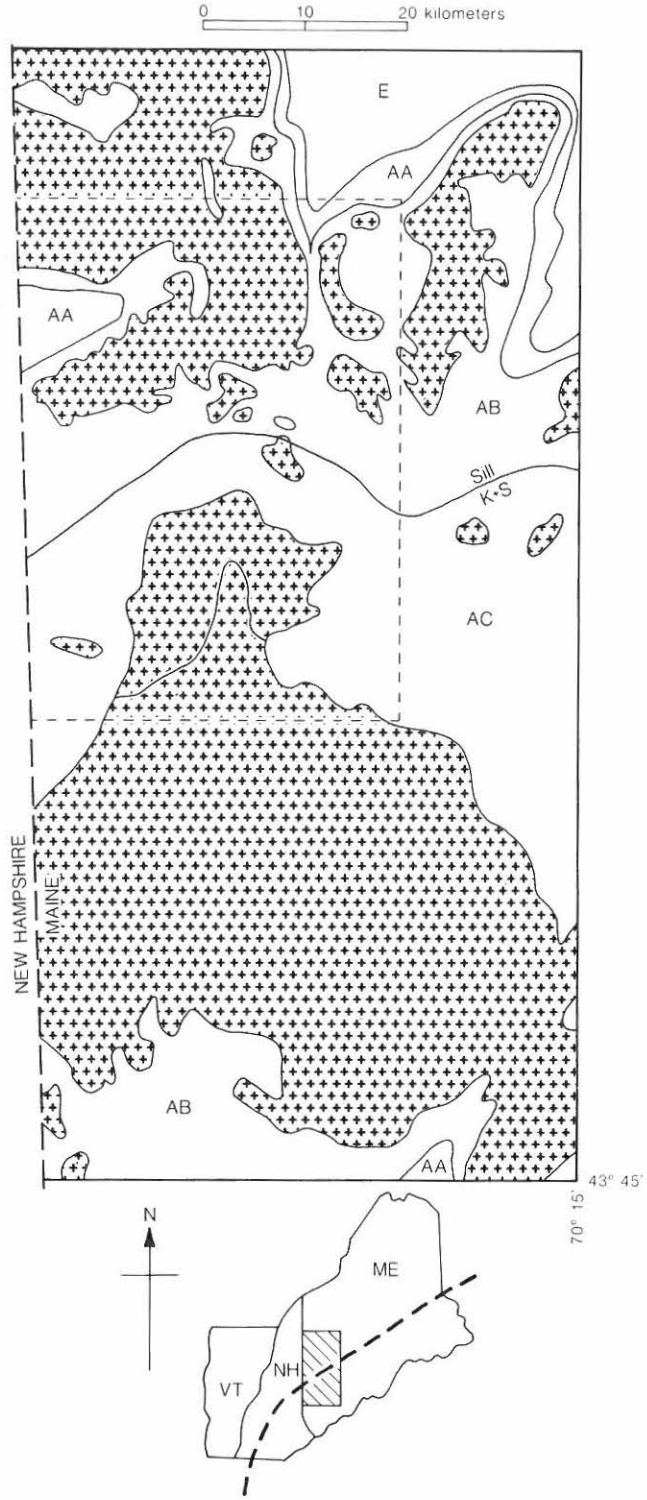

Figure 1. Metamorphic map after Guidotti et al., (1983) of southwestern Maine. Metamorphic facies designation. E -- epidote amphibolite characterized by Fe-garnet in pelites; AA -- lower amphibolite characterized by staurolite and kyanite or andalusite in pelites; AB -- medium amphibolite characterized by sillimanite + K-feldspar (K+S) in pelites. Igneous rocks are designated by pluses. The boxed-in area is shown in more detail in Figure 2. The axis of the Kearsarge-central Maine synclinorium is depicted by the dashed line on the index map.

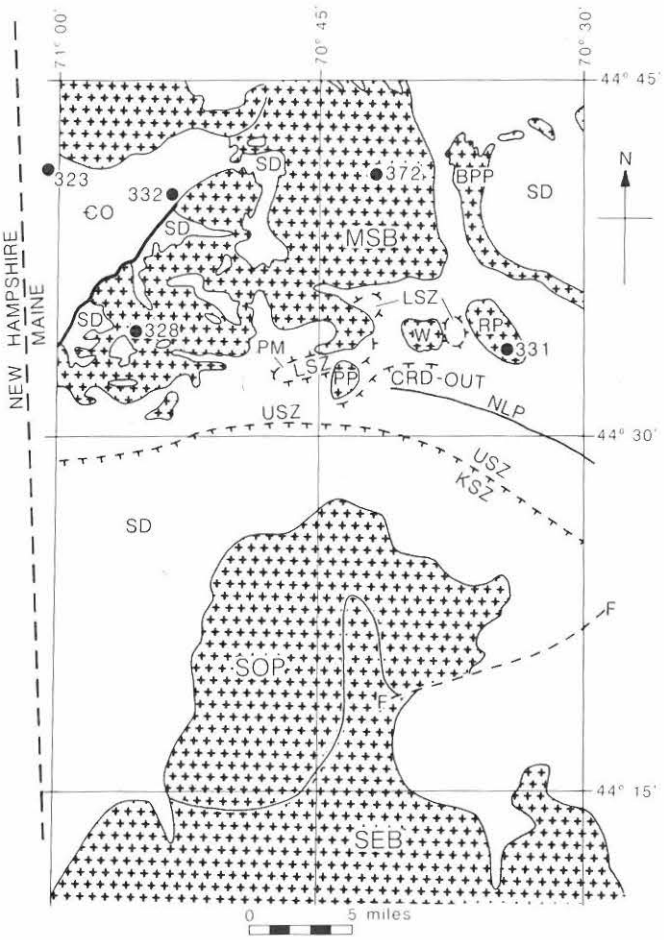

Figure 2. Metamorphic and plutonic map of region north of the Sebago batholith. (See Fig. 1 for index map). €O -- Cambro-Ordovician strata; SD -- Siluro-Devonian strata. Hachured line -- isograds; LSZ -- lower sillimanite, USZ -- upper sillimanite, and KSZ -- K-feldspar + sillimanite zones; CRD-out -- breakdown of cordierite. Heavy black line -- boundary of CO and SD strata. F -- fault. Pluses designate granitic bodies; MSB -- Mooselookmeguntic, BPP -- Bunder Pond, RP -- Rumford, W -- Whitecape, PP -- Plumbago, SOP -- Songo, and SEB -- Sebago intrusions. PM -- Puzzle Mountain. Solid circles with numbers give sample localities and hornblende ages of samples in Figure 4 and Table 2.

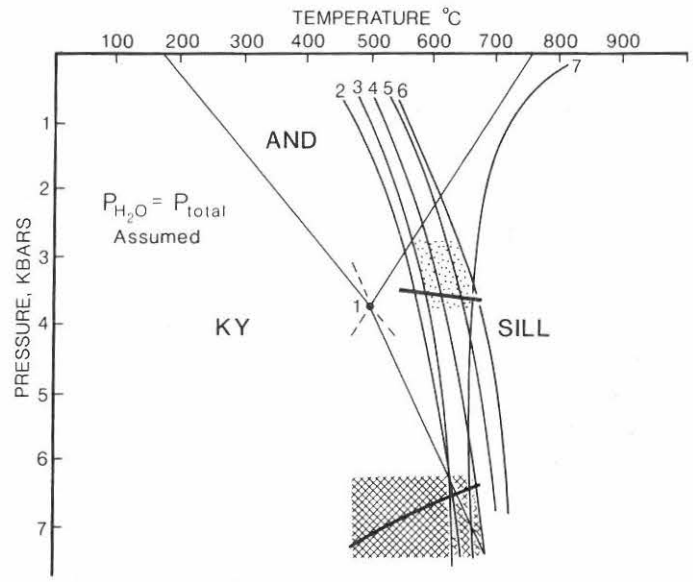

Figure 3. PT grid of equilibria relevant to the Sebago area with $P_{H2O} = P_{Total}$ after Guidotti et al., (1986).

(1) Alumino-silicate triple point, Holdaway (1971).
(2) Paragonite breakdown, Chatterjee (1972).
(3) St + Chl = Al-Sil + Bio, Guidotti (1974).
(4) St = Al-Sil + Gn + Bio, Hoschek (1969).
(5) Ab + Mu + Al-Sil + Ksp, Chatterjee and Froese (1975).
(6) Muscovite breakdown, Chatterjee and Johannes (1974).
(7) Granite minimum melt curve, Tuttle and Bowen (1958).

Stippled area -- range of PT conditions to the north of the batholith (see text). Cross-hatched area -- range of PT conditions to the south of the batholith. Heavy lines -- metamorphic field gradients.

sillimanite zone, many of the pelitic rocks are migmatitic and up to 50% of the rocks are pegmatitic. This mixed zone is also intruded by many small plutons and dikes, and the distinction between granitic rocks with metamorphic inclusions and metamorphic rocks with granitic inclusions is often arbitrary.

At the southern end of the Sebago batholith, fresh euhedral kyanite coexisting with biotite occurs, indicating that pressures of metamorphism were substantially higher than to the north (Thomson 1986; Thomson and Guidotti, this volume). Once again, the isograds are roughly parallel to the boundary of the batholith, but the mixed zone of pegmatitic and migmatitic bodies does not occur. The metamorphic grade again increases towards the batholith passing from garnet to staurolite to kyanite

TABLE 1. INFERRED TEMPERATURES OF METAMORPHISM FOR THE SEBAGO AREA.

	North	South
KF/US	660	
US/LS	615	625-700
LS/ST	590	600-625
KY/ST		500-600
ST/G		450-480

KF – K-feldspar + sillimanite
US – upper sillimanite
LS – lower sillimanite
ST – staurolite
KY – kyanite
G – garnet
Estimates for the south based on Thomson (1986) and Thomson and Guidotti (this volume).

to lower sillimanite zone over a total distance of 10 km. Temperatures of metamorphism range from 450°C in the garnet zone to 700°C in the lower sillimanite zone based on a petrogenetic grid approach (Thomson, 1986; Thomson and Guidotti, this volume). This is consistent with garnet-biotite geothermometry which gives a temperature range of 475-625°C. Pressures of metamorphism are less certain. A petrogenetic grid approach with $P_{H2O} = P_{Total}$ yielded pressures in the range of 6.0 to 7.0 kbar (Thomson, 1986; Thomson and Guidotti, this volume). However, if $P_{H2O} < P_{Total}$ is considered, pressures could be as low as 5.0 kbar. Again, the metamorphic field gradient is nearly isobaric. The conditions of metamorphism at the southern end are also shown in Figure 3 for $P_{H2O} = P_{Total}$, and the inferred temperatures of metamorphism for the various isograds are listed in Table 1.

GEOCHRONOLOGY

The ages of several plutons in western Maine have been previously determined by isotopic age dating. The crystallization age of the Mooselookmeguntic pluton was found to be 371±6 Ma using the Rb/Sr whole rock isochron method (Moench and Zartman, 1976). (All ages discussed in this paper are calculated using the currently accepted decay constants (Steiger and Jager, 1977)). Lux et al. (in press) used Rb/Sr whole rock isochron methods to determine an age of 382 Ma for the Songo pluton. The Sebago pluton has also been dated by both U/Pb and Rb/Sr methods and was found to have a crystallization age of 325 Ma by Aleinikoff et al. (1985) and 332 Ma by Hayward and Gaudette (1987). Two U/Pb isotopic ages for monazites from the Sebago batholith have also been measured (Aleinikoff et al., 1985). Both fell on concordia curves and gave ages of 272 and 282 Ma. These were interpreted to be cooling ages dating the time at which the samples cooled below 600°C, the closure temperature for Pb in monazites.

Lux and Guidotti (1985) argued that the K-feldspar + sillimanite-grade metamorphism that surrounds the Sebago batholith developed during Carboniferous rather than Devonian time. In support of that argument they presented $^{40}Ar/^{39}Ar$ ages for hornblendes from the southern Mooselookmeguntic pluton and the Songo pluton, which were approximately equal to, or distinctly younger than, the age of the Sebago pluton. They suggested that these hornblendes were reset during a period of Carboniferous metamorphism and that, during that metamorphism, the samples had been raised above 500°C, the closure temperature for Ar in hornblende.

Twenty-one new $^{40}Ar/^{39}Ar$ ages are presented here (Tables 2 and 3). Five of these ages are for hornblendes from the Mooselookmeguntic pluton, a small, related, satellite pluton, and the adjacent Ammonoosuc formation (Fig. 2). The release spectra are shown in Figure 4. The sixteen remaining ages are for micas (eight biotite-muscovite pairs) from the Sebago pluton and the release spectra are given in Figure 5. Analytical methods are presented elsewhere (Lux et al., in press).

The new isotopic ages for hornblende provide further evidence for Carboniferous metamorphism. With one exception, the ages of the hornblendes range from 328-332 Ma. These are equal to the crystallization age of the Sebago pluton to within the limits of analytical uncertainty. However, the one exception is perhaps the most important data point. The sample locality of this hornblende is in the Mooselookmeguntic pluton and lies to the north of the other four sample localities. It has a plateau age

TABLE 2. INCREMENTAL RELEASE DATA FOR HORNBLENDE SAMPLES FROM THE MOOSELOOKMEGUNTIC AND RUMFORD PLUTONS AND FROM THE AMMONOOSUC FORMATION.

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles ^{39}Ar	^{39}Ar %Total	%^{40}Ar Rad	K/Ca	Age(Ma)
			84-51-H		J = 0.004873			
765	380.8	1.873	0.5608	1.5	0.5	48.3	0.26	1295 ± 38
880	326.0	2.133	0.7296	2.2	0.7	29.2	0.23	778 ± 58
940	108.5	2.380	0.1726	3.9	1.3	46.4	0.21	448 ± 16
975	82.21	4.941	0.1445	4.1	1.4	44.6	0.01	321 ± 14
985	63.78	10.43	0.0897	5.4	1.8	59.9	0.05	309 ± 10
1005	50.35	13.58	0.0360	18.1	6.0	84.0	0.04	330 ± 6
1025	50.05	13.71	0.0353	26.8	8.9	84.5	0.04	329 ± 4
1040	48.41	14.38	0.0290	35.2	11.7	88.6	0.03	331 ± 4
1045	48.51	14.19	0.0283	30.5	10.1	88.7	0.03	334 ± 5
1060	47.75	14.32	0.0274	33.1	11.0	89.4	0.03	330 ± 4
1070	46.27	13.97	0.0219	38.2	12.7	92.1	0.04	331 ± 4
1080	46.18	13.73	0.0213	27.0	8.9	92.1	0.04	331 ± 4
1095	45.77	13.92	0.0212	22.8	7.5	92.5	0.04	329 ± 5
1110	46.20	14.96	0.0198	25.8	8.5	95.1	0.03	338 ± 4
Fuse	46.16	15.17	0.0192	27.3	9.1	95.2	0.03	337 ± 4
TOTAL				308.8	100.0			341 ± 3
PLATEAU AGE								331 ± 3

Moles ^{39}Ar has been multiplied times 10^{14}

TABLE 2. (CONTINUED)

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles ^{39}Ar	^{39}Ar %Total	%^{40}Ar Rad	K/Ca	Age(Ma)
			84-55-H		J = 0.004833			
880	223.3	1.3248	0.2559	6.6	0.6	56.5	0.39	937 ± 17
960	85.76	1.138	0.1080	10.7	0.9	54.2	0.43	418 ± 8
985	79.12	2.356	0.1247	7.6	0.6	47.3	0.21	337 ± 9
1005	74.77	4.682	0.1184	7.1	0.6	49.1	0.10	321 ± 8
1020	69.77	6.320	0.0911	8.6	0.7	57.9	0.08	344 ± 7
1025	57.79	6.656	0.0418	20.6	1.7	74.1	0.07	363 ± 5
1040	57.47	6.697	0.0359	19.9	1.7	76.6	0.07	373 ± 8
1050	52.61	6.462	0.0192	46.5	3.9	83.6	0.08	374 ± 5
1065	51.43	6.356	0.0152	78.8	6.6	85.3	0.08	374 ± 5
1075	50.32	6.268	0.0123	104.9	8.8	86.7	0.08	372 ± 4
1085	49.61	6.156	0.0120	208.0	17.5	87.6	0.08	371 ± 4
1110	48.81	6.080	0.0086	308.0	25.9	88.4	0.08	369 ± 4
Fuse	48.53	6.189	0.0070	359.7	30.3	89.4	0.08	370 ± 4
TOTAL				1187.20	100.0			373
PLATEAU AGE								372 ± 4
			82-30-H		J = 0.004800			
960	101.8	2.053	0.1759	20.0	3.8	49.1	0.24	388 ± 8
985	108.3	10.66	0.2224	4.8	0.9	40.0	0.05	343 ± 12
1000	105.0	12.01	0.2233	3.7	0.7	38.0	0.04	319 ± 17
1010	72.16	13.91	0.1077	8.2	1.6	57.3	0.04	330 ± 9
1025	53.48	14.37	0.0447	23.0	4.4	77.3	0.03	330 ± 5
1040	47.83	14.38	0.0258	61.5	11.8	86.2	0.03	329 ± 4
1050	49.93	14.47	0.0328	36.1	6.9	82.7	0.03	330 ± 5
1065	49.62	14.47	0.0306	56.7	10.9	83.9	0.03	332 ± 4
1075	56.11	14.23	0.0516	56.2	10.8	74.4	0.03	334 ± 5
1085	48.36	14.23	0.0263	78.1	15.0	86.1	0.03	332 ± 4
1100	49.50	14.08	0.0294	49.8	9.6	84.6	0.03	334 ± 4
1125	47.03	14.34	0.0214	51.6	9.9	88.8	0.03	333 ± 4
Fuse	45.32	14.14	0.0152	70.6	13.6	92.4	0.03	334 ± 4
TOTAL				520.2	100.0			334
PLATEAU AGE								332 ± 4
			82-33-H		J = 0.004915			
925	720.8	8.891	1.8140	0.6	0.2	25.7	0.06	1174 ± 73
970	419.3	8.762	0.9192	0.8	0.3	35.4	0.06	993 ± 55
990	208.5	11.22	0.5178	1.0	0.3	27.0	0.04	444 ± 37
1010	112.8	14.61	0.2511	1.7	0.6	35.2	0.03	325 ± 20
1025	105.3	13.97	0.2348	1.6	0.5	35.1	0.04	304 ± 2
1050	46.83	14.42	0.0237	26.2	8.9	87.3	0.03	334 ± 4
1075	43.89	13.55	0.0190	52.2	17.7	89.5	0.04	322 ± 4
1085	43.29	14.69	0.0165	57.4	19.5	91.3	0.03	324 ± 4
1100	43.96	13.48	0.0177	43.1	14.6	90.3	0.04	325 ± 4
1110	44.46	13.42	0.0184	36.9	12.5	90.0	0.04	327 ± 4
1135	43.73	13.18	0.0154	54.7	18.5	91.8	0.04	328 ± 4
Fuse	45.74	13.04	0.0209	18.6	6.3	88.6	0.04	311 ± 4
TOTAL				294.8	100.0			330
PLATEAU AGE								323 ± 3
			82-40-H		J = 0.006445			
765	1257	3.673	0.9154	0.6	0.1	78.5	0.13	3606 ± 52
800	214.0	1.190	0.1262	5.0	0.7	82.6	0.41	1373 ± 15
940	107.8	1.831	0.0654	7.7	1.1	82.1	0.27	816 ± 10
960	67.30	2.565	0.0670	5.5	0.8	70.6	0.19	484 ± 7
975	47.89	3.704	0.0520	5.7	0.8	67.9	0.13	347 ± 6
985	39.12	4.726	0.0288	9.7	1.4	78.3	0.10	329 ± 5
990	37.06	5.042	0.0210	12.1	1.7	83.2	0.10	332 ± 6
1000	34.73	5.464	0.0134	16.4	2.4	88.6	0.09	332 ± 5

Moles ^{39}Ar has been multiplied times 10^{14}

TABLE 2. (CONTINUED)

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles ^{39}Ar	^{39}Ar %Total	%^{40}Ar Rad	K/Ca	Age(Ma)
		82-40-H			J = 0.006445			
1005	33.91	5.640	0.0105	19.8	2.9	90.8	0.09	332 ± 4
1010	32.82	5.851	0.0069	35.7	5.2	93.8	0.08	332 ± 4
1015	32.73	5.893	0.0065	54.3	7.8	94.1	0.08	332 ± 4
1025	32.81	5.885	0.0080	37.7	5.4	92.8	0.08	329 ± 4
1030	32.49	5.929	0.0073	44.7	6.4	93.4	0.08	328 ± 4
1040	32.62	5.943	0.0082	36.2	5.2	92.6	0.08	327 ± 4
1045	32.22	6.039	0.0063	36.2	5.2	94.2	0.08	328 ± 4
1050	31.95	6.073	0.0055	45.8	6.6	94.9	0.08	328 ± 4
1065	31.88	6.043	0.0053	36.9	5.3	95.1	0.08	328 ± 4
1070	31.82	6.119	0.0055	37.1	5.3	94.9	0.08	327 ± 4
1075	31.75	6.169	0.0056	36.3	5.2	94.8	0.08	326 ± 4
1080	32.16	6.202	0.0063	35.6	5.1	94.2	0.08	328 ± 4
1085	32.00	6.234	0.0055	26.7	3.8	94.9	0.08	328 ± 4
1090	31.95	6.130	0.0050	59.2	8.5	95.3	0.08	329 ± 4
1125	31.99	6.062	0.0038	58.2	8.4	96.5	0.08	333 ± 4
1135	33.74	5.791	0.0097	21.5	3.1	91.5	0.08	333 ± 4
Fuse	37.95	5.130	0.0257	9.2	1.3	80.0	0.10	327 ± 5
TOTAL				693.9	100.0			346
PLATEAU AGE								328 ± 2

Moles ^{39}Ar has been multiplied times 10^{14}

TABLE 3. INCREMENTAL RELEASE AND TOTAL FUSION DATA FOR BIOTITES (B) AND MUSCOVITES (M) FROM THE SEBAGO PLUTON.

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles ^{39}Ar	^{39}Ar %Total	%^{40}Ar Rad	Age(Ma)
		SBG-3-B			J = 0.005758		
800	25.48	1.497	8.073	152.2	9.6	90.5	224.9 ± 2.4
880	25.31	0.1906	1.710	400.7	25.4	97.9	240.5 ± 2.3
960	25.62	0.4734	1.801	271.6	17.2	97.8	243.1 ± 2.4
1005	25.26	0.4863	0.5956	377.3	23.9	99.2	243.1 ± 2.3
1055	25.02	0.1265	0.1006	286.6	18.2	99.7	242.2 ± 2.3
Fuse	25.01	0.2253	0.0268	89.9	5.7	98.6	242.3 ± 2.6
TOTAL				1578.4	100.0		240.5 ± 2.3
PLATEAU AGE							242.2 ± 3.3
		SBG-4-B			J = 0.005753		
800	24.53	1.828	4.8306	173.3	9.7	94.0	224.8 ± 3.7
890	23.87	0.3454	0.8669	459.1	25.8	98.8	229.4 ± 2.4
960	23.85	0.6561	0.8844	236.5	13.3	98.7	229.2 ± 2.3
1005	23.56	0.3675	0.3440	392.1	22.0	99.4	228.0 ± 2.2
1055	23.58	0.4582	0.2304	367.5	20.6	99.5	228.5 ± 2.2
Fuse	24.07	1.102	1.007	151.1	8.5	98.6	230.9 ± 2.6
TOTAL				1779.6	100.0		228.6 ± 2.4
PLATEAU AGE							229.2 ± 3.3
		SBG-5-B			J = 0.00573		
825	25.29	0.3016	5.446	206.2	12.1	93.5	229.2 ± 2.3
915	24.21	0.09377	0.6039	410.4	24.0	99.1	232.4 ± 2.2
985	24.33	0.2254	1.275	275.5	16.1	98.3	231.7 ± 2.3
1030	24.11	0.1284	0.3107	397.0	23.2	99.5	232.3 ± 2.3
1080	24.16	0.2315	0.2732	325.8	19.1	99.5	232.8 ± 2.2
Fuse	24.72	0.2258	2.015	93.4	5.5	97.4	233.3 ± 2.7
TOTAL				1708.4	100.0		232.0 ± 2.3
PLATEAU AGE							232.5 ± 2.4

$^{37}Ar/^{39}Ar$ has been multiplied times 10^2 $^{36}Ar/^{39}Ar$ has been multiplied times 10^3 Moles ^{39}Ar has been multiplied times 10^{14}

TABLE 3. (CONTINUED)

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles ^{39}Ar	^{39}Ar %Total	%^{40}Ar Rad	Age(Ma)
		SBG-6-B			J = 0.005708		
800	24.66	0.4329	7.465	187.1	13.9	90.9	217.2 ± 2.2
880	24.53	0.1512	0.9743	316.9	23.5	98.7	233.4 ± 2.2
960	24.87	0.4543	2.003	221.2	16.4	97.5	233.8 ± 2.3
1000	24.44	0.2633	0.7141	334.5	24.8	99.0	233.4 ± 2.2
1050	24.29	0.1494	0.2879	238.3	17.7	99.5	233.1 ± 2.3
Fuse	23.68	0.8912	3.791	51.7	3.8	95.1	218.2 ± 15.9
TOTAL				1349.6	100.0		230.6 ± 2.8
PLATEAU AGE							233.4 ± 1.8
		SBG-7-B			J = 0.005729		
810	24.55	0.3997	8.354	133.8	6.4	89.8	214.6 ± 2.8
890	24.71	0.1058	1.264	641.4	30.8	98.3	235.1 ± 2.2
975	25.38	0.2552	1.649	311.0	14.9	97.9	240.1 ± 2.4
1025	24.99	0.1370	1.044	601.4	28.8	98.6	238.2 ± 2.3
Fuse	24.80	0.3103	0.5281	398.2	19.1	99.2	237.9 ± 2.3
TOTAL				2085.8	100.0		236.0 ± 2.3
PLATEAU AGE							238.8 ± 3.6
		SBG-8-B			J = 0.005718		
815	26.14	0.1954	7.119	103.7	6.3	91.8	231.9 ± 3.2
890	25.92	0.1764	2.008	488.7	29.5	97.6	243.7 ± 2.4
985	25.87	0.1694	1.197	343.8	20.7	98.5	245.3 ± 2.4
1025	25.73	0.04781	0.7406	364.4	22.0	99.0	245.3 ± 2.4
Fuse	25.63	0.06597	0.6286	358.8	21.6	99.1	244.7 ± 2.4
TOTAL				1659.3	100.0		243.9 ± 2.4
PLATEAU AGE							244.8 ± 2.8
		SBG-9-B			J = 0.005733		
800	26.29	0.5400	6.229	94.1	5.9	92.9	236.3 ± 7.5
890	25.71	0.1907	2.529	565.8	35.8	96.9	240.9 ± 4.6
980	25.72	0.3723	1.866	352.3	22.3	97.7	242.8 ± 6.3
1025	25.39	0.4298	0.4747	398.6	25.2	99.3	243.6 ± 10.5
Fuse	25.42	0.7184	1.718	171.7	10.8	97.9	240.5 ± 12.6
TOTAL				1582.4	100.0		241.7 ± 7.5
PLATEAU AGE							242.0 ± 4.1
		SBG-10-B			J = 0.005767		
800	26.61	0.3606	9.208	95.7	5.7	89.6	232.5 ± 6.1
890	25.00	0.08246	1.411	522.1	31.3	98.2	238.8 ± 3.4
990	25.11	0.07546	1.279	409.0	24.5	98.3	240.2 ± 4.9
1025	24.52	0.01216	0.6592	389.0	23.3	99.0	236.4 ± 3.0
Fuse	25.20	0.06797	0.8861	251.9	15.1	98.8	242.1 ± 2.9
TOTAL				1667.7	100.0		238.7 ± 3.7
PLATEAU AGE							238.5 ± 5.0
		SBG-3-M			J = 0.005753		
840	28.51	0.2755	10.99	76.0	3.5	88.5	244.4 ± 3.2
915	26.86	0.05356	4.678	479.2	21.9	94.7	246.4 ± 2.3
970	26.18	0.03200	2.487	482.9	22.1	97.0	246.1 ± 2.4
1017	25.99	0.03000	1.715	460.0	21.1	97.9	246.4 ± 2.3
1090	25.78	0.03151	0.9083	641.5	29.4	98.8	246.7 ± 2.4
Fuse	27.38	0.1309	5.437	44.1	2.0	94.0	249.0 ± 3.9
TOTAL				2183.7	100.0		246.4 ± 2.4
PLATEAU AGE							246.4 ± 1.8

$^{37}Ar/^{39}Ar$ has been multiplied times 10^2 $^{36}Ar/^{39}Ar$ has been multiplied times 10^3 Moles ^{39}Ar has been multiplied times 10^{14}

TABLE 3. (CONTINUED)

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles ^{39}Ar	^{39}Ar %Total	%^{40}Ar Rad	Age(Ma)
		SBG-4-M			J = 0.005753		
840	27.63	4.194	15.63	47.9	2.4	83.2	223.9 ± 13.1
915	25.54	1.120	3.235	135.0	6.7	96.1	238.3 ± 2.4
975	25.44	0.1598	2.052	579.3	28.9	97.5	240.6 ± 3.0
1015	25.22	0.3890	1.479	338.2	16.9	98.1	240.1 ± 2.5
1085	25.03	0.7289	0.06826	621.6	31.0	99.0	240.5 ± 3.0
Fuse	25.58	8.285	1.144	284.3	14.2	98.6	244.4 ± 2.8
TOTAL				2006.3	100.0		240.5 ± 3.1
PLATEAU AGE							240.4 ± 1.8
		SBG-5-M			J = 0.00571		
800	26.02	0.6518	12.72	17.1	0.7	85.4	215.5 ± 6.2
890	26.26	0.8031	6.364	79.4	3.3	92.7	234.8 ± 2.7
950	26.37	0.3024	4.703	350.9	14.5	94.6	240.2 ± 2.3
1000	25.40	0.07460	1.965	792.6	32.9	97.6	238.8 ± 2.3
1065	25.42	0.3050	1.829	460.8	19.1	97.7	239.3 ± 2.3
Fuse	25.42	3.368	1.005	710.9	29.5	98.7	241.5 ± 2.3
TOTAL				2411.5	100.0		239.6 ± 2.4
PLATEAU AGE							239.9 ± 3.6
		SBG-6-M			J = 0.005719		
840	29.75	0.2630	19.10	43.8	2.6	80.9	232.6 ± 6.2
905	26.60	0.01515	4.817	99.8	5.8	94.5	242.4 ± 2.5
970	26.43	0.06616	4.408	503.9	29.4	94.9	241.9 ± 2.3
1010	25.94	0.05964	2.700	324.9	19.0	96.8	242.0 ± 2.3
1070	25.64	0.006918	1.636	424.7	24.8	98.0	242.1 ± 2.4
Fuse	25.55	0.09338	0.9723	314.7	18.4	98.7	243.1 ± 2.4
TOTAL				1711.9	100.0		242.0 ± 2.5
PLATEAU AGE							242.3 ± 2.2
		SBG-7-M			J = 0.005729		
870	28.54	0.3791	12.59	61.6	3.2	86.8	239.5 ± 3.8
950	25.55	0.006895	1.985	840.7	43.3	97.6	240.8 ± 2.3
1000	25.46	0.7322	1.422	433.6	22.3	98.2	241.5 ± 2.3
1050	25.41	0.1553	1.339	368.8	19.0	98.3	241.3 ± 2.4
Fuse	25.33	2.547	0.4974	239.0	12.3	99.3	242.8 ± 2.7
TOTAL				1943.7	100.0		241.2 ± 2.4
PLATEAU AGE							241.2 ± 3.6
		SBG-8-M			J = 0.005705		
880	27.23	1.263	9.629	76.9	5.1	89.4	234.7 ± 3.2
930	26.32	0.5054	3.284	280.8	18.5	96.2	243.4 ± 2.4
990	25.81	0.29674	1.499	472.5	31.2	98.1	243.4 ± 2.4
1050	25.70	0.64622	1.407	342.8	22.6	98.2	242.8 ± 2.5
Fuse	25.74	7.388	0.4780	343.0	22.6	99.3	245.6 ± 2.4
TOTAL				1515.9	100.0		243.3 ± 2.4
PLATEAU AGE							243.2 ± 2.0
		SBG-9-M			J = 0.005762		
880	28.41	0.9474	12.83	101.7	8.8	86.5	239.0 ± 10.7
940	26.28	0.1752	3.784	519.9	45.1	95.6	243.9 ± 2.3
1000	25.72	0.1362	2.438	362.7	31.4	97.0	242.4 ± 2.5
Fuse	25.24	0.1670	0.2299	169.3	14.7	99.6	244.0 ± 2.7
TOTAL				1153.6	100.0		243.0 ± 3.2
PLATEAU AGE							243.4 ± 3.0

$^{37}Ar/^{39}Ar$ has been multiplied times 10^2 $^{36}Ar/^{39}Ar$ has been multiplied times 10^3 Moles ^{39}Ar has been multiplied times 10^{14}

TABLE 3. (CONTINUED)

Temp °C	$^{40}Ar/^{39}Ar$	$^{37}Ar/^{39}Ar$	$^{36}Ar/^{39}Ar$	Moles ^{39}Ar	^{39}Ar %Total	%^{40}Ar Rad	Age(Ma)
		SBG-10-M			J = 0.005759		
880	28.21	0.3592	10.56	83.7	4.5	88.8	243.1 ± 2.8
950	26.66	0.09105	5.233	251.1	13.5	94.1	243.3 ± 2.4
990	25.77	0.01400	2.600	690.0	37.0	96.9	242.3 ± 2.3
1060	25.55	0.03828	1.394	373.1	20.0	98.2	243.5 ± 2.4
Fuse	25.49	0.03960	1.106	465.4	25.0	98.6	243.8 ± 2.4
TOTAL				1863.3	100.0		243.1 ± 2.4
PLATEAU AGE							243.2 ± 2.4

$^{37}Ar/^{39}Ar$ has been multiplied times 10^2 $^{36}Ar/^{39}Ar$ has been multiplied times 10^3 Moles ^{39}Ar has been multiplied times 10^{14}

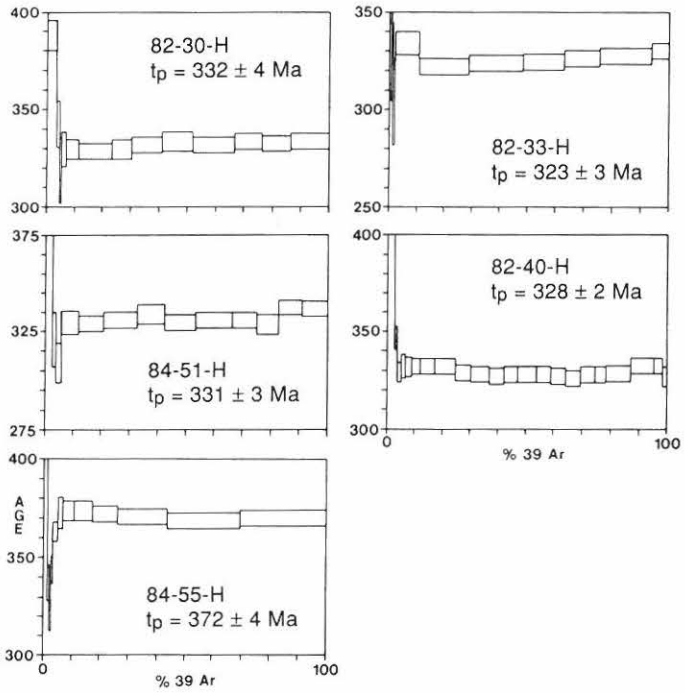

Figure 4. Age spectra of hornblende from samples of the Mooselookmeguntic pluton, a small satellite pluton, and the Ammonoosuc Formation. Note that the vertical scales are variable and that small increments with anomalously old ages have been omitted. The horizontal scales are identical. Individual box heights represent 2 sigma analytical uncertainties.

of 372 Ma which is identical to the crystallization age of the Mooselookmeguntic pluton. This demonstrates that ambient temperature was below the closure temperature for Ar in hornblende, 500°C, and that the pluton cooled rapidly below that temperature following emplacement. It also delineates the northern limit of the Carboniferous metamorphic event. Apparently, temperatures in excess of 500°C were experienced only up to 30-50 km north of the present-day surface outcrop of the Sebago batholith.

Mica ages from throughout the Mooselookmeguntic pluton are all younger than 305 Ma (Lux, 1986), implying that ambient temperature in the region before 305 Ma was above 350°C. Thus, ambient temperature at the time of emplacement of the Sebago batholith is constrained to be between approximately 350°C and 500°C.

The new $^{40}Ar/^{39}Ar$ ages for muscovites and biotites from the Sebago batholith range in age from 246 to 240 Ma, and 245 to 229 Ma respectively. These ages are interpreted to date the time that the micas last cooled through their closure temperatures, 280°C for biotite (Harrison et al., 1985) and 350°C for muscovite (Cliff, 1985), in response to regional uplift and erosion. There is a general trend towards younger ages to the southwest indicating that either: (1) the region uplifted more rapidly in the northeast, or (2) the region was buried more deeply to the southwest. The second explanation requires differential uplift across the region, with the present erosion surface exposing deeper levels of the crust towards the southwest. This is consistent with the geometry of the region (De Yoreo et al., 1989; Lux et al., 1986; Hon and Schulman, 1983) as is discussed below.

The average discordance between the muscovite and biotite ages is 5 Ma implying that, on average, the region cooled at a rate of 15°C/MY between approximately 245 Ma and 230 Ma. Previous mica age discordance determinations on nearby, shallower plutons indicate average cooling rates for the region of 1-3°C/MY from 370 Ma to 270 Ma and 6-9°C/MY from 270 Ma to 250 Ma (De Yoreo et al., 1989). These cooling rates will be used to constrain the erosion rates used in our thermal model.

GEOMETRY OF THE PLUTONS

In the first regional gravity survey of Maine, Kane and Bromery (1968) noted that the plutons in the high-grade metamorphic terrane had very little gravity signature and, hence, were likely to be relatively thin bodies. Later gravity studies by Hodge et al. (1982) in Maine and by Neilson et al. (1976) in New Hampshire comfirmed this observation. They concluded that the thickness of these plutons ranged from 0.5 km to 2 km. Hodge et al. (1982) estimated the thickness of the Sebago batholith to

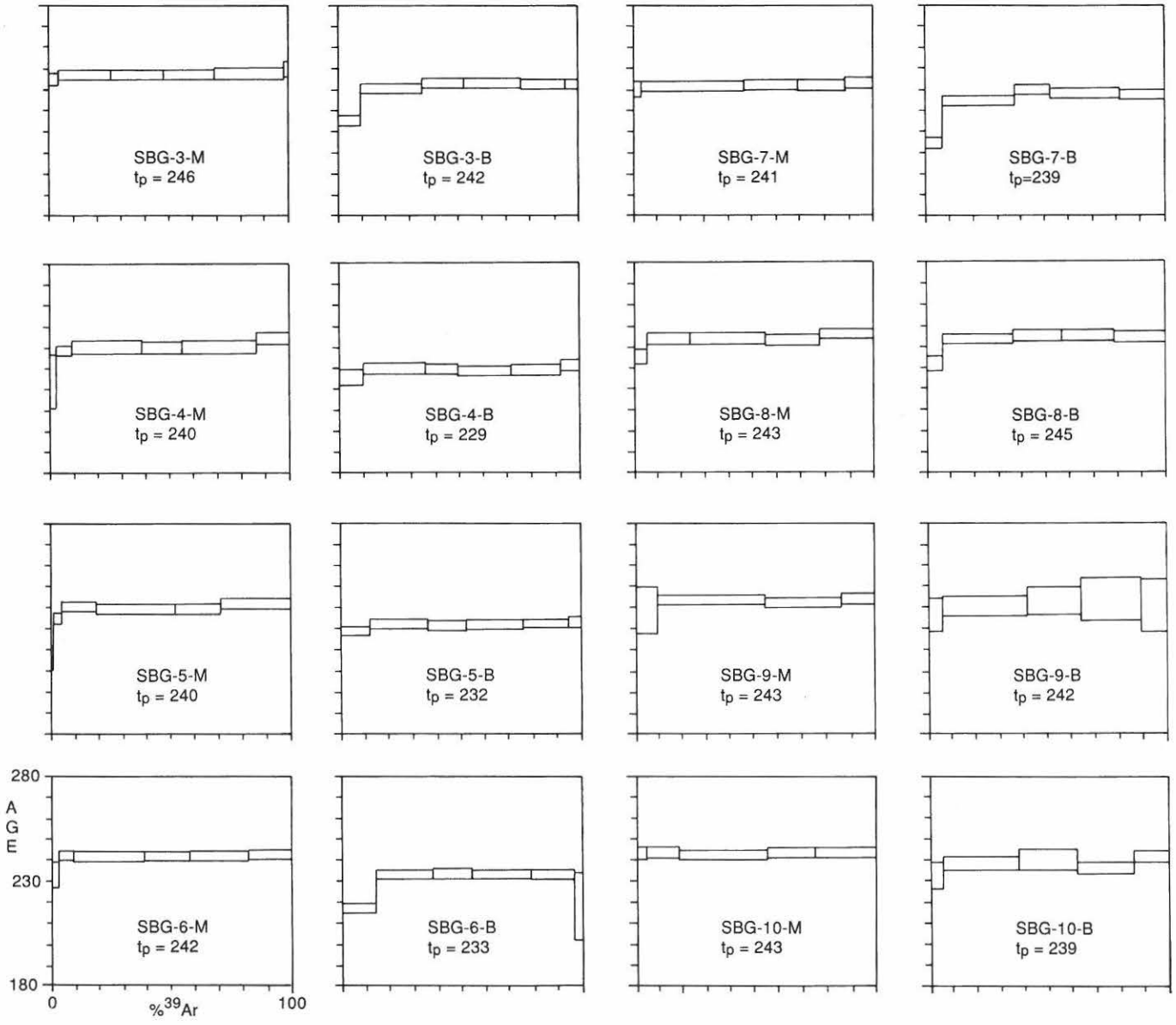

Figure 5. Age spectra of biotite (B) and muscovite (M) from samples of the Sebago pluton. The vertical and horizontal scales are identical and are as shown in the lower left. Individual box heights represent 2 sigma uncertainties.

be approximately 1 km. Detailed gravity surveys across a number of the plutons in western Maine (Carnese, 1983) showed them to be thin tabular masses that homotropically dip to the northeast at about 3 degrees. Many extend far beyond their surface contacts. For example, although the Mooselookmeguntic and Redington plutons are separated on the surface by 10-15 km, gravity measurements suggest that the Mooselookmeguntic pluton is approximately 2 km thick and actually extends underneath the Redington pluton, staying within a kilometer of the surface in the intervening region. In addition, peak pressures of metamorphism from northeastern Maine to central Massachusetts along the Kearsarge-central Maine synclinorium imply a gentle northeastward plunge of approximately 3 degrees (De Yoreo et al., 1989). It is reasonable to assume that the Sebago batholith also dips to the northeast at about 3 degrees and extends beyond its surface contact. Given the present southwest-northeast surface exposure of 45 km for the Sebago batholith, this dip results in an estimated thickness of about 2 km. In its present day configuration the batholith should be very thin near its southwest contact and thicken to nearly 2 km at the northeast contact and beyond (see Fig. 6). The average thickness below the exposed surface is then about 1 km, in agreement with the estimate of Hodge et al. (1982). With this geometry, the metamorphic terrane to the north of the Sebago batholith lies above its upper surface while that to the south once lay below its lower surface (Fig. 6).

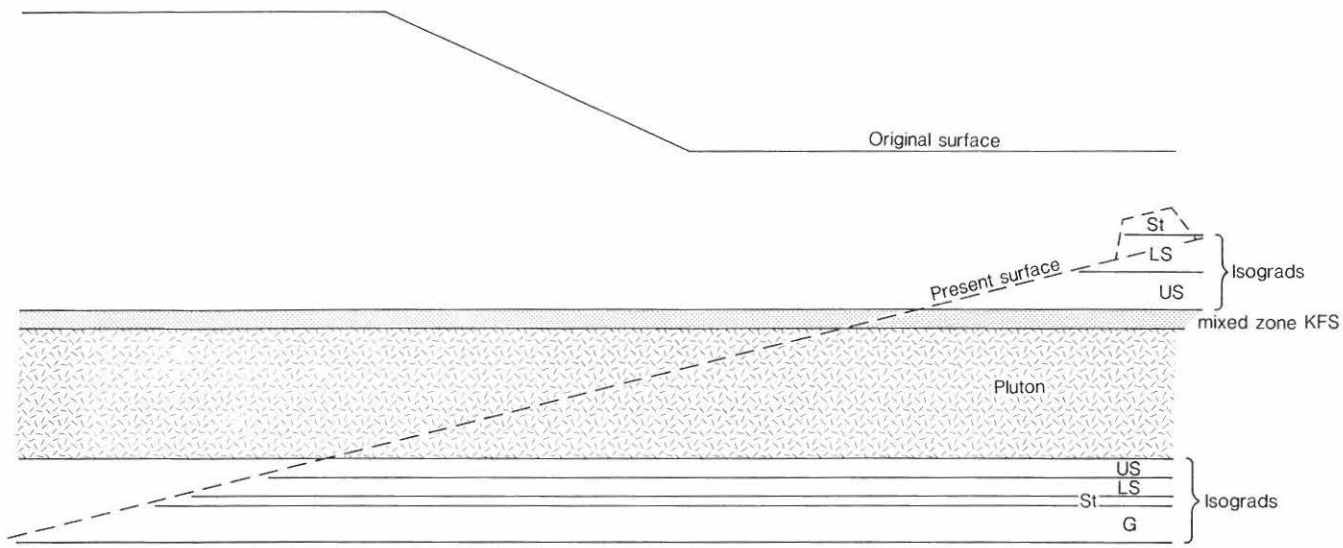

Figure 6. Schematic diagram showing proposed geometry for the Sebago batholith and surrounding metamorphic terrane. Original surface allows for increasing topographic relief to the southwest. Metamorphic zones: G -- garnet, St -- staurolite, LS -- lower sillimanite, US -- upper sillimanite, KFS -- K-feldspar + sillimanite.

Conservative estimates of 3.8 and 5.5 kbar for the northern and southern ends of the Sebago batholith respectively imply a change in depth of burial of about 6 km from north to south. However, the assumption that the pluton dips at 3 degrees can account for only half of this change. Errors in the pressure estimates or an increase in topographic relief to the southwest as shown in Figure 6 are possible sources of this discrepancy.

THERMAL MODEL

The central question of this paper is: was the Sebago batholith capable of transferring the amount of heat necessary to produce the surrounding regional, high-grade metamorphism, or were other sources such as hot, hydrous fluids or high, transient, mantle flux required? In order to address this question it is necessary to solve the heat diffusion equation (Carslaw and Jaeger, 1959) which for one dimension is given by:

$$(1)\quad \partial^2 T(x,t)/\partial x^2 - (1/K)\partial T(x,t)/\partial t = (u/K)\partial T(x,t)/\partial x - A(x)/\Lambda$$

where T(x, t) is the temperature at position x and time t, K is the thermal diffusivity, u is the erosion rate, Λ is the thermal conductivity, and A(x) is the distribution of heat production in the crust due to radiogenic heating.

We seek a solution of Equation 1 for the solidification and cooling of a liquid in a semi-infinite solid with an arbitrary initial temperature distribution, T(x,0). In general, this is an intractable problem and we are forced to make a number of assumptions. However, the geometry of the problem at hand leads to enormous simplification and enables us to generate a solution that is accurate to a good approximation. The least lateral dimension of the batholith is greater than 40 km, the surface area is greater than 2.5×10^3 km^2, and the thickness is only 2 km. Furthermore, the petrologic studies discussed previously indicate that the upper surface of the batholith at the time of emplacement was at approximately 14 km depth. Thus, three simplifications can be made. First, since the ratio of heat that flows perpendicular to the sheet to that which flows parallel to the sheet is on the order of the square of the width divided by the square of the thickness, the problem is essentially one dimensional. Second, in the time it takes for the sheet to solidify, the heat diffuses roughly one pluton thickness away from the boundary of the pluton, allowing us to ignore the boundary at the earth's surface. We assume that the magmatic sheet is in an infinite medium up until the time of solidification, t_s. Third, since the pluton is thin, little error is made in assuming that the temperature of the country rock affected during solidification is initially independent of position, i.e. $T(x, t \leq t_s) = T_o$

This problem has been studied in detail by numerous authors (Jaeger, 1964; Carslaw and Jaeger, 1959; Irvine, 1970). For the situation depicted in Figure 7, the solution for $t \leq t_o$ is given by

$$(2)\quad T(x,t) = T_o + (T_m - T_o)(\mathrm{erfc}(x/(4Kt)^{0.5}))/(1 + \mathrm{erf}\lambda)\quad t \leq t_s,\ x \geq x_m$$
$$x_m = -(4Kt)^{0.5}\lambda$$
$$t_s = d^2/(16K\lambda^2)$$

where T_o is the temperature of the country rock, T_m is the temperature of the magma, x_m is the position of the solidification surface, d is the thickness of the pluton, and λ is given by

$$(3)\quad L\pi/c(T_m - T_o) = e^{-\lambda 2}/\lambda(1 + \mathrm{erf}\lambda)$$

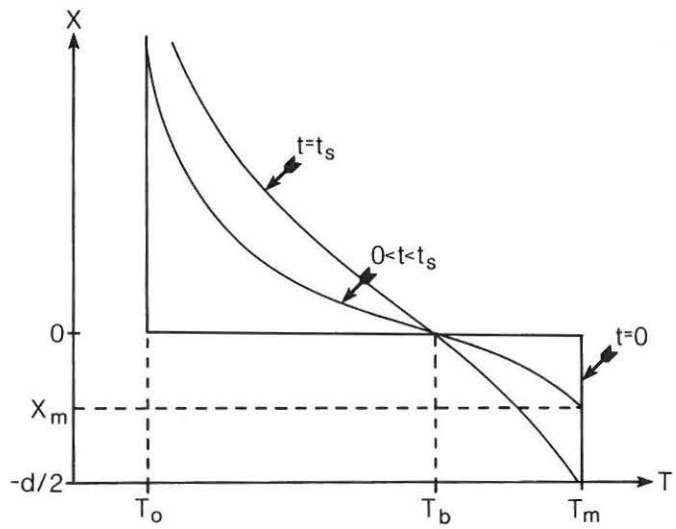

Figure 7. Temperature distributions near a solidifying igneous slab of thickness d with an initial temperature, T_m, in an infinite solid with initial temperature, T_o. $t = 0$: $T = T_o$ for $x > 0$, $x < -d$ and $T = T_m$ for $-d < x < 0$. $0 < t < t_s$: $T = T_m$ for $-d + x_m < x < x_m$. $t = t_s$: the solidification boundary, x_m, is at $x = -d/2$. $T = T_b$ at the pluton country-rock boundary for $0 < t \leq t_s$.

Here L and c are the latent heat and the specific heat of the magma respectively. Equation 2 assumes that the thermal properties of the magma and the country rock are identical. The pluton solidifies symmetrically from top and bottom.

For $t > t_s$ the solution is obtained numerically by applying a Crank-Nicholson implicit finite difference technique (Crank, 1981) to Equation 1, with $T'(x,t_s) = T(x,0) + T(x,t_s) - T_o$ as the initial temperature distribution.

The values of a number of parameters need to be specified. The thickness of the sheet, d, is taken to be 2 km. The thermal diffusivity, thermal conductivity, and the specific heat are given values for "average" crustal rocks with $K = 32\ km^2/MY$, $\Lambda = 2.5$ W/mK, and $c = 1.1$ J/gK. (For a discussion of the values of these parameters see England and Thompson, 1984). Based on the latent heats of oxides in the CRC Handbook of Chemistry and Physics (Weast and Astle, 1982), a value of 320 J/g is a reasonable average for L.

Varying K, Λ, c, and L over the range of relevant values has little effect on the result provided T_o and T_m are not changed. In contrast, variations in T_m and T_o can have significant effects. Fortunately, studies on plutons of similar composition and depth of emplacement provide reasonable limits on T_m. Typical emplacement temperatures for S-type granites are in the range of 750-850°C (Clemens et al., 1986; Clemens and Wall, 1981; Wyborn et al., 1981; Clemens, 1984). A value of 800°C is taken for T_m.

The initial temperature distribution in the country rock, T(x,0), is the most uncertain parameter. It is a function not only of the thermal properties of the rocks, but of the mantle heat flux, Q_m, the distribution of radiogenic heat producing elements in the crust, A(x), and the burial and erosion history of the region.

For the thermal diffusivity and the thermal conductivity we again choose values of 32 km^2/MY and 2.5 W/mK. The mantle heat flux is assumed to be that of present day New England for which $Q_m = 35\ mW/m^2$ (Birch et al., 1968; Jaupart et al., 1981).

The burial history is one of sedimentation from about 440 to 400 Ma followed by rapid Acadian thickening at about 400 Ma. Some 10-15 km of deep-water sediments were deposited prior to the Acadian orogeny (Moench and Zartman, 1976). The pressures of metamorphism in the region indicate initial depths of burial of the present erosion surface of between 10 to 15 km (Holdaway et al., 1987) and the thickness of the present day synclinorium is 10-12 km (Stewart et al., 1985). Thus the sedimentary pile was roughly doubled in thickness by Acadian folding and thrusting. For our model we take the sedimentation rate to be 0.3 km/MY for 42 million years (442-400 Ma). The sediments are homogeneously thickened by a factor of two instantaneously at 42 million years (400 Ma).

In order to avoid overestimating T_o and hence the thermal effect of the Sebago batholith, these sediments were modeled as being deposited on an old craton. In the model, they are deposited onto a basement whose volumetric heat production is similar to that of the Precambrian basement of the Adirondacks for which the total contribution to the surface heat flow from crustal radiogenic sources is 10 mW/m^2 (Birch et al., 1968). If this is distributed uniformly over 30 km, then $A(x) = A_o = 0.33$ uW/m^3.

The heat production in the sediments is then constrained by the present day distribution of crustal heat production for New England for which, on average, $A_o = 4.2$ uW/m^3 distributed uniformly over 7 km (Birch et al., 1968). In the model presented here, $A_o = 2.25$ W/m^3 for the sediments. After thickening and erosion, the average present day *contribution* to surface heat flow from crustal sources is properly obtained. The present day *distribution* of crustal heat production is thought to be a result of upward enrichment of radiogenic elements by the metamorphism and plutonism following the Acadian deformation.

As discussed above, the cooling rate, up until 325 Ma, at the depths of interest, was about 1-2°C/MY. Since the ambient temperature was between 300 and 500°C, the average gradient was 20-35°C/km. Geothermal gradients are likely to be steepest near the surface, so we take, as a starting point, an average gradient to the depth of interest of 20°C/km, giving a first guess at the erosion rate of 0.05 km/MY. The cooling rates predicted by the model with this erosion rate were found to be consistent with those determined through isotopic age dating and lead to muscovite and biotite ages of 260 Ma and 240 Ma respectively.

Note that no care has been taken to ensure that the original crustal thickness is obtained after erosion. Some 25 km of material are deposited onto 30 km of crust but not all of it is removed. This is done with the belief that 5-10 km of lower crustal material are melted and transported upward as granitic

magmas in response to the crustal thickening (De Yoreo et al., 1989; England and Thompson, 1986). Also note that the initial deposition of the sediments onto stable, continental crust contradicts the principle of isostasy. It is more likely that the basement for these sediments was a subsiding basin. However, the region today is underlain by normal, continental, sialic basement (Stewart et al., 1986; Stewart et al., 1985; Taylor and Toksoz, 1982) so it is clear that in some way this material was thickened and emplaced onto normal, continental crust. A number of different models have been tried, and in each case it is this deformational event which controls the subsequent thermal evolution. As long as the thickness of the pile and the distribution of the volumetric heat production is unchanged, the thermal histories are nearly identical within 10 million years after thickening. In short, the deformational event erases any memory of the period of sedimentation.

Equation 1 was solved numerically for the above parameters using a Crank-Nicholson implicit, finite difference method with grid spacing of 1 km and 1 MY. The effect of the intrusion of the Sebago batholith was then calculated as described above. The geotherms for 400 Ma, 395 Ma, 370 Ma, and 325 Ma as well as the pressure-temperature-time (P-T-t) paths for crustal horizons initially buried to 10, 20, 30, 40, and 50 km are shown in Figure 8. At 325 Ma, the ambient temperature at 14 km is about 370°C, and the cooling rates at shallower levels during the cooling phase are nearly 1°C/MY at the depths of interest. Thus the geochronologic constraints are satisfied.

In the absence of igneous intrusions, the resultant metamorphic grade for the crustal horizon passing through 14 km at 325 Ma is greenschist facies. Figure 9 shows the predicted effect of the intrusion of a 2 km thick magmatic sheet. The pluses and diamonds give the geotherms at 400 Ma and 325 Ma respectively and the triangles give the array of maximum temperatures (or metamorphic field gradient) obtained following the emplacement and cooling of the magma, which takes 40,000 years to solidify. The cross-hatched band shows the observed metamorphic field gradient from the 500°C paleoisotherm to the K-feldspar + sillimanite isograd, assuming a dip of about 3 degrees and the temperatures listed in Table 1. The agreement is excellent. The K-feldspar + sillimanite zone, which roughly coincides with the mixed zone and is highly migmatized, is predicted to have temperatures in excess of 660°C. According to the melting relations for pelitic rocks of Thompson and Algor (1977), abundant migmatization in this zone would have been likely.

It is conceivable that similar arrays of maximum temperatures at the depth range of interest could be obtained with some scheme of burial, erosion, thermal parameters, and crustal heat production. However, it would result in melting the entire crust below about 15 km, which has clearly not happened. In addition, the direction of the P-T-t paths would be counter-clockwise as in Figure 8. They would cross the andalusite-sillimanite reaction curve in a retrograde fashion. As noted previously, this is contradictory to observation; this reaction curve was crossed in a prograde direction. As can be seen in Figure 10, the P-T-t paths caused by an igneous intrusion will cross all reaction curves in a prograde fashion and do so nearly isobarically as is observed both to the north and to the south of the Sebago batholith.

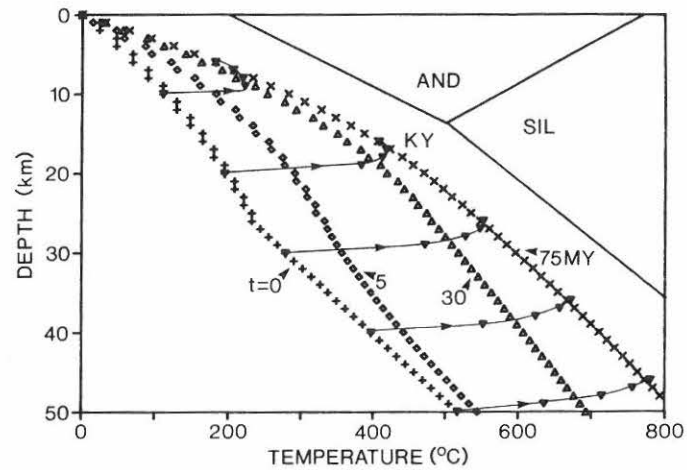

Figure 8. Thermal history for southwestern Maine based on burial and erosion history and thermal properties described in text. Pluses -- t = 0 (400 Ma); diamonds -- t = 5 MY (395 Ma); open triangles -- t = 30 MY (370 Ma); and crosses -- 75 MY (325 Ma). Closed triangles with solid lines -- P-T-t paths for horizons buried to 10, 29, 39, 40, and 50 km at t = 0; each triangle represents 20 MY. AND, SIL and KY -- andalusite, sillimanite, and kyanite. Alumino-silicate triple point after Holdaway (1971).

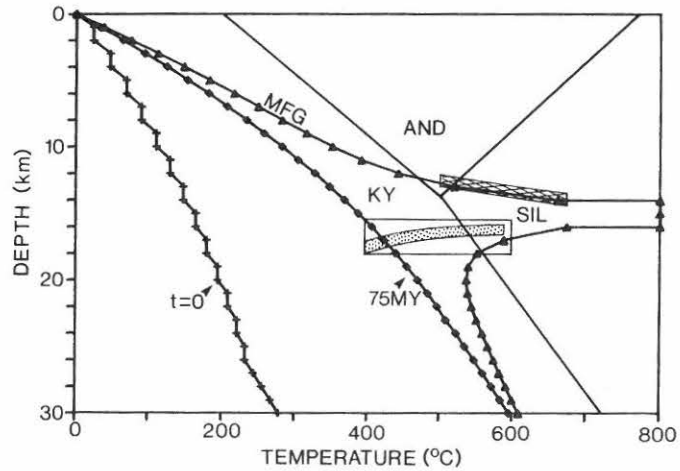

Figure 9. Effect of a 2 km thick igneous intrusion emplaced below 14 km on the temperature profile at 75 MY (325 Ma) of Figure 8. Pluses -- t = 0 (400 Ma); diamonds -- 75 MY (325 Ma); triangles -- array of maximum temperatures or metamorphic field gradient (MFG) experienced during solidification and cooling of the pluton. Cross-hatched band -- observed conditions of metamorphism south of the Sebago batholith assuming $P_{H2O} < P_{Total}$. Stippled band -- approximate MFG south of the Sebago batholith (see text). AND, SIL, and KY -- andalusite, sillimanite and kyanite. Alumino-silicate triple point after Holdaway. (1971).

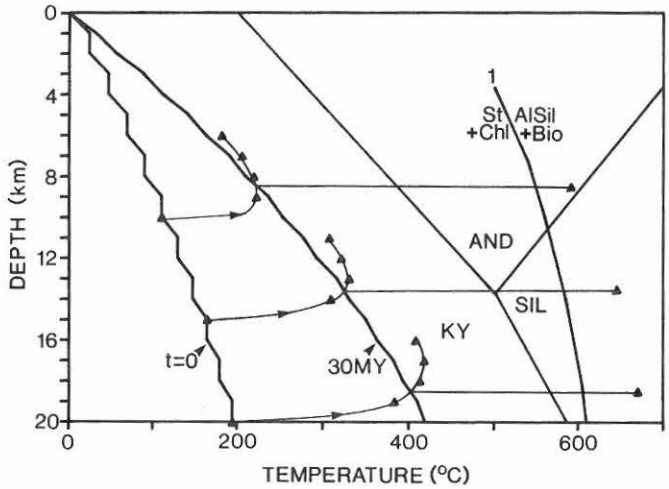

Figure 10. Effect of igneous intrusions at 30 MY on the P-T-t paths in Figure 8 for horizons buried to 10, 15 and 20 km at t = 0. Each horizon is placed in contact with the upper boundary of a 2 km thick magmatic sheet at 30 MY. (1) St = Chl + Al-Sil + Bio from Guidotti (1974).

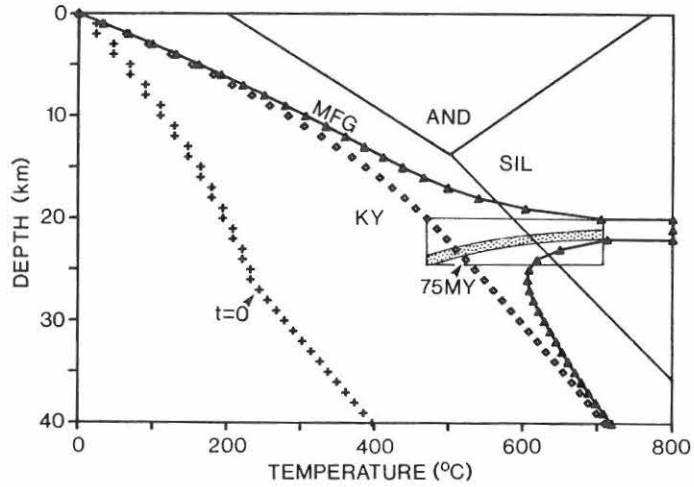

Figure 11. Effect of 2 km thick igneous intrusion emplaced below 20 km on the temperature profile at 75 MY (325 Ma) of Figure 8. Pluses -- t = 0 (400 Ma); diamonds -- t = 75 MY (325 Ma); triangles -- array of maximum temperatures or metamorphic field gradient (MFG) experienced during solidification and cooling of the pluton. Large open box -- observed conditions of metamorphism south of the Sebago batholith assuming $P_{H2O} = P_{Total}$. Stippled band -- approximate MFG south of the Sebago batholith (see text). AND, SIL, KY -- andalusite, sillimanite, kyanite. Alumino-sillicate triple point after Holdaway (1971).

Application of the model to the southern end of the batholith is more difficult for two reasons: (1) The pressure is very uncertain. (2) The study area is in a cove of the Sebago batholith where both the perimeter of the batholith and the isograds run parallel to to the dip direction. If a 3 degree dip is maintained, then the pressures of metamorphism would have been about 5.5 kbar, indicating $P_{H2O} < P_{Total}$. If we assume $P_{H2O} = P_{Total}$, then pressures were about 6.5 kbar, indicating a sharp increase in topographic relief from north to south at the time of emplacement. A traverse perpendicular to the isograds is perpendicular to the dip direction; thus pressure would have been nearly constant and it is then necessary to take into account the effect of being near the edge of the magmatic sheet. Unfortunately, the relationship of the isograds to the boundary of the igneous body is complicated by their location in the cove. With these complications in mind, the results of a straightforward application of the model to the southern end are presented in Figures 9 and 11 for the two pressure ranges discussed above. The large open box shows the range of conditions and the stippled band approximates the inferred metamorhpic field gradient (Thomson, 1986; Thomson and Guidotti, this volume). The sharp decrease in temperature with increasing depth below the pluton indicates that: (1) the Sebago batholith was the heat source for the metamorphism; (2) edge effects become important in this area.

Although the thermal effect of the pluton is large, it falls off rapidly with distance. Temperatures in excess of 500°C are experienced only out to approximately 1.5 km above the upper surface. For a dip of 2-3 degrees this is equivalent to a distance along the erosion surface of 30-50 km, in good agreement with the geochronologic data discussed above. These results demonstrate that, with only normal mantle heat flow and a lower limit of crustal radiogenic heat production, the Sebago batholith itself is the likely heat source for the surrounding regional metamorphism provided it extends to the northeast 30 to 50 km beyond its surface contact.

An upper-limit calculation was also performed. The volumetric heat production of the sediments was held at 2.25 uW/m^3 while that of the basement was increased to the value of present-day New England. The results are nearly identical. Ambient temperature at 14 km increases to 440°C at 325 Ma. Maximum temperatures within 2 km of the pluton increase by 30-45°C, and the slope of the predicted metamorphic field gradient is unchanged.

The effect of varying the melt temperature, T_m , was also examined. Variations in T_m of ± 50°C result in changes in the maximum temperatures near the pluton of <30°C.

DISCUSSION

The model presented here should be considered a lower estimate of the effects of the Sebago batholith for a number of reasons, aside from the modest values chosen for the thermal parameters. We have assumed that the magma was emplaced in a single episode, that the magma has a melting point rather than a melting range, and that the thermal diffusivity of the melt equals that of the solid. If the magma was injected in pulses on a 0.01 - 1.0 MY time scale, if the magma solidified over a range in temperature, or if the decreased thermal diffusivity of the melt was considered, the solidification and cooling time would have

been prolonged and the thermal effects would have been more extensive.

The results of the model require the batholith to extend 30-50 km beyond its northern contact. The zone of pegmatites which wraps around the northeastern side of the batholith (Fig. 2) may provide evidence for this northeastern extension. We propose that these pegmatites are apophyses of the Sebago batholith, extending upward from the underlying sheet. The most northern of these pegmatites, the Whitecap Mountain pegmatite, (Fig. 2), is about 35 km north of the batholith, and there is evidence that it is Carboniferous in age rather than Devonian (Guidotti et al., 1986). If this northerly extension does not exist, the model may still apply if the Sebago batholith is only one of a number of related and possibly interconnected sill-like granite bodies intruded into the region at about 325 Ma ago.

Even if the above arguments are correct, the problem of a primary heat source still remains. However, as is shown in Figure 8, even with the lower-limit calculation, temperatures in excess of 800°C are experienced at the bottom of the crust by 325 Ma. In the upper limit, temperatures above 800°C are experienced everywhere below 40 km depth by 325 Ma. Thus it is likely that granitoid magmas would have been generated in response to Acadian thickening. Chamberlain and England (1985) reached a similar conclusion in examining the thermal evolution of the Kearsarge-central Maine synclinorium in New Hampshire.

As discussed above, Aleinikoff et al. (1985), obtained U-Pb ages on both zircons and monazites from the Sebago batholith. They found that the U-Pb age of the monazite was 50 million years younger than that of the zircon and suggested that the Sebago batholith, intruded at 325 Ma, did not cool below 600°C until 275 Ma. Furthermore, they suggest that cooling to 600°C was due to uplift and that, therefore, ambient temperature was 600°C. If the depth of emplacement of the Sebago batholith was 14 km, this would lead to an average gradient of over 40°C/km and pervasive melting of the crust below 15 km depth. In addition, for any reasonable radioactive productivity for the crust, the mantle heat flux would have had to have been 2-3 times that of present-day New England.

Aleinikoff et al. (1985) also suggest that the 10 million year difference in monazite ages between the two samples studied had two possible causes: (1) Because it was near the edge of the pluton, the sample near the northeast margin of the batholith cooled more rapidly than the sample near the center of the batholith. (2) The northeast portion of the batholith was uplifted more rapidly, causing a regional cooling gradient. The first explanation is unlikely to be correct because the batholith is so thin (≤2 km) that both samples would have cooled primarily from top and bottom and not from the edges. (Both sample localities are far more than 2 km from the edge.) Furthermore, for a reasonable ambient temperature (≤400°C), the time required to cool below 600°C is on the order of 10,000-100,000 years and not 10 million years. The second explanation is also unlikely because, once again, it required an ambient temperature of >600°C at 14 km. One can only conclude that either: (1) The closure temperature for Pb loss in monazite is hundreds of degrees lower than assumed. (2) The actual monazite ages are tens of millions of years older than the calculated ages. (3) The monazite ages represent a reheating event for the Sebago batholith. While there is no petrologic evidence for the latter, it should be noted that several $^{40}Ar/^{39}Ar$ release spectra for hornblendes from the region have diffusion gradients whose minimum ages are in the range of 270-290 Ma (Lux and Guidotti, 1985; Lux, unpub. data). These could also be interpreted as a result of a Permian heating event. However, these diffusion gradients are only observed in a few samples from the region. This suggests that if they are due to reheating, they may well be the result of local heat sources.

CONCLUSION

We have presented geochronologic data which provide an estimate of the ambient temperature at the time of the intrusion of the Sebago batholith, give a picture of the progression of cooling throughout the Sebago from 245 Ma to 230 Ma, and delineate the northern limit of the Carboniferous regional metamorphic event (M_C) in western Maine.

We have also presented a thermal model for the sheet-like Sebago batholith which suggests that M_C is a deep-level contact aureole of the batholith itself and that the pluton is likely to be a result of lower crustal melting in response to Acadian thickening. We hypothesize that the Sebago batholith, or closely related granitic sheets, extend at least 30 to 50 km to the northeast and that the metamorphism to the north of the Sebago batholith was produced above the batholith while that to the south was produced below the batholith.

ACKNOWLEDGMENTS

This work was supported by EPSCOR grant ISP-8011448. We thank C. P. Chamberlain and M. J. Holdaway for their careful and thoughtful reviews.

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Maine Geological Survey
Studies in Maine Geology: Volume 3
1989

Carboniferous Barrovian Metamorphism in Southern Maine

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ABSTRACT

Field and petrologic observations in the Windham-Gorham area of southern Maine indicate that a part of the area contains kyanite as the aluminosilicate polymorph. This occurrence is indicative of a medium-pressure (Barrovian) metamorphism and is in sharp contrast to surrounding areas in Maine and New Hampshire which are typically characterized by low-pressure metamorphism (andalusite and/or sillimanite instead of kyanite). The grade of metamorphism ranges from garnet to upper sillimanite zone, and the present pattern of isograds is a result of a static (post-tectonic) Barrovian event. In the context of available petrogenetic grids, the mineral parageneses and isogradic reactions lead to an estimate for the minimum and maximum pressures of metamorphism at 6.25 and 7.5 kbars (20-25 km), respectively. Temperatures estimated from the petrogenetic grid approach range from 500° - 700°C. Geothermometry and geobarometry based on exchange reactions provide temperature and pressure estimates for the kyanite zone which range from 490° - 540°C and 5.5-6.4 kbars (18-22 km) (assuming an ideal solution model), respectively.

The sequence of events postulated for the Windham-Gorham area is as follows: (1) Acadian upright, tight, isoclinal folding which controls the present map pattern (D_1), (2) a widespread (Acadian?) metamorphic event (M_1) to staurolite grade, (3) localized deformation (D_2) as a direct result of the intrusion of the Sebago batholith 325 million years ago, (4) a static, post-tectonic Barrovian-type metamorphism (M_2) which is spatially related to the underside of the Sebago batholith and which caused both prograde and retrograde metamorphism of the M_1 assemblages, and (5) a late, sporadic retrograde metamorphic event (M_3) near the batholith. It appears that the Windham-Gorham area is the surface expression of a deeper level of the crust, preserving rocks of a Carboniferous metamorphic event.

INTRODUCTION

The higher grade, regionally metamorphosed rocks of pelitic bulk compositions in southern Maine typically contain andalusite and/or sillimanite (Thompson and Norton, 1968). Hence, according to Miyashiro (1973), they would be classified as a low-pressure baric type. As expected, other common pelitic minerals include biotite, muscovite, garnet, and staurolite. However, there is a relatively small area near the town of South Windham containing kyanite as the Al_2SiO_5 polymorph. This would be more characteristic of the medium-pressure baric type of Miyashiro (1973). Although collected for years by mineral enthusiasts (Cook Road in Windham has yielded kyanite since the late 1800's), the kyanite has never before been investigated

in a petrologic sense. The occurrence of kyanite and its Barrovian implication are quite significant in the interpretation of the geologic and metamorphic history of southern Maine.

The purpose of this investigation is to ascertain the metamorphic history of the Windham-Gorham area and provide a tectonic interpretation, especially in the context of the kyanite occurrence and the recently determined Carboniferous age of the Sebago batholith (Hayward and Gaudette, 1984; Aleinikoff, 1984; Aleinikoff et al., 1985).

The area studied includes approximately 40 square miles and is located near the towns of Gorham and South Windham, about 15 miles northwest of Portland. In a regional sense it is located on the eastern limb of the Kearsarge-central Maine synclinorium. It lies just south of and includes a portion of the Sebago batholith, the largest pluton in Maine (Fig. 1).

GEOLOGIC SETTING

Previous Geologic Work

Previous geologic work in southern Maine includes studies by Katz (1917), Perkins and Smith (1925), Fisher (1941), and Osberg (1968). Much of the more recent geologic mapping in southern Maine has been carried out by Hussey (1968, 1971, 1981, 1985). These works have provided the geologic framework for this study. Following Billings (1956), Hussey (1968) placed the rock units of the present study area (then called the Berwick and Eliot Formations) within the Merrimack Group.

Hussey (1971), in his description of the structure, stratigraphy, and lithology of the units comprising the Portland quadrangle (including the Windham-Gorham area), introduced the name Windham Formation and correlated it with the Eliot Formation of Katz (1917) and the Waterville Formation of central Maine. The Berwick Formation was interpreted to lie conformably above the Windham Formation and correlated with the Vassalboro Formation to the north (as indicated by Osberg, 1968). Both units were considered to be members of the Merrimack Group (Hussey, 1971).

Hussey's (1981) work in the Lower Androscoggin-Casco Bay area also included the present study area. In that study the units were not designated as members of the Merrimack Group due to uncertainties in radiometric dating (Lyons et al., 1982). However, Hussey (1981) still preferred to correlate the Berwick with the Vassalboro Formation. He indicated the Vassalboro Formation to be Late Ordovician(?) to Early Silurian(?) and to conformably underlie the Windham Formation of Early Silurian(?) age. This change in stratigraphic interpretation was in response to the studies of Osberg (1980) in south-central Maine. The recent Bedrock Geologic Map of Maine (Osberg et al., 1985) follows this stratigraphic interpretation.

Stratigraphy and Lithology

A major part of the study area consists of metamorphosed sedimentary strata which are now calc-silicate and biotite granofels, marbles, and mica schists. Igneous rocks include granite (Sebago batholith) and minor mafic intrusive rocks too small to be shown at the scale of mapping. The Windham-Gorham area includes northeast-trending metamorphosed strata of Late Ordovician(?) to Early Silurian(?) age, specifically the Vassalboro and Windham Formations (Fig. 2). The name Vassalboro is used herein for that unit which in the study area was originally designated as the Berwick Formation (Katz, 1917; Hussey, 1968, 1971). This has been done in order to avoid confusion with the recent reinterpretation of the Berwick Formation in southeastern New Hampshire as a Late Precambrian formation (Bothner et al., 1984).

Stratified Rocks. The Vassalboro Formation consists of a thick sequence of quartz + plagioclase + biotite ± hornblende granofels (75-80%) with lesser amounts (10-15%) of calc-silcate granofels and micaceous schist.

The Windham Formation consists of three lithologies: (1) a pelitic schist, (2) a ribbon limestone, and (3) calc-silicate and biotite granofels. The principal lithology is a thin-bedded, two-mica + garnet + quartz + plagioclase schist with staurolite present as metamorphic grade increases, and abundant sillimanite near the Sebago batholith. Locally, the high-pressure aluminosilicate polymorph kyanite is present. For a detailed description of lithologies refer to Thomson (1985) and Hussey et al. (1986).

Intrusive Rocks. The Sebago batholith, mainly a peraluminous, two-mica granite, lies partly within the study area (Fig. 2). Gravity data suggest that the batholith is a rather thin (1 km±), sub-horizontal sheet (Kane and Bromery, 1968; Hodge et al., 1982). Previously thought to have been part of the New Hampshire Plutonic Series, the Sebago batholith has been dated recently as Carboniferous (325 Ma) by Hayward and Gaudette (1984), Aleinikoff (1984), and Aleinikoff et al. (1985).

In addition to the Sebago batholith, there are numerous basaltic and diabasic dikes of variable thickness (both parallel to and crosscutting the structural grain of the country rock). These dikes have not been affected by any subsequent metamorphic activity and are thought to be related to Triassic intrusive activity (Hussey, 1971).

Structure

The work of Hussey (1971, 1981, 1985) has provided most of the geologic framework for this study with only a few minor changes. The general structure of the area consists of a series of synforms and antiforms that strike approximately N30°E and have gentle plunges to the north and south (Fig. 2). A generalized cross-section (A-A') through the north-northeast portion of the area is shown as well.

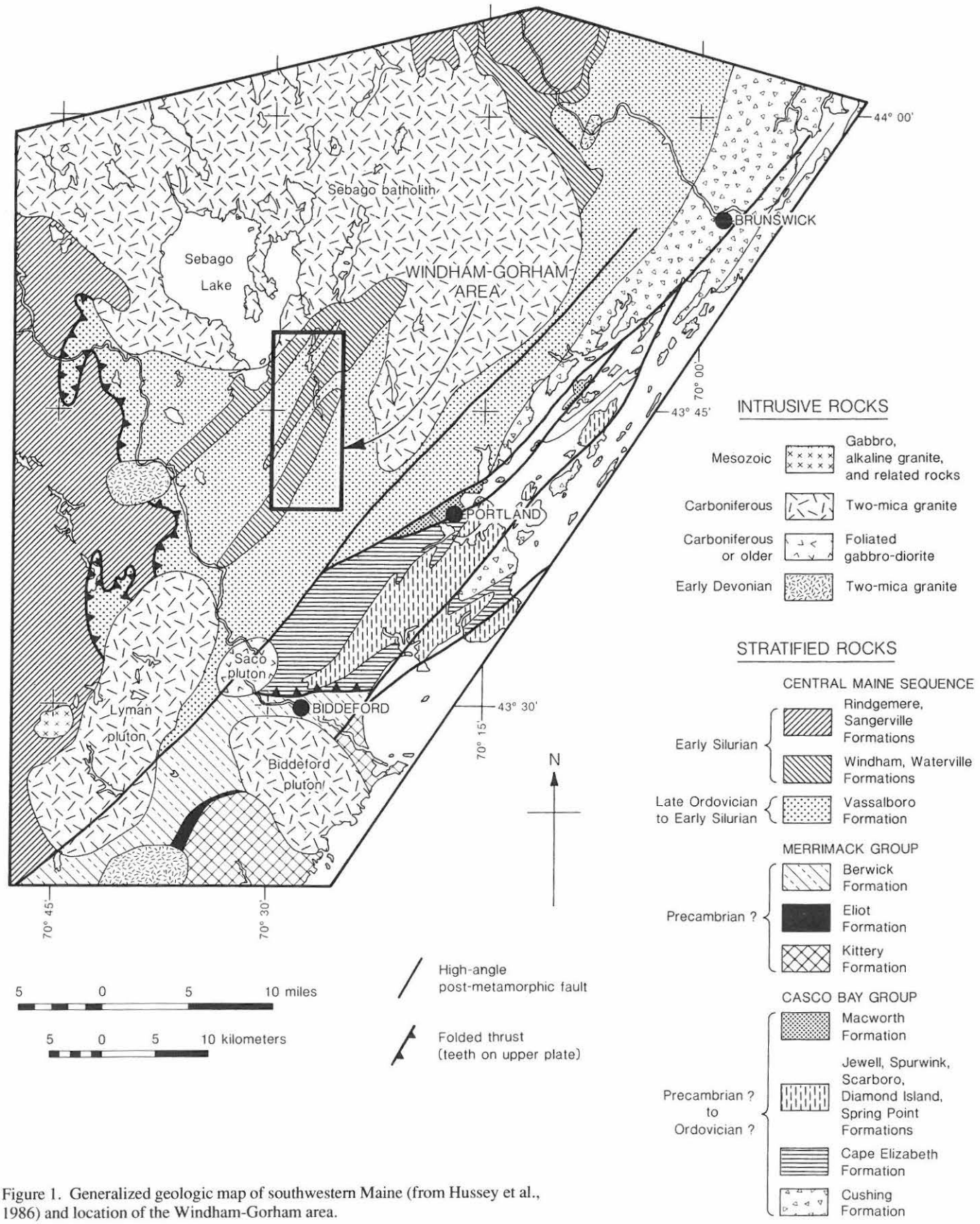


Figure 1. Generalized geologic map of southwestern Maine (from Hussey et al., 1986) and location of the Windham-Gorham area.

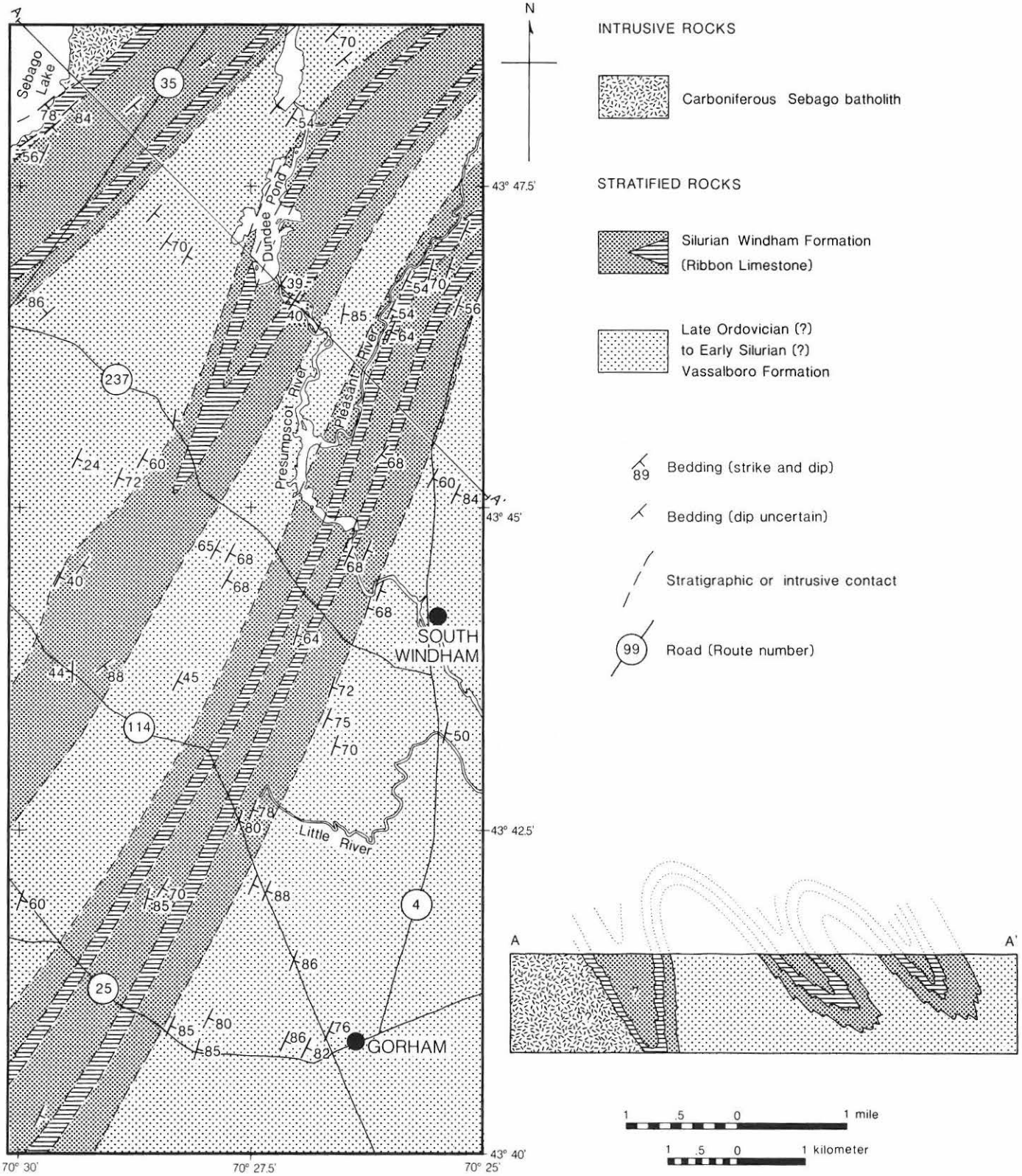


Figure 2. Generalized geologic map and cross section of the Windham-Gorham area (modified from Hussey, 1985).

The rocks of the study area are multiply deformed. The area was previously interpreted by Hussey (1981) to have been deformed by an early event that produced large-scale recumbent folds (F_1), and then a later event that caused large-scale upright to slightly overturned folds (F_2). However, Hussey's (1985) most recent interpretations suggest that the study area shows evidence only of the later (F_2) deformation. Both F_1 and F_2 folds are a result of the Acadian orogeny of Early Devonian age (Hussey et al., 1986).

Minor structures (crenulation cleavages and tight isoclinal folds), and hence the degree of deformation, increase as the Sebago batholith is approached. These F_3 minor structures are directly related to the intrusion of the Sebago batholith during the Carboniferous and so are Carboniferous in age (Thomson, 1985; Hussey et al., 1986).

REGIONAL METAMORPHIC SETTING

Many detailed studies of the metamorphic rocks have been carried out in parts of Maine near the study area. Here, discussion is directed only at those most relevant to the present study. Guidotti (1970) presented a discussion of the metamorphism near Rangeley, north of the study area. The metapelites in that region range in grade from biotite to K-feldspar + sillimanite zone and show evidence of polymetamorphism. Particular emphasis has been placed on part of Guidotti's (1970) Figure 1, Area B. There the rocks range from garnet to upper sillimanite zone. Guidotti (1970) suggested three Acadian events for this area. M_1 was a low-grade event followed by M_2, an andalusite + staurolite + biotite (+ cordierite locally) event. A subsequent metamorphic event (M_3) reached sillimanite grade near plutons, thereby demonstrating a spatial relationship between metamorphism and plutonism. M_3 resulted in a "hinge-effect" (Guidotti, 1970), as it caused prograde metamorphism of andalusite-staurolite grade rocks near the plutons and a retrogradation of andalusite-staurolite grade rocks farther away. Depending upon position in the M_3 metamorphic gradient, the andalusite and staurolite (and cordierite) produced by the M_2 event were partially to completely replaced by pseudomorphs.

Guidotti (1970) also discussed the metamorphism in an area immediately north of the Sebago batholith (see Fig. 1 of Guidotti, 1970: Area D). Some of the rocks there are migmatites that are systematically related to the K-feldspar + sillimanite isograd and probably formed by partial melting (Guidotti, pers. commun., 1985). This is in marked contrast to the present study area where the rocks only reach the upper sillimanite zone and show no evidence of migmatization.

A fairly similar metamorphic history was suggested by Novak and Holdaway (1981) for the area just west of Augusta. The M_2 and M_3 events of Novak and Holdaway (1981) may be similar to Guidotti's (1970) M_3 event. All of these sillimanite-producing events are spatially related to plutons.

Holdaway et al. (1982) recognized four Acadian events (M_1-M_4) in west-central Maine. There the M_2 assemblages (andalusite-staurolite, locally cordierite) are affected by a later, somewhat higher pressure metamorphic event (M_3) which involved a range of grades. The M_3 regional metamorphism is spatially related to plutons throughout most of the area. Finally, M_4 was a local contact event which produced staurolite, andalusite, and cordierite around some of the Devonian plutons in the region.

Hussey (1981) stated that the rocks in the Windham-Gorham area were metamorphosed by a low-pressure, intermediate (Buchan type) event. He also noted, however, that the occurrence of kyanite may indicate the higher pressure Barrovian-type metamorphism.

All early studies considered the metamorphism to be of Acadian age but more recently Lux and Guidotti (1985), DeYoreo et al. (1985), and Guidotti et al. (1986a, 1986b) have presented evidence for a Carboniferous thermal event in western Maine, related to the Carboniferous Sebago batholith. A combination of petrologic evidence and disturbed K-Ar ages for Devonian plutons (Songo and Mooselookmeguntic plutons) in the area studied by Guidotti (1970) led to the suggestion that the high-grade K-feldspar + sillimanite isograd of western Maine is of Carboniferous (Holdaway et al., 1982; Guidotti et al., 1983) rather than Devonian age as previously claimed. The regional metamorphic history of Maine may be more complex than previously thought and may involve widespread Carboniferous regional or contact metamorphism superimposed on an Acadian metamorphic terrane. As developed later, this has implications for the metamorphism of the stratified rocks in the Windham-Gorham area due to their close proximity to the Sebago batholith.

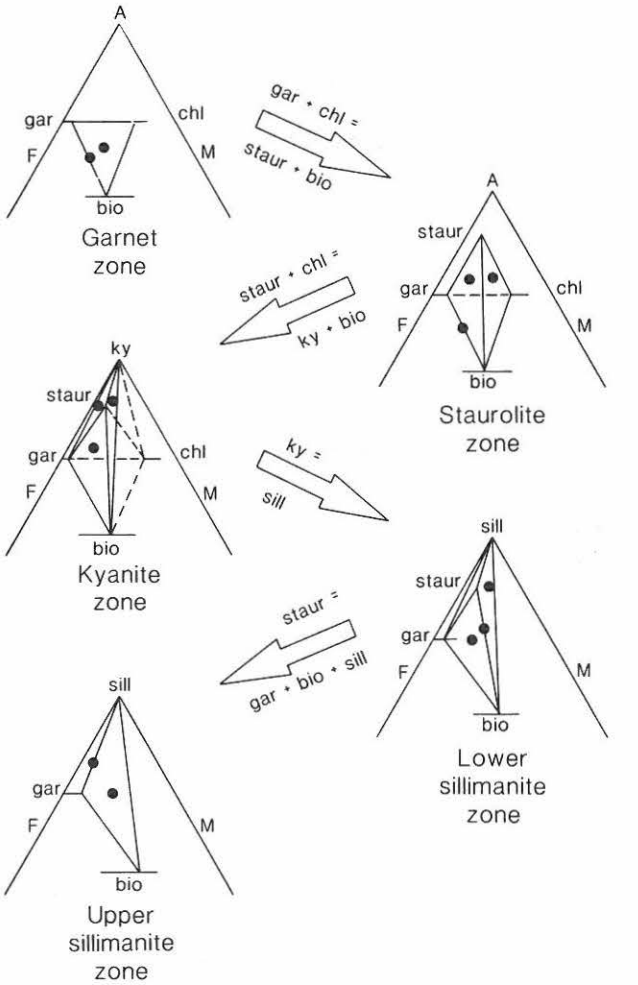
PETROLOGY OF THE METAMORPHIC ROCKS

Petrography

A total of 250 stations were recorded from which 101 pelitic schist samples were collected for detailed petrographic and petrologic research. Equilibrium assemblages of pelitic schists have been plotted on Thompson's (1957) AFM diagram (Fig. 3). The grade of metamorphism increases from the garnet zone in the southeast to the upper sillimanite zone in the northwest as the Sebago batholith is approached (Fig. 4). Important textural characteristics are discussed below and shown schematically in Figure 5a-l. For a more detailed petrographic discussion, see Thomson (1985).

Garnet Zone. The lowest grade mineral assemblages observed belong to the garnet zone. Rocks in this zone were previously interpreted to belong to the staurolite zone due to a reported occurrence of staurolite in a thin pelitic layer of the Vassalboro Formation by Hussey (1971; pers. commun., 1985). Based upon the AFM projections of the observed assemblages (Fig. 3), it appears that garnet zone is a better designation.

Biotite and garnet within this zone commonly exhibit a helicitic overprint texture with inclusion trails continuous with the external schistosity (Fig. 5a,b). Less commonly, the garnets



Garnet zone
bio + gar + Mg-chl
bio + gar

Staurolite zone
bio + gar + staur + Mg-chl
bio + gar + staur
bio + staur + Mg-chl
bio + gar

Kyanite zone
bio + gar + staur + ky
bio + gar + staur
bio + staur + ky
bio + staur + Mg-chl
bio + gar

Lower sillimanite zone
bio + gar + staur + sill
bio + gar + staur
bio + staur

Upper sillimanite zone
bio + gar + sill
gar + sill

+ white mica, quartz, and plagioclase
± ilmenite, graphite, and pyrrhotite

Figure 3. Schematic AFM topologies and observed mineral assemblages for the Windham-Gorham area (dashed lines represent those assemblages stable with respect to a component other than those taken into account by the AFM projection, e.g. CaO, MnO).

appear as anhedral, elongated grains parallel to the foliation. In many samples euhedral garnet porphyroblasts, as well as Mg-rich chlorite, are arranged randomly within the groundmass and the external schistosity is truncated at the boundaries of the porphyroblasts (Fig. 5f,e). Finally, in a few cases, garnets have been observed which contain an internal schistosity that is inconsistent with the external foliation (Fig. 5h).

Staurolite Zone. The assemblages present in the staurolite zone are listed in Figure 3 and have been used to establish the mineral compatibility diagram (topology) shown. The four phase assemblage bio + gar + staur + Mg-chl seems to be inconsistent with Thompson's (1957) mineral facies concept (i.e., the number of phases exceeds the number of components). However, it has been suggested in such cases (Guidotti, 1970; Evans and Guidotti, 1966) that garnet may be stabilized with respect to a component other than those taken account of on the AFM projection (e.g., MnO or CaO). The analyses discussed in the next section show that these components are present in garnet. Thus they are believed to stabilize garnet so that it is present in the bio + staur + Mg-chl field of the AFM diagram.

Biotite, Mg-chlorite, and garnets of the staurolite zone exhibit the same textural characteristics as described for the garnet zone. Large staurolite poikiloblasts are dominated by the helicitic overprint texture (Fig. 5c) and commonly contain euhedral garnets as inclusions.

Kyanite Zone. The kyanite zone is characterized by the key mineral assemblage quartz + plagioclase + biotite + muscovite ± garnet ± staurolite ± kyanite (and ± some Mg-rich chlorite). A complete listing of assemblages present in the kyanite zone and the resulting mineral facies diagram are presented in Figure 3.

Garnet is either inclusion-free or poikiloblastic. Helicitic overprint texture is observed in many specimens (Fig. 5b). Staurolite is present as large poikiloblasts exhibiting helicitic overprint texture. However, some samples contain staurolites which have an internal schistosity which is inconsistent with the surrounding foliation (Fig. 5i). Garnet is commonly found as euhedral inclusions in staurolite.

Kyanite, although commonly associated with quartz pods, is also found as euhedral porphyroblasts within the groundmass. The porphyroblasts truncate the foliation at their grain boundaries without any deflection (Fig. 5g). Occasionally, inclusion trails of ilmenite, continuous with the external schistosity, pass straight through a kyanite porphyroblast (Fig. 5d).

Lower Sillimanite Zone. The aluminosilicate present in this zone is sillimanite instead of kyanite. Kyanite has been observed in some outcrops but it is generally associated with quartz pods and veins. The assemblages present are listed and shown graphically in Figure 3.

Chlorite is invariably an Fe-rich variety and is believed to be present only as an alteration of biotite. It is found closely associated with biotite and as larger laths within quartz veins. It is probably due to late hydrothermal activity.

Garnet is present within the groundmass and also within coarse-grained muscovite pseudomorphs after staurolite. Presumably the latter were once inclusions within the staurolite grains that are now replaced by pseudomorphs.

Staurolite usually occurs as grains exhibiting helicitic overprint texture (Fig. 5c). However, in some samples the staurolite has an internal schistosity, marked by a pattern of inclusions,

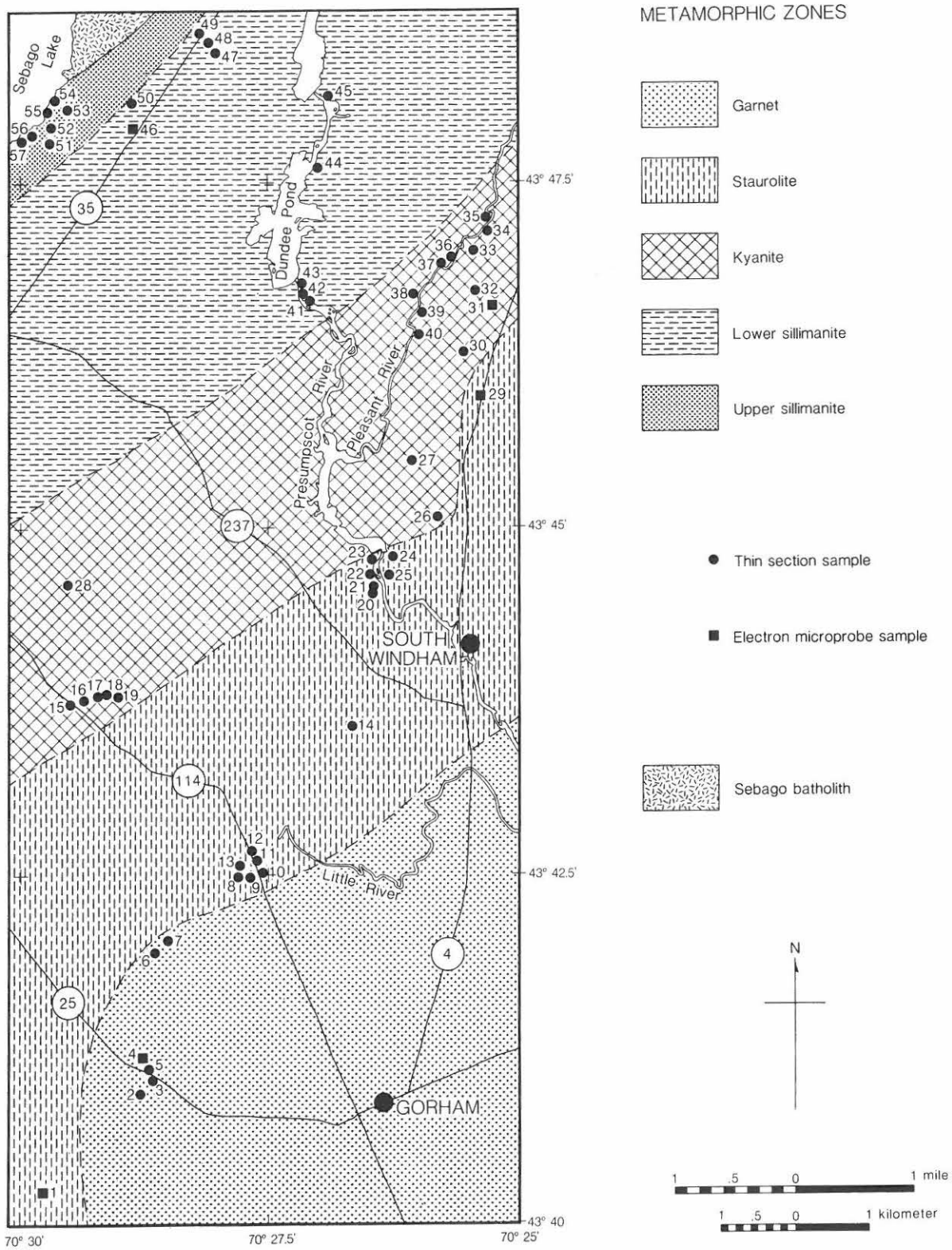


Figure 4. Sample locality and metamorphic map of the Windham-Gorham area (only samples from the pelitic lithology shown).

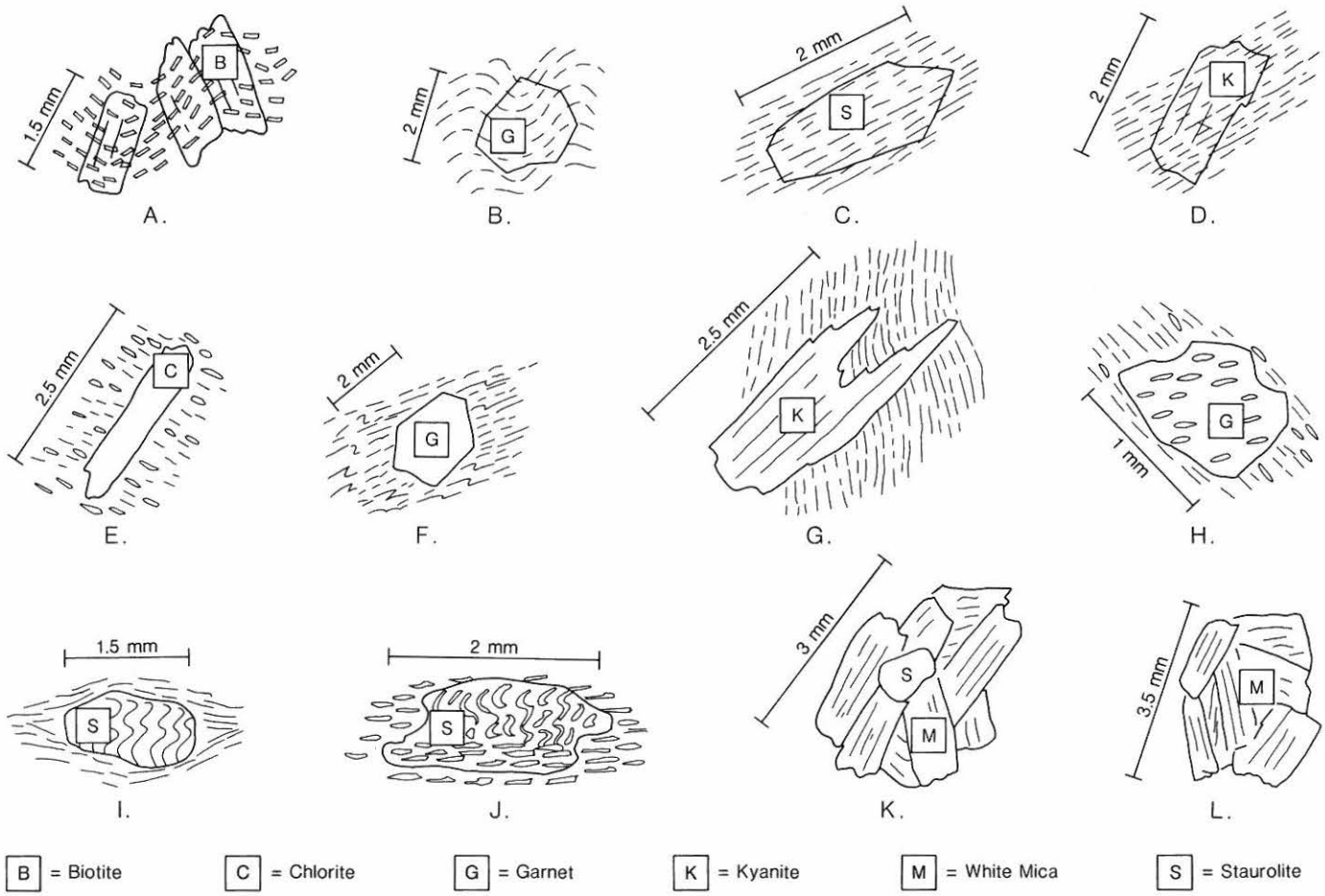


Figure 5. Mineral textures (see text for discussion). (a-d) Post-tectonic helicitic overprint texture in biotite, garnet, staurolite, and kyanite. (e-g) Post-tectonic chlorite, garnet, and kyanite. (h-i) Pre-tectonic garnet and staurolite. (j) Pre- and post-tectonic staurolite. (k-l) Partial and complete pseudomorphs after staurolite.

which is folded in a manner inconsistent with the external foliation within the thin section (Fig. 5i). Occasionally, grains with this latter pattern are rimmed with optically continuous poikiloblastic staurolite with an inclusion pattern which is continuous with the external foliation (Fig. 5j). Staurolite also occurs as relict cores within coarse-grained muscovite pseudomorphs (Fig. 5k).

Sillimanite first appears in the lower sillimanite zone as fibrolitic masses occasionally closely associated with biotite. The fibrolitic masses are oriented parallel to the foliation.

Upper Sillimanite Zone. This zone is characterized by the complete disappearance of staurolite, which is commonly replaced by pseudomorphs of coarse-grained muscovite. The most common assemblage is quartz + plagioclase + muscovite ± garnet ± sillimanite. Other assemblages and the AFM topology are presented in Figure 3. Chlorite, where present, is clearly a retrograde Fe-rich chlorite and so is not considered in the AFM topology.

Garnet occurs as isolated anhedral to subhedral grains lacking any significant inclusion patterns. Sillimanite, as fibrolitic aggregates, occurs oriented parallel to the foliation throughout the groundmass. Such aggregates are commonly intergrown with biotite.

Electron Microprobe/Chemical Analyses

Minerals in six polished thin sections (one or two sections from each metamorphic zone with the exception of the upper sillimanite zone, Fig. 4) were analyzed with the MAC 400S three spectrometer electron microprobe at the University of Maine at Orono. A 15 kV potential was used with a .02 microampere beam current and a 2-3 micron beam size. Each analysis was conducted for 10 seconds or 10,000 counts. Garnet, biotite, plagioclase, and staurolite were analyzed using silicate standards for most elements. The analytical data were reduced using the Bence and Albee (1968) correction procedure. The results of these analyses are presented in Tables 1-4.

Analytical data for biotite, plagioclase, and staurolite are average compositions from 1-3 grains in each sample.

TABLE 1. CHEMICAL ANALYSES OF BIOTITE.*

Zone	Garnet	Staurolite		Kyanite		Lo sill
Locality	4	1	29	31	31	46
SiO_2	35.74	35.65	36.34	37.30	36.96	36.29
Al_2O_3	19.45	19.91	19.77	19.58	19.69	19.57
TiO_2	1.45	1.42	1.49	1.71	1.84	1.79
FeO_t	18.73	19.35	18.69	15.75	15.74	17.85
MgO	10.02	10.16	10.37	11.21	11.65	8.77
MnO	0.16	0.04	0.15	0.25	0.32	0.23
K_2O	8.39	8.56	8.21	8.56	8.57	9.07
Na_2O	0.77	0.05	0.00	0.47	0.67	0.55
Total	94.71	95.14	95.00	94.83	95.24	94.13
		Formula based on 22 oxygens				
Si	5.423	5.452	5.532	5.549	5.461	5.529
Al^{iv}	2.577	2.518	2.468	2.451	2.539	2.471
Total	8.000	8.000	8.000	8.000	8.000	8.000
Al^{vi}	0.900	1.071	1.080	0.980	0.904	1.045
Ti	0.165	0.041	0.170	0.188	0.205	0.201
Fe_t	2.377	2.475	2.378	1.956	1.953	2.272
Mg	2.266	2.315	2.352	2.484	2.576	1.990
Mn	0.018	0.006	0.002	0.030	0.039	0.026
Total	5.726	5.908	5.982	5.638	5.677	5.534
K	1.621	1.699	1.594	1.624	1.621	1.762
Na	0.277	0.015	0.000	0.135	0.191	0.161
Total	1.898	1.819	1.594	1.759	1.812	1.923
Mg/Fe_t	0.953	0.935	0.989	1.269	1.319	0.876
$Mg/(Mg+Fe_t)$	0.488	0.483	0.497	0.559	0.569	0.467

* Average analyses of 3 to 5 biotite grains

TABLE 2. CHEMICAL ANALYSES OF PLAGIOCLASE.*

Zone	Kyanite		Lo sill
Locality	31	31	46
SiO_2	61.75	60.57	62.81
Al_2O_3	24.71	24.18	23.30
MgO	0.10	0.11	0.08
CaO	5.97	5.96	4.33
Na_2O	8.26	8.49	8.89
K_2O	0.21	0.18	0.30
Total	101.11	99.47	99.71
		Formula based on 32 oxygens	
Si	10.865	10.836	11.154
Al	5.122	5.089	4.854
Mg	0.023	0.026	0.016
Ca	1.120	1.139	0.821
Na	2.814	2.945	3.059
K	0.044	0.037	0.062
Total	19.988	20.081	19.966
Ab	70.7	71.5	77.6
An	28.2	27.6	20.8
Or	1.1	0.9	1.6

* Average of 5 analyses on 1-2 grains.

TABLE 3. CHEMICAL ANALYSES OF GARNET.*

Zone	Garnet		Staurolite				Kyanite				Lo sill	
	Rim	Core	Rim	Core	Rim	Core	Rim	Core	Rim	Core	Rim	Core
Locality	4		1		29		31		31		46	
SiO_2	37.01	37.23	37.16	36.27	37.57	37.09	37.53	37.28	37.54	37.43	36.82	37.04
Al_2O_3	22.36	22.54	22.09	21.75	21.68	20.98	21.65	21.80	21.62	21.64	22.52	22.02
FeO_t	30.75	23.62	35.69	32.20	33.24	28.14	29.19	25.09	29.48	29.92	31.14	31.43
MgO	2.17	1.22	2.49	1.64	2.43	0.97	3.39	2.04	3.02	1.96	3.19	3.36
MnO	4.12	8.18	0.79	3.95	2.76	9.43	4.74	10.86	5.28	8.25	4.88	5.62
CaO	3.73	5.34	2.65	3.57	3.72	3.14	2.76	2.21	2.88	1.90	2.19	1.96
Total	100.13	98.13	100.87	99.38	101.40	99.75	99.26	99.28	99.74	101.09	100.74	101.44
					Formula based on 12 oxygens							
Si	2.960	3.005	2.962	2.949	2.981	3.016	3.006	3.007	3.003	2.990	2.929	2.940
Al^{iv}	0.040	–	0.038	0.051	0.019	–	–	–	–	0.010	0.071	0.060
Total	3.000	3.005	3.000	3.000	3.000	3.016	3.006	3.007	3.003	3.000	3.000	3.000
Al^{vi}	2.068	2.143	2.037	2.032	2.008	2.009	2.043	2.072	2.038	2.028	2.040	2.000
Fe_t	2.057	1.594	2.382	2.188	2.207	1.914	1.955	1.693	1.972	1.998	2.071	2.085
Mg	0.258	0.144	0.295	0.198	0.287	0.117	0.402	0.244	0.359	0.232	0.377	0.396
Mn	0.277	0.559	0.051	0.272	0.186	0.649	0.321	0.741	0.355	0.558	0.328	0.377
Ca	0.319	0.461	0.225	0.310	0.314	0.272	0.235	0.189	0.238	0.162	0.186	0.165
Total	2.911	2.758	2.953	2.968	2.994	2.952	2.913	2.867	2.924	2.950	2.962	3.023

* Rims are the averages of two analyses just coreward from the retrograded rims. (See text)

TABLE 4. CHEMICAL ANALYSES OF STAUROLITE.*

Zone	Staurolite		Kyanite	
Locality	1	29	31	31
SiO_2	28.15	27.34	27.47	27.46
Al_2O_3	54.04	54.72	55.50	55.12
TiO_2	0.49	0.55	0.50	0.53
FeO_t	14.27	13.77	13.91	13.73
MgO	0.98	1.66	1.16	1.35
ZnO	0.31	0.07	0.00 **	0.00 **
Total	98.24	98.10	98.54	98.20
		Formula based on 24 oxygens		
Si	4.062	3.942	3.944	3.953
Al	9.196	9.309	9.363	9.351
Ti	0.051	0.059	0.050	0.056
Fe	1.720	1.660	1.668	1.651
Mg	0.211	0.355	0.244	0.287
Zn	0.029	0.004	0.000	0.000
Total	15.269	15.329	15.300	15.299

* Average of 5 analyses on 1-3 staurolite grains.
** Below detectibility limits due to high background and low count rates.

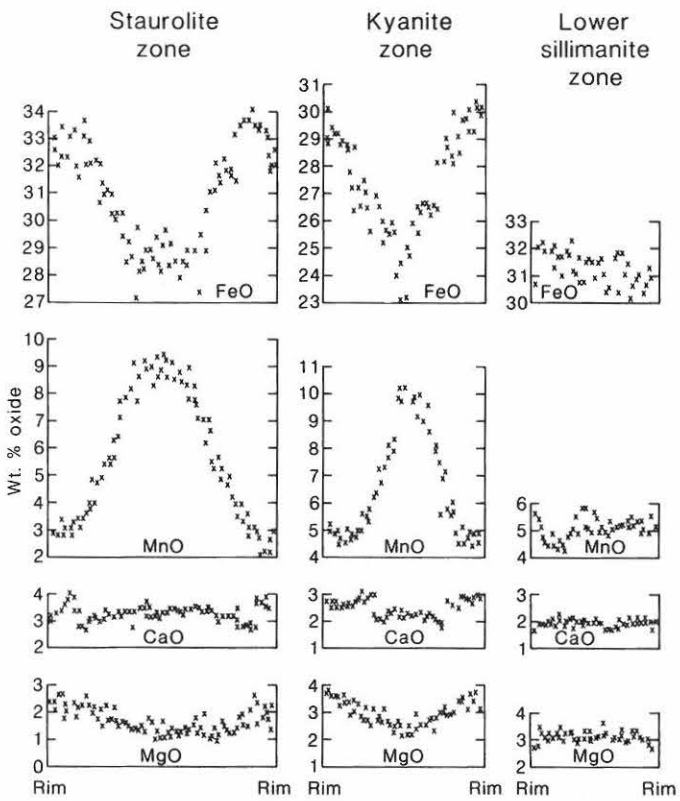


Figure 6. Garnet profiles from the staurolite, kyanite, and lower sillimanite zones showing a variation in FeO, MnO, CaO, and MgO (wt.% oxide) across a rim to core to rim traverse (at intervals of 10 to 15 microns).

Plagioclase analyses were done as close as possible to the rim of grains.

Garnets from each metamorphic zone, with the exception of the upper sillimanite zone, were analyzed along rim to core to rim traverses across grains at an interval of approximately 10-15 microns. Grains selected for analyses had near-vertical boundaries and were analyzed through the center of the grain. Electron microprobe analyses in Table 3 give average rim and average core compositions in each metamorphic zone. Figure 6 shows the resultant compositional profiles for MnO, MgO, CaO and FeO. The compositional profiles are comparable to other garnet zoning studies by Tracy et al. (1976). Because the garnet compositional profiles in the present study exhibit a retrogradational "lip" as evidenced by a slightly increased MnO content (Tracy et al., 1976), the rim compositions reported in Table 3 are actually the average of two analyses just inward from the "lip".

White mica from kyanite-bearing assemblages was analyzed by X-ray powder diffractometry to test for the occurrence of paragonite, the sodium end-member of the white mica solid solution series. An assemblage most likely to have paragonite (Guidotti, 1984) was chosen, but it was not detected.

PETROGENESIS

Polymetamorphism and Times of Crystallization

Previous studies throughout the New England Appalachians have shown much of the region to be polydeformed as well as polymetamorphosed. The present study also shows evidence of polymetamorphism as clearly seen by the presence of coarse-grained muscovite pseudomorphs after staurolite and various other textures discussed in the previous section. An interpretation of these polymetamorphic textures follows which incorporates evidence for both prograde and retrograde metamorphism.

The interpretation of the metamorphic history of an area relies to a significant extent on microstructural information obtained from thin section petrographic analysis. For example, inferences on the timing of crystallization relative to deformation are commonly based on the microstructural relationships between porphyroblasts and the surrounding matrix. Vernon (1978) lists some common features used to determine temporal relationships between porphyroblast growth and deformation and includes: (1) the nature of the relationship between porphyroblast margins and the surrounding schistosity and (2) the nature of the relationship between internal and external schistosity (S_i and S_e, respectively). Such features have been described earlier for the rocks of this study and are considered here for purposes of ascertaining whether the porphyroblasts are pre-, syn-, or post-tectonic.

Pre-tectonic Crystallization. Both Spry (1969) and Zwart (1962) claim that a later foliation which wraps around a porphyroblast indicates that the porphyroblast is pre-tectonic. However, Vernon (1978, p. 297) suggests alternative interpretations

for such features, while Ferguson and Harte (1975) state that a pre-tectonic interpretation appears clear when there is an internal foliation (S_i) which is inconsistent with external foliation (S_e). Although not common, this is the situation with some of the garnets and staurolites within the Windham-Gorham area.

Some samples show garnets that are characterized by an internal schistosity which is inconsistent with the external foliation (Fig. 5h). From these observations, the garnets are interpreted to be, at least in part, pre-tectonic with respect to the latest deformational event.

Staurolite occasionally exhibits similar evidence suggesting a pre-tectonic texture. Figure 5j is a sketch of a staurolite from within the lower sillimanite zone that is in part, pre-tectonic and in part, post-tectonic. The interior of the grain (upper part in sketch) has a pattern of inclusions which is folded in a manner which is clearly inconsistent with the external foliation within the same thin section. However, the outer portion (lower part of grain in sketch) of the staurolite has an inclusion pattern which is consistent with the external foliation (S_e and S_i passes from one end of the grain to the other with no deflection at grain boundaries). The interior of the grain is interpreted to be pre-tectonic with respect to the deformation that formed S_e whereas the outer portion of the grain is post-tectonic. The interior pattern of inclusions is apparently recording a preexisting deformation of a foliation which has been obliterated by subsequent events. Figure 5i also displays a staurolite grain which exhibits an inclusion pattern which is inconsistent with the external foliation. This pattern is a record of a previous foliation since destroyed by later events.

Post-tectonic Crystallization. Criteria presented by Vernon (1978) for post-tectonic crystallization of porphyroblasts include: (1) S_i is continuous with S_e, being either straight or folded, (2) porphyroblasts are arranged randomly with respect to S_e, (3) S_e is not deflected around porphyroblast boundaries (it is truncated), and (4) porphyroblasts show no evidence of deformation. Post-tectonic growth of porphyroblasts and groundmass minerals may also result in randomly oriented crystals in place of preferred orientation, polygonization of bent grains, and the formation of pseudomorphs (Spry, 1969).

In the study area phases such as biotite, garnet, staurolite, and, in several cases, kyanite, exhibit helicitic overprint texture (Fig. 5a-d). As emphasized by Ferguson and Harte (1975), this texture is probably unequivocal evidence of post-tectonic growth of a porphyroblast. The included minerals within these porphyroblasts are of the same nature as the matrix minerals and clearly pass from one side of the porphyroblast to the other in a manner that is continuous with the external schistosity.

Porphyroblasts which both truncate and deflect an external schistosity may be interpreted as pre-, syn-, or post-tectonic (with regard to S_e) and thus are ambiguous (Ferguson and Harte, 1975). However, the criteria of truncation of S_e by an inclusion-free porphyroblast in the absence of deflection (Fig. 5e-g) suggests post-tectonic porphyroblast growth. Although helicitic overprint textures are rarely observed in kyanite (and Mg-chlorite) in the Windham-Gorham area, the fact that in all cases S_e is abruptly truncated by the porphyroblasts of kyanite appears to be sufficient evidence for suggesting post-tectonic growth (following the assumption that truncation can only originate through the replacement of S_e by a growing porphyroblast - Ferguson and Harte, 1975).

Polygonization of bent grains, random orientation of porphyroblasts and the partial to complete pseudomorphs of muscovite after staurolite are all indicative of post-tectonic metamorphism and have been observed in the Windham-Gorham area.

The evidence suggests that biotite, garnet, staurolite, and kyanite mainly grew after tectonic activity that formed S_e ceased. Inasmuch as the last deformation was localized and associated with the intrusion of the Carboniferous Sebago batholith, the porphyroblasts (or parts of them as in Fig. 5j) are as young as Carboniferous. This would appear to relate well with the distribution of mineral facies involving these phases which are spatially related to the Sebago batholith (Fig. 4).

In summary, study of thin sections from the Windham-Gorham area has revealed that it is a polymetamorphic terrane. The evidence in this and the previous section suggest that, with respect to the deformations that produced the foliations seen in these rocks, minerals such as biotite, garnet and staurolite are in part pre-tectonic but mainly post-tectonic. Phases such as kyanite, chlorite, and sillimanite are post-tectonic. There are at least two periods of porphyroblast growth.

Prograde and Retrograde Metamorphism

The Windham-Gorham area shows evidence of both prograde and retrograde metamorphism. Perhaps the best displayed evidence for this lies in the partial to complete pseudomorphic replacement of staurolite by coarse-grained muscovite in the lower and upper sillimanite zones, respectively. However, a pseudomorph of coarse-grained muscovite, presumably after staurolite, was also observed within the garnet zone (Fig. 4, Location 6; Thomson, 1985). It follows that rocks initially in the staurolite zone of metamorphism were upgraded to sillimanite grade by prograde reactions near the Sebago batholith, but farther away (southeast) were downgraded by retrograde metamorphism. In between (in the present staurolite and kyanite zones), the conditions of the two metamorphic events were similar, such that staurolite remained as a stable phase. Moreover, the textures observed in the staurolite and kyanite zones indicate at least two generations of staurolite growth. Similar relationships were observed by Guidotti (1970) in the Rangeley area of western Maine (see previous discussion in this paper) where metamorphism, like that in the present study area, was related to the intrusion of a pluton.

Further evidence of retrograde metamorphism lies in the observation of the Fe-chloritization of biotite in the lower and upper sillimanite zones. This was a sporadic, localized late

retrogradational event unrelated to the event that caused the pseudomorphism of staurolite discussed above.

Parageneses, Reactions, and Conditions of Metamorphism

The AFM topology changes observed from one metamorphic zone to another may be related by a series of isogradic reactions (depicted by the arrows in Fig. 3). The essential aspects of the isograds observed in the Windham-Gorham area are presented in Figures 3 and 4. Most of the reactions deduced from the AFM topologies are dehydration reactions. The sole exception is the solid-solid transition of kyanite to sillimanite.

Reactions. The detailed reactions shown in Figure 3 are:

(1) Garnet to staurolite zone:
chl + gar + musc = staur + bio + qtz + H_2O

(2) Staurolite to kyanite zone:
staur + chl + musc = ky + bio + qtz + H_2O

(3) Kyanite to lower sillimanite zone:
ky = sill

(4) Lower sillimanite to upper sillimanite zone:
musc + staur + qtz = sill + bio + gar + H_2O

The reactions above are "discontinuous reactions" in that new phases are formed and old phases destroyed by changes in tie line configurations or polymorphic transformations. These discontinuous reactions define the isograds of the study area.

Petrogenetic Grid Geothermometry and Geobarometry. Two approaches were employed to provide an estimate for the conditions of metamorphism. By comparing the observed mineral assemblages and AFM topologic changes with experimental data, one is able to obtain some general estimates of the metamorphic conditions under which the rocks were formed. This method, based on a petrogenetic grid, provides us with a maximum and minimum limit of metamorphic conditions. However, since experimental data (and thermodynamic calculations also) are based on presumed equilibrium conditions, the P-T estimates for natural parageneses are valid only to the extent that the rocks approached equilibrium. It appears that the rocks of the Windham-Gorham area have approached equilibrium based on criteria presented by Vernon (1977). All phases represented in the topologies shown in Figure 3 are in contact with one another and show no evidence of replacement or reaction (except in the case of white mica pseudomorphs after staurolite) and grain boundaries are sharp and smooth. In addition, the present assemblages observed in the study area are common ones and the changes observed, both chemically and mineralogically, are systematic and the expected ones in a progressive metamorphic terrane. Furthermore, each grade of metamorphism may be represented graphically on an AFM topology diagram and the mineral changes may be mapped as isograds.

The positions of the discontinuous reactions that define the isograds have been approximated in P-T space based on thermodynamic calculations and laboratory experiments (Fig. 7). The references for these reactions are shown in the figure caption. It should be noted that there are large uncertainties in the placement of those reactions involving staurolite. The aluminosilicate triple point used is that of Holdaway (1971) and is located at approximately 500°C and 3.8 kbars. Moreover, for simplicity, it is initially assumed that $P_{H2O} = P_{total}$.

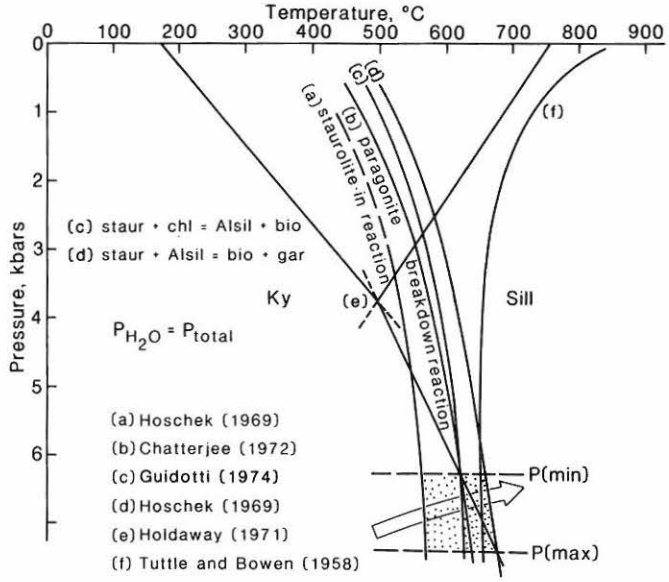


Figure 7. P-T diagram showing minimum and maximum pressures and temperatures of metamorphism in the Windham-Gorham area.

As noted previously, no paragonite was found in the kyanite-bearing rocks. Thus, these rocks must have exceeded the conditions of the paragonite breakdown reaction. Because the paragonite breakdown curve was crossed in the kyanite stability field, the intersection of the paragonite breakdown curve and the kyanite-sillimanite curve marks the minimum pressure (P[MIN]) of metamorphism for these rocks (6.25 kbars).

Although the polymorphic transition of kyanite to sillimanite (reaction 3) is not well documented in the field (for instance, kyanite and sillimanite have not been observed in contact with one another), other evidence suggests that this polymorphic transition has taken place. Most importantly, the breakdown of staurolite occurs wholly within the sillimanite field via reaction 4. Hence, the intersection of reaction 4 with the kyanite-sillimanite transition (based on Holdaway's 1971 data) provides a good estimate of maximum pressure (P[MAX]) with the assumption that $P_{H2O} = P_{total}$ (7.5 kbars).

The petrogenetic grid approach also enables one to roughly estimate the temperature limits of metamorphism. The stippled region in Figure 7 represents the pressure and temperature range under which the rocks of the Windham-Gorham area were metamorphosed (in the staurolite to upper sillimanite zone).

In addition to the petrogenetic grid approach, estimates of the conditions of metamorphism were made using both garnet-biotite geothermometry (Thompson, 1976; Ferry and Spear, 1978) and plagioclase-garnet-Al_2SiO_5-quartz geobarometry (Ghent, 1976) based on exchange equilibria and solid-solid reaction equilibria, respectively.

Exchange Reaction (Garnet-Biotite) Geothermometry. The Mg-Fe^{2+} partitioning between coexisting garnet and biotite varies with temperature but is relatively independent of pressure. Hence, it is a potential geothermometer. The exchange reaction is:

$$\underset{\text{phlogopite}}{KMg_3Si_3AlO_{10}(OH)_2} + \underset{\text{almandine}}{Fe_3Al_2Si_3O_{12}} = \underset{\text{annite}}{KFe_3Si_3AlO_{10}(OH)_2} + \underset{\text{pyrope}}{Mg_3Al_2Si_3O_{12}} \quad (5)$$

and the K_D (distribution coefficient) is [(Mg/Fe)bio/(Mg/Fe)gar] (Thompson, 1976) or [(Mg/Fe)gar/(Mg/Fe)bio] (Ferry and Spear, 1978). The calibrations used in the present study are from Thompson (1976) and Ferry and Spear (1978) at 2.07 kbars.

The results of the garnet-biotite geothermometry are presented in Table 5 for the garnet, staurolite, kyanite, and lower sillimanite zones. Temperature estimates range from 490° - 602°C (Thompson, 1976 calibration) and 477° - 623°C (Ferry and Spear, 1978 calibration). The results are at least grossly consistent with an increased grade of metamorphism from the southeast to the northwest as the Sebago batholith is approached. Temperatures estimated by exchange equilibria compare favorably with those estimated by employment of a petrogenetic grid approach.

TABLE 5. GEOTHERMOMETRY AND GEOBAROMETRY RESULTS

Zone	Garnet	Staurolite		Kyanite		Lo sill
Sample	4	1	29	31	31	46
		Garnet-Biotite Geothermometry, °C				
Thompson (1976)	490	492	490	537	501	602
Ferry and Spear(1978)	477	479	477	537	490	623
		Ideal Solid Solution Model Geobarometry				
X_{An}				.282	.276	.208
X_{gross}				.081	.081	.063
log K_D				-1.630	-1.590	-1.560
T°K				810	763	896
P (bars)				6408	5572	8714
		Geobarometry Modified for Effects of Non-ideality *				
T°K				810	763	896
P (bars)				7429	6788	9613

* See Ghent (1976) for method of calculation.

Plagioclase-Garnet-Al_2SiO_5-Quartz Geobarometry. The assemblage consisting of plagioclase, garnet, quartz, and an aluminosilicate is a common one in regionally metamorphosed metapelites. Ghent (1975, 1976) has shown that the reaction:

$$\underset{\text{anorthite}}{3CaAl_2Si_2O_6} = \underset{\text{grossular}}{Ca_3Al_2Si_3O_{12}} + \underset{\text{Al-sil}}{2Al_2SiO_5} + \underset{\text{quartz}}{SiO_2} \quad (6)$$

provides information which can be calibrated for estimating pressures (and temperatures) of metamorphism. The P-T curves for the pure end-member reaction 6 with kyanite and sillimanite as the aluminosilicate polymorphs have been derived by Ghent (1975, 1976).

Ideal Solid Solution Model

Assuming an ideal solid solution model, a term log K_D is added to each equation where log K_D = 3logXgross - 3logXAn and Xgross = mol CaO/(mol CaO + mol FeO + mol MgO + mol MnO) in garnet and XAn = mol CaO/(mol CaO + mol Na_2O + mol K_2O) in plagioclase. Following the reasoning of Ghent (1976) (in the Esplanade Range, British Columbia), FeO used in the calculations is total FeO.

Two samples from the kyanite zone (Fig. 4, locality 31) and one sample from the lower sillimanite zone (Fig. 4, locality 46) were used to calculate pressures of metamorphism. The temperatures obtained from the Ferry and Spear (1978) calibrations were employed in solving for pressure. Pressures calculated (assuming an ideal solution model) were 5.6 - 6.4 kbars for the kyanite zone and 8.7 kbars for the lower sillimanite zone with an uncertainty of approximately 400 bars (Ghent, 1976) or higher (Lang and Rice, 1985). All information and results are reported in Table 5.

Modification for Effects of Non-Ideality

The correlation of the distribution coefficient with metamorphic grade is complicated by the various substitutions for Fe and Mg in both biotite and garnet (Saxena, 1969). For this reason, the ideal solid solution model may be inappropriate.

Pressures estimated by taking into account non-ideality of phases (see Ghent, 1976) range from 6.8 - 7.4 kbars in the kyanite zone and 9.6 kbars in lower sillimanite zone. The results are tabulated in Table 5 and are considerably higher than those estimated assuming an ideal solid solution model.

Discussion of P-T Results

Table 6 summarizes the results for the conditions of metamorphism calculated and/or estimated by the various methods discussed above. The Ferry and Spear (1978) and Thompson (1976) geothermometer results have a slight discrepancy (Table 5). This can be expected since the two geothermometers were calibrated differently (see Essene, 1982, Fig. 3).

TABLE 6. COMPARISON OF THE TWO APPROACHES FOR ESTIMATING CONDITIONS OF METAMORPHISM.*

	Temperature, °C	Pressure, Kbars
Petrogenetic grid approach	500 - 700	6.25 - 7.5
Geothermometry and geobarometry approach	477 - 623	5.5 - 6.4 (kyanite zone)
		8.7 (lo sill zone)

* Pressures based on an ideal solid solution model.

In general, however, both calibrations lead to results indicating an increase in temperature from the garnet zone to the upper sillimanite zone as the Sebago batholith is approached. Finally, the petrogenetic grid approach envelopes the temperatures calculated using exchange equilibria assuming that $P_{H2O} = P_{total}$. It would appear reasonable to state that the rocks of the Windham-Gorham area were metamorphosed at temperatures from 450° to 700°C.

The petrogenetic grid approach yields pressures from 6 to 8 kbars (20 - 25 km) (Table 6). Pressures calculated using geobarometric calibrations are more difficult to interpret. Pressures calculated for the kyanite zone, assuming an ideal solid solution model, seem fairly reasonable as compared with the petrogenetic approach. However, the pressures calculated for the lower sillimanite zone appear to be high, especially in terms of the spatial model proposed in the next section. The unrealistically high pressure result for the lower sillimanite zone may be attributed to several factors including poor electron microprobe analyses or analysis of phases in disequilibrium (i.e., retrograded rims). Faulty temperature estimates based on exchange equilibria can also change the pressures calculated. For instance, a 10°C increase in temperature will cause about a 200 bar increase in pressures calculated by Ghent's (1976) geobarometer. All of these factors as well as the various assumptions used (i.e., ideal solution model) may lead to problems with the estimate of the conditions of metamorphism. Despite these problems, it can be stated with reasonable certainty that rocks of the kyanite zone were formed under Barrovian-type metamorphic conditions.

GEOLOGIC AND METAMORPHIC HISTORY

It is clear from this study that the Windham-Gorham area is polydeformational and polymetamorphic. In light of the information presented above, a sequence of events is suggested below for the study area. A summary of this discussion is presented in Table 7.

According to Hussey (1985), the Kearsarge-central Maine sequence (including the Vassalboro, Waterville, Sangerville, Perry Mountain, Smalls Falls, and Madrid Formations) exhibits evidence of an early folding event (F_1). These F_1 folds are large-scale recumbent, west-facing isoclines, the evidence for which has been reported by Osberg (1968, 1980) in the Waterville-Vassalboro area. These folds were later deformed by large-scale upright to slightly overturned tight isoclinal folds (F_2) (Hussey, 1985). It is these F_2 isoclinal folds which control the map pattern observed in Figure 2. In this study, the writer has designated this deformational event as D_1, since evidence for the recumbent folding event seen in the north is not observed in the study area (Hussey, 1985). This deformational event is probably Acadian in age.

During, or possibly after the initial (D_1) deformational event, a metamorphic event (M_1) took place and metamorphosed the entire area to at least staurolite grade. Evidence for this event includes partial to complete coarse-grained muscovite pseudomorphs after staurolite as discussed in a previous section. This metamorphic event probably occurred during the Acadian as well. The present work suggests that the intrusion of the Sebago batholith may have been accompanied by localized deformation in the study area. This localized deformational event is referred to as D_2 in the present study.

The area underwent a second metamorphic event (M_2) essentially coincident with the intrusion of the Sebago batholith in Carboniferous time. The metamorphic intensity of this event is spatially controlled by the intrusion of the Sebago batholith. It was a post-tectonic Barrovian-type metamorphism ranging from garnet to upper sillimanite grade. The present isograd pattern is a direct consequence of the 325 Ma metamorphic event. The significance of this post-tectonic metamorphism

TABLE 7. DEFORMATIONAL AND METAMORPHIC HISTORY OF THE WINDHAM-GORHAM AREA.

Time	Deformation	Metamorphism	Characteristics
Acadian	D_1		Upright, tight isoclinal folding leading to the present map pattern.
Acadian		M_1	To staurolite grade throughout the entire study area.
Carboniferous	D_2		Localized deformation as a result of the intrusion of the Sebago batholith (325 Ma).
Carboniferous		M_2	Post-tectonic Barrovian-type metamorphism ranging from garnet to upper sillimanite zone. Produced the present isograd pattern causing a "hinge effect" on the M_1 assemblages.
Carboniferous (?)		M_3	Retrograde metamorphism near batholith causing partial chloritization of biotite.

produced by the emplacement of the Sebago batholith will be discussed in the next section.

The M_2 metamorphic event caused prograde metamorphism of staurolite grade rocks to the upper sillimanite zone near the batholith. Staurolites from the M_1 event are partially to completely replaced by coarse-grained muscovite in the lower and upper sillimanite zones, respectively. On the other hand, the M_2 event caused retrogradation of staurolite grade rocks to the garnet zone in the south as seen by the presence of a coarse-grained muscovite pseudomorphs after staurolite in one locality (discussed previously). Hence, the area has a "hinge effect" much like that discussed by Guidotti (1970) in the Rangeley area of Maine. The M_2 staurolite and kyanite zone metamorphic temperatures were probably roughly similar to those during the M_1 metamorphic event. For this reason, the staurolite in these zones shows no evidence of pseudomorphism. Instead, as discussed in a previous section with respect to Carboniferous age events, the staurolites (and garnets) exhibit both pre-tectonic (M_1) and post-tectonic (M_2) textures.

Finally, at some time after M_2, a third metamorphism occurred (M_3) which caused retrograde metamorphism in at least the northwestern part of the study area. The evidence for this localized event lies in the observation of the Fe-chloritization of biotite in the lower and upper sillimanite zones near the Sebago batholith. The event was probably caused by the late movement of fluids.

REGIONAL GEOLOGIC IMPLICATIONS

In the context of the Caledonide orogen, post-Acadian plutonism and metamorphism is well documented in the Canadian and southern Appalachians and western Europe. Plutons in the southern Appalachians exhibit ages ranging from 265 - 325 Ma (Fullagar and Butler, 1979). In New England, small-scale Permian metamorphism has been documented in Rhode Island and eastern Massachusetts (O'Hare and Gromet, 1985).

Recent studies in southern Rhode Island, New Hampshire, and Maine have indicated Carboniferous and younger ages for granitic plutons. Kocis et al. (1978) found the Narragansett Pier granite in southern Rhode Island to be 257 Ma. A pluton in Milford, N.H. has been dated as 275 Ma (Aleinkoff et al., 1979) and the Lyman pluton of southern Maine as 322 Ma (Gaudette et al., 1982). Most recently, the Sebago batholith of southern Maine has been found to be 325 Ma by two independent dating techniques (Aleinikoff, 1984; Aleinikoff et al., 1985; Hayward and Gaudette, 1984). Thus, the known range for post-Acadian New England plutons is 275 - 325 Ma.

Lux and Guidotti (1985) and Guidotti et al. (1986a, 1986b) present evidence for Carboniferous metamorphic activity in the area on the north side of the Sebago batholith. $^{40}Ar/^{39}Ar$ ages for hornblendes from two Acadian plutons which are cut by the Sebago batholith display disturbed to re-set release spectra indicative of cooling ages near 308 Ma. This, together with petrologic observations, suggests that the high-grade K-feldspar + sillimanite metamorphism in that area occurred in Carboniferous rather than Devonian time and was closely associated with the intrusion of the Sebago batholith. Lux and Guidotti (1985) suggest that the high grade-metamorphic conditions had ended by about 300 Ma so the Carboniferous metamorphism occurred between 325 - 305 Ma.

The present study provides additional evidence of a Carboniferous metamorphism in Maine. However, it suggests important differences from other metamorphic regimes in Maine in that the evidence indicates that this was a medium-pressure, Barrovian-type metamorphism. A mechanism for attaining the pressures necessary to produce kyanite-grade metamorphism must be considered.

Spatial Relations of the Sebago Batholith and the Metasediments of the Windham-Gorham Area

Gravity studies of the Sebago batholith suggest that it is a thin (approx. 1 km) tabular sheet, dipping gently to the north (Kane and Bromery, 1968; Hodge et al., 1982). Alternatively, Guidotti (pers. commun., 1985) has calculated its thickness based on the map pattern to be as much as 4 km. Its maximum width is nearly 90 km in a northwest-southeast direction. Holdaway et al. (1982) and Cheney and Guidotti (1979) suggested a depth of emplacement of 10-12 km (3.5 kbars). More recently, thermal modeling for the intrusion of the batholith suggests that it was intruded at a depth of about 14 km (about 4 kbars) (DeYoreo et al., 1985; Guidotti et al., 1986a, 1986b). Previous studies on the metamorphism to the north of the Sebago batholith (Evans and Guidotti, 1966; Guidotti, 1970) indicate a spatial relationship of the K-feldspar + sillimanite isograd with the batholith. Evans and Guidotti (1966) and Guidotti (1970) suggest that the isogradic surfaces are essentially flat-lying and parallel to the tabular Sebago batholith. Petrogenetic grid information indicates that the rocks to the north of the Sebago batholith were metamorphosed above the batholith at pressures of about 3.3 kbars.

In this study it has been argued that the post-tectonic Barrovian-type metamorphism (M_2) is related to the intrusion of the batholith and thus is Carboniferous in age. It would appear that this metamorphism is related to that in western Maine discussed by Lux and Guidotti (1985).

Following the reasoning of Guidotti (1970), the isograds of the present study are presumed to be essentially flat-lying. They are irregularly trending, widely spaced isograds, as would be expected if they are flat-lying. Although the present work has shown that the isograds are spatially related to the Sebago batholith, the pressures estimated by both a petrogenetic grid approach and geobarometric approach are considerably higher than those estimated on the north side of the batholith. In the context of the Sebago batholith being a thin sheet with a gentle northerly dip, it is suggested that the Windham-Gorham area was metamorphosed on the underside of the batholith during Car-

boniferous time with the garnet zone at the deepest level and the upper sillimanite zone at the shallowest level, closest to the batholith. The arrow depicted in Figure 7 represents a possible P-T trajectory for the rocks of the study area.

Figure 8 presents a schematic cross-section of the crust just after the intrusion of the batholith in Carboniferous time. It also shows the approximate location of the Windham-Gorham area (based on regional geologic setting plus the petrogenetic grid and geobarometry calculations) as well as the Rangeley area to the north of the batholith studied by Lux and Guidotti (1985).

Field studies suggest that the uplift rate to the south was slightly greater than that of the north (based on higher metamorphic grades in the south) so that different levels of the crust are exposed at the surface today. The dashed line in Figure 8 represents the present erosional surface assuming that the models suggested by field studies are correct. It appears that the Windham-Gorham area is the surface expression of a deeper level of the crust preserving rocks of a Carboniferous metamorphic event.

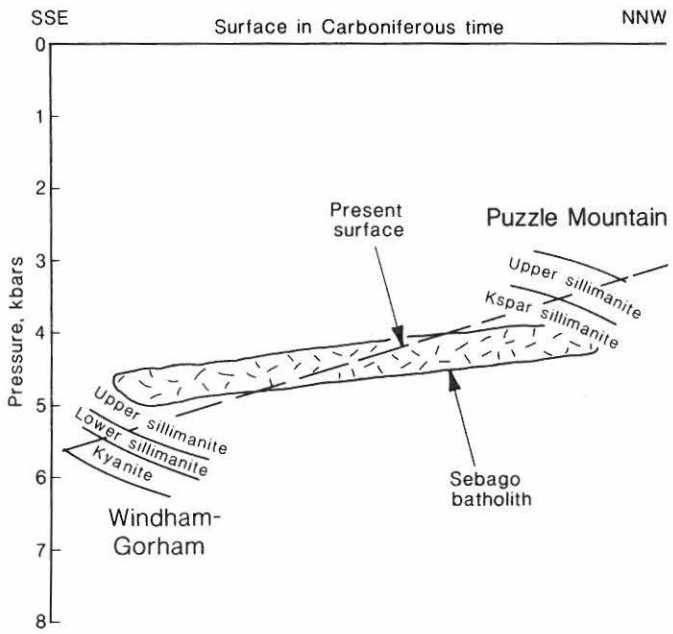


Figure 8. Schematic cross-section through the earth's crust showing the surface during Carboniferous time and the present erosional surface.

ACKNOWLEDGMENTS

The writer greatly appreciates professional support from C.V. Guidotti, A. M. Hussey II and P. H. Osberg. Thanks are also extended to P. Maggiore for endless help with electron microprobe analyses, H. Lang for a lengthy and thorough review, an anonymous reviewer's constructive comments, and A. Young for manuscript preparation. Financial support was provided by the Department of Geological Sciences and the Graduate Student Board, University of Maine at Orono, Sigma Xi, and EPSCOR Grant ISP-8011448 and NSF Grant EAR-8204254 (to C.V. Guidotti).

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Maine Geological Survey
Studies in Maine Geology: Volume 3
1989

Implications of the Rb-Sr and O Isotopic Systematics and Geochemistry of Some Two-mica Granites in Northern New England

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ABSTRACT

Rb-Sr whole rock analyses of six two-mica granites in northern New England yield whole rock Devonian and Carboniferous crystallization ages: Chelmsford granite - 399±6 Ma; Effingham granite - 319±13 Ma; Fitzwilliam granite - 383±6 Ma; Hooksett granite 400±7 Ma; Sebago granite - 335±6 Ma; and Sunapee granite - 364±5 Ma. Rb-Sr mineral ages of the Effingham (263±4 Ma), Fitzwilliam (326±15 Ma; 267±12 Ma), and Sebago (268±9 Ma) granites confirm a Carboniferous to Permian mineral resetting event in northern New England as observed in other studies. The Chelmsford (0.7102±0.0009), Fitzwilliam (0.7090±0.0005), and Sunapee (0.7133±0.0006) granites yield the highest initial $^{87}Sr/^{86}Sr$ ratios and are the only granites among those investigated to yield initial $^{87}Sr/^{86}Sr$ ratios >0.708. The other granites, the Effingham (0.7048±0.0006), Hooksett (0.7066±0.0013), and Sebago (0.7048±0.0006), yield initial $^{87}Sr/^{86}Sr$ ratios that fall within the range reported for I-type granites. This latter group of granites crops out in the central portion of the Kearsarge-central Maine synclinorium. The distribution of initial $^{87}Sr/^{86}Sr$ ratios suggests a geographic trend in northern New England, with initial $^{87}Sr/^{86}Sr$ ratios > 0.7085 obtained from granites that were emplaced immediately east of the Bronson Hill anticlinorium (Fitzwilliam and Sunapee), or southeast of the Norumbega fault system (Chelmsford), while the granites of the central part of the synclinorium yield initial $^{87}Sr/^{86}Sr$ ratios <0 .7066.

Oxygen isotope data for the Chelmsford (+10.5 to +12.6), Fitzwilliam (+8.6 to +11.1), and Sebago (+7.5 to +11.1) granites imply the involvement of a source or contaminant subjected to low temperature alteration in the generation of at least these three granites. Petrography and the available CIPW norms are consistent for the granites of this study and yield peraluminous characteristics (primary muscovite and biotite; corundum-normative). The low initial Sr ratios of the central Kearsarge-central Maine synclinorium granites, however, suggest that a Rb-depleted source is more likely for these granites than the Rb-enriched source indicated by initial $^{87}Sr/^{86}Sr$ ratios for the Chelmsford, Fitzwilliam, and Sunapee granites.

INTRODUCTION

Analyses of the Appalachian orogen with regard to its tectonostratigraphic terranes have stimulated studies designed to assess the possibilities for discrimination between the crust-basement packages that have been identified by geophysical and geochemical examinations throughout the Appalachians (e.g., Ando et al., 1983; Ando et al., 1984; Brown et al., 1983; Stewart et al., 1986; Williams and Hatcher, 1983; Zen, 1983). Recently, lead-isotopic data collected by Ayuso (1986) from Devonian granites and country rock in Maine, and oxygen and strontium isotopic data collected by Andrew et al. (1983) from granites in

central and eastern Maine have demonstrated that isotopic discrimination between basements or granite sources in Maine is possible.

The purpose of this study is to consider the petrographic, Rb-Sr and O isotopic, and other geochemical characteristics of some two-mica granites of the Kearsarge-central Maine synclinorium of northern New England in light of the types of studies discussed above. Included in this study are the Sebago granite of southwestern Maine; the Effingham, Fitzwilliam, Hooksett, and Sunapee granites of New Hampshire; and the Chelmsford granite of northeastern Massachusetts (Fig. 1). Previous Rb-Sr isotopic work pertaining to the Lyman granite of southwestern Maine (Gaudette et al., 1982) will also be included as the Sebago and Effingham granites appear to be closely related to the Lyman granite spatially (see Fig. 1) and petrographically. The characterization of these granites and their source(s) has tectonic and petrogenetic implications for the New England region, as well as significance with regard to fundamental questions of mantle-crust interaction dynamics.

GENERAL GEOLOGIC OCCURRENCE, PETROGRAPHY, AND GEOCHEMISTRY

The plutons examined in this study crop out within the broad lithotectonic province known as the Kearsarge-central Maine synclinorium. Major northeast-trending faults suggest general east-west divisions within the synclinorium to which the granites may be relegated (see Fig. 1). Thus, the Chelmsford granite may be described as occupying the most southeasterly of these divisions, a zone that lies northwest of the Clinton-Newbury fault and southeast of the Flint Hill and Nonesuch River-Norumbega faults and the Massabesic anticlinorium. The Hooksett and Lyman granites occupy a zone northwest of the Nonesuch River-Norumbega fault and southeast of the Concord tectonic zone.

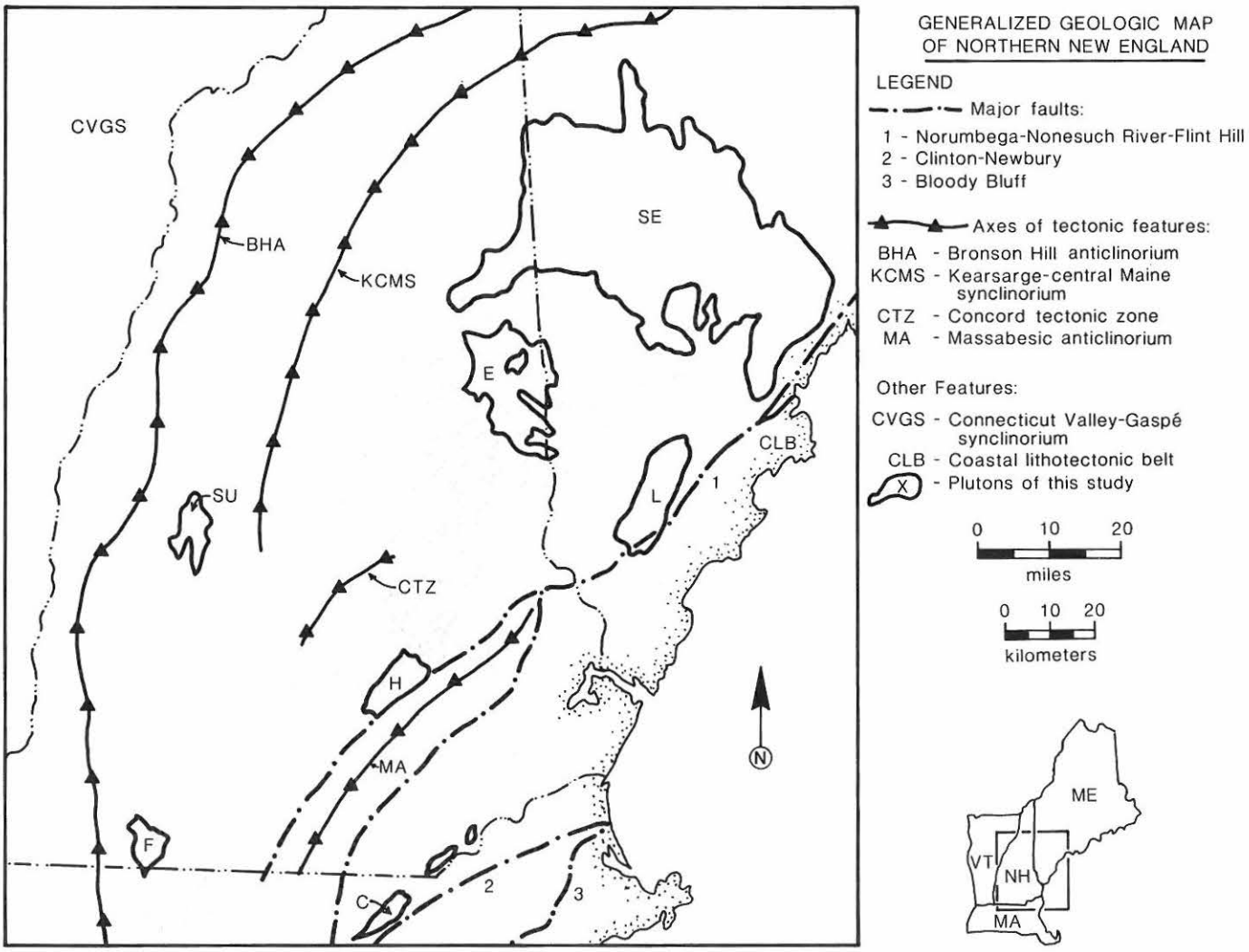


Figure 1. Generalized geologic map of northern New England plutons of this study. Chelmsford (C), Effingham (E), Fitzwilliam (F), Hooksett (H), Lyman (L), Sebago (SE), and Sunapee (SU).

The Lyman granite is actually cut by the Nonesuch River-Norumbega fault on its southern end and may be slightly offset by the fault (Gaudette et al., 1982). The Effingham and Sebago granites lie to the north of an eastern extension of the Concord tectonic zone, east of the axis of the Kearsarge-central Maine synclinorium, while the Fitzwilliam and Sunapee granites crop out to the west of the synclinorium axis and immediately to the east of the Bronson Hill anticlinorium. The exact nature of the basement(s?) underlying the Kearsarge-central Maine synclinorium is unknown, with both Gander- and Avalon-type basements being suggested as possible candidates (e.g., Williams and Hatcher, 1983 and Lyons et al., 1982, respectively).

Nielson et al. (1976) and Hodge et al. (1982) determined through gravity studies that the two-mica granites of New Hampshire and Maine are thin, sheet-like bodies, having thicknesses of up to 2.5 km. These granites are generally massive and homogeneous in their centers and become more dike-like toward the outer perimeters. In most cases the granites may be described as pristine. Pegmatites and xenoliths of the surrounding country rock are associated with the two-mica granites near the zone of contact with the country rock (Fowler-Billings, 1949; Chapman, 1952; Billings, 1956; Virgin, 1965; Gaudette et al., 1982; Hayward, 1983b; Aleinikoff et al., 1985).

All of the granites of this study intrude high-grade metamorphic rocks. The New Hampshire granites intrude Paleozoic metasediments of amphibolite facies (Billings, 1956). The Sebago granite metamorphosed and deformed the sediments into which it was emplaced. These sediments are of medium to high rank amphibolite facies. The Lyman granite intrudes metasediments of low to medium amphibolite and epidote-amphibolite facies (Osberg et al., 1985).

Additionally, many of the two-mica granites of northern New England crosscut the three older members of the New Hampshire Plutonic Series (Billings and Keevil, 1946; Chapman, 1952; Billings, 1956; Lyons and Livingston, 1977). The Spaulding, Bethlehem, and Kinsman intrusive suites comprise the three older units of the New Hampshire Plutonic Series and were emplaced during the Acadian tectonic event. The two-mica granites of this series have been grouped together as the Concord intrusive suite (Lyons et al., 1982). The intrusion of the two-mica granites was post-tectonic (Fowler-Billings, 1949; Billings, 1956; Lyons and Livingston, 1977; Lyons et al., 1982). Though there are mineralogical similarities between the intrusive suites, they do not represent a differentiation series. Use of the New Hampshire Plutonic Series terminology in this study is strictly historical.

The primary mineralogy of the granites of this study consists of quartz, potassium feldspar (almost invariably microcline that can be perthitic in texture; rarely orthoclase), plagioclase (An_{3-28}), biotite, and muscovite. Primary muscovite ranges from one to twelve modal percent and has been identified according to the criteria established by Miller et al. (1981). Accessory minerals vary in occurrence and can include garnet, magnetite, ilmenite, rutile, titanite, apatite, allanite, monazite,

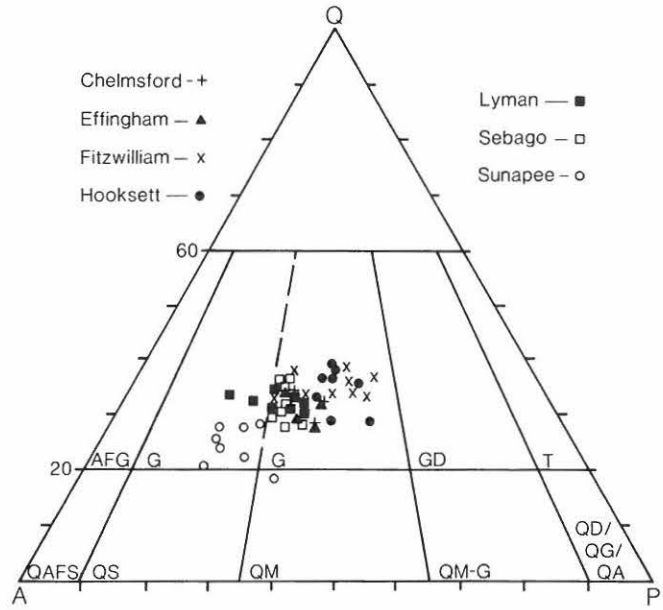


Figure 2. Q-A-P modal composition plot of the two-mica granites of this study. Q-A-P diagram fields: AFG = alkali feldspar granite, G = granite, GD = granodiorite, T = tonalite, QAFS = quartz alkali feldspar syenite, QS = quartz syenite, QM = quartz monzonite, QM-G = quartz monzo-gabbro, QD/QG/QA = quartz diorite/quartz gabbro/quartz anorthosite.

and zircon. Secondary muscovite is a common alteration product of plagioclase, commonly oriented along the twin lamellae and cleavage planes. Epidote, chlorite, and carbonate are also typical secondary minerals in these granites. No geographic pattern in modal mineralogy has been discerned for these granites.

As can be seen from Figure 2, a Q-A-P modal composition plot (Streckeisen, 1976), the granites display little range in composition. Published (Billings and Wilson, 1964) and unpublished (U.S. Geological Survey, courtesy of E. L. Boudette, 1983; H. E. Gaudette and J. Wilband, 1984) weight percent analyses and CIPW norms of some granites of the Concord intrusive suite are provided in Appendix I. With the exception of one sample, E460, all of the analyses are corundum-normative. Primary muscovite and biotite are present in the modes of all of the granites and this result is therefore not surprising. The CIPW norms are misleading, however. Armstrong and Boudette (1985) discussed the inadequacy of the CIPW norm to characterize two-mica granites as a result of the absence of normative micas in this system. Indeed, normative hypersthene and diopside, among other normative minerals, appear in some of the analyses in spite of the fact that they are not mineralogically present in the rocks themselves. Thus, the primary mineralogy of granites provides a more accurate characterization of the granites than do the CIPW norms, as has been advocated by others, notably Miller (1985).

Chappell and White (1974) recognized two different granite types in the Lachlan fold belt that could be discriminated on a

geochemical basis in terms of source material. These two granite types were designated S-type, implying a sedimentary source, and I-type, implying an igneous source, respectively. These designations correspond roughly to Ishihara's (1977) ilmenite (S-type) and magnetite (I-type) series.

Two-mica granites often exhibit a number of characteristics that have been ascribed to S-type granites including mineralogy and isotopic signatures. The genesis and evolution of two-mica granites, however, present a number of problems to investigators. For instance, the initial Sr ratios determined for some two-mica granites in New Hampshire and Maine are surprisingly low (<.707; Gaudette et al., 1982; Hayward, 1983a, 1983b; Hayward and Gaudette, 1984), implying that there was little or no involvement of a radiogenic ^{87}Sr-enriched component in the genesis of these particular plutons.

PREVIOUS ISOTOPIC WORK

U-Pb* zircon analyses of the Chelmsford granite yielded an average ^{207}Pb/^{206}Pb age of 389±5 Ma (Zartman and Naylor, 1984). This result is in agreement with the average ^{207}Pb/^{206}Pb age of 433±5 Ma obtained for the Ayer granite (Zartman and Naylor, 1984), a body that field relations indicate is cut by the Chelmsford granite. The results of Rb-Sr* whole rock analyses of New Hampshire granites of this study and the Sebago granite, and Rb-Sr mineral analyses for the Effingham pluton, were summarized in meeting abstracts (Hayward, 1983a; Hayward and Gaudette, 1984) and will be presented in greater detail below. Aleinikoff et al. (1985) determined an U-Pb crystallization age of 325±3 Ma from zircons and monazites from two whole rock samples of the Sebago granite, in agreement with the Rb-Sr whole rock age of 332±11 Ma obtained by Hayward and Gaudette (1984).

The Sunapee granite, analyzed by Lyons and Livingston (1977), produced two Rb-Sr whole rock isochrons, one yielding an age of 351±11 Ma with an initial ^{87}Sr/^{86}Sr ratio of 0.7107±0.0009, and a second yielding an age of 323±3 Ma with an initial Sr ratio of 0.7144±0.0004. The isochron that gave the younger age was derived through the omission of one granodioritic sample that plotted below the 351 Ma isochron. The Sunapee granite was included in this study due to the wide variation in age and initial ^{87}Sr/^{86}Sr ratio produced by the two distinct isochrons of the original study.

The Milford granite has been analyzed isotopically by Aleinikoff et al. (1979). The zircon and monazite U-Pb data yielded an age of 275±10 Ma. The authors reported the presence of two zircon populations in the Milford granite and interpreted this finding as strong evidence that the granite was produced through crustal anatexis, possibly of the Massabesic paragneiss that the granite intrudes.

In addition to the two Carboniferous ages derived from the Sunapee and Milford granites, Gaudette et al. (1982) obtained a Rb-Sr whole rock age of 322±12 Ma and initial Sr ratio of 0.7044±0.0009 from the Lyman pluton of southwestern Maine. This pluton has been chronologically and petrographically associated with the Sebago and Effingham granites (Hayward and Gaudette, 1984).

METHODS

Fresh fifty kilogram samples of the granites were selected, crushed, and pulverized for Rb-Sr and O isotope analyses using standard procedures such as those outlined by Jeffrey and Hutchinson (1981). Minerals were obtained through a combination of magnetic and heavy liquid separations. Sample digestion for Rb-Sr whole rock and mineral isotope dilution analyses was accomplished through the introduction of an HF-$HClO_4$ mixture to acid-cleaned teflon capsules containing a known mass of sample that had been mixed previously with known masses of ^{87}Rb and ^{84}Sr and evaporated to dryness. The acid digestion was conducted in a loosely covered capsule at atmospheric pressure and 150°C on a hot plate for twelve hours or overnight. The caps were then removed from the capsules and the HF-$HClO_4$ mixture was evaporated at 150°C, and then completely dried at 200°C. After driving off the HF-$HClO_4$ mixture from the capsule, the sample was further digested and converted to chloride form by the consecutive introduction of 6.2 N HCl and 2.0 N HCl. When chloride conversion was complete, the sample was taken up in 1.0 ml of 2.0 N HCl. A calibrated volume of HCl was passed through Dowex 50wx-8 (200-400 mesh chloride form) cation exchange resin to separate Rb and Sr from 0.1 ml aliquots of each sample. The Rb and Sr separates were dried on a hot plate at 150°C, converted to nitrate form, and then run on Ta filaments in a Ta_2O_5-H_2O slurry on the nine-inch radius, sixty-degree deflection, Nier-type solid source mass spectrometer at the University of New Hampshire (see Hayward, 1983b for more details). Rb-Sr analyses of the Chelmsford granite and some of the samples of the Sebago granite were performed using the loading procedure of Popp et al. (1986), and therefore the Rb and Sr separates were simply dried and loaded as chlorides. Sr analyses were conducted for 40 cycles of data; 4 measurements of the ^{87}Sr/^{86}Sr ratio, and one each of the ^{84}Sr/^{86}Sr, ^{85}Sr/^{86}Sr, and ^{88}Sr/^{86}Sr ratios per cycle were collected and averaged. Rb analyses were conducted for 20 cycles of data; 2 ratios of ^{87}Rb/^{85}Rb measurements were collected and averaged per cycle.

Whole rock and mineral ages and initial Sr ratios were calculated by the York (1969) least squares regression method. Mean standard weighted deviation values give an indication of the quality of the data for each isochron. On the basis of replicate analyses, the reproducibility of the ^{87}Rb/^{86}Sr and ^{87}Sr/^{86}Sr ratios is better than 0.24% and 0.09% respectively, for the whole rock samples. Errors for age and initial ratio calculations are given at 95% confidence interval (2 sigma). Oxygen isotope values are

* All U-Pb and Rb-Sr ages reported or discussed in this paper have been calculated or recalculated using ^{238}Uλ =1.55125 x 10^{-10}/yr and ^{235}U λ=9.8485 x 10^{-10}/yr or ^{87}Rbλ =1.42

TABLE 1. Rb/Sr CONCENTRATIONS AND Rb/Sr, $^{87}Rb/^{86}Sr$, AND $^{87}Sr/^{86}Sr$ RATIOS FOR WHOLE ROCK SAMPLES.

Pluton	Rb (ppm)	Sr (ppm)	Rb/Sr	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$
Chelmsford					
JAH8521	293	47	6.30	0.8148	18.40
22	252	82	3.09	0.7613	8.97
23	294	75	3.93	0.7764	11.45
24	228	68	3.34	0.7656	9.72
25	244	60	4.07	0.7764	11.85
26	234	65	3.58	0.7702	10.43
27	287	91	3.16	0.7626	9.17
28	272	94	2.89	0.7577	8.42
30	277	99	2.79	0.7561	8.12
Effingham					
UNH 289	298	239	1.25	0.7215	3.62
290	274	236	1.16	0.7203	3.38
291	276	235	1.17	0.7203	3.41
292	226	271	0.83	0.7159	2.41
458	270	242	1.16	0.7194	3.24
459	225	239	1.01	0.7181	2.98
460	237	239	0.99	0.7174	2.73
Fitzwilliam					
NH/MND 1	206	142	1.45	0.7318	4.21
2	209	141	1.48	0.7320	4.30
3	187	141	1.33	0.7301	3.86
5	220	119	1.85	0.7390	5.35
6	238	100	2.38	0.7463	6.88
7	235	46	5.11	0.7864	14.21
8	232	129	1.80	0.7380	5.23
9	220	130	1.69	0.7358	4.92
10	213	132	1.61	0.7341	4.68
Hooksett					
NH/SUNC1	286	53	5.40	0.7967	15.80
2	260	184	1.41	0.7302	4.09
3	275	59	4.66	0.7838	13.65
4	240	47	5.11	0.7929	14.76
5	292	52	5.62	0.7978	16.30
6	280	61	4.59	0.7838	13.48
7	257	52	4.94	0.7883	14.35
8	222	77	2.88	0.7535	8.38
Lyman					
LY-1	154	316	0.49	0.7106	1.39
2	295	133	2.22	0.7339	6.33
3	226	123	1.85	0.7291	5.27
4	318	95	3.34	0.7480	9.55
5	220	301	0.73	0.7133	2.08
Sebago					
UNH 464	245	119	2.06	0.7333	5.98
465	284	306	0.93	0.7177	2.69
466	311	76	4.09	0.7612	11.88
467	187	320	0.58	0.7128	1.69
469	299	233	1.29	0.7231	3.73
JAH85161	262	243	1.08	0.7220	3.13
163	288	119	2.42	0.7417	7.02
164	305	112	2.72	0.7432	7.90
166	334	88	3.80	0.7572	11.02
167	267	154	1.73	0.7276	5.04
168	256	339	0.76	0.7147	2.18
169	259	349	0.74	0.7139	2.15
Sunapee					
NH/SUNP1	142	85	1.67	0.7387	4.85
2	136	71	1.92	0.7423	5.57
3	151	84	1.80	0.7405	5.27
4	151	106	1.42	0.7342	4.11
6	184	51	3.61	0.7681	10.54
7	210	44	4.77	0.7849	13.75
8	210	43	4.88	0.7864	14.21

given relative to standard mean ocean water (SMOW) and were provided by Geochron Laboratories of Cambridge, MA.

ISOTOPIC RESULTS

Rb and Sr concentrations, Rb/Sr, $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ ratios of whole rock samples obtained from the granites of this study are displayed in Table 1. The Rb and Sr concentrations, Rb/Sr, $^{87}Rb/^{86}Sr$ and $^{87}Sr/^{86}Sr$ ratios for the mineral samples are displayed in Table 2. Rb-Sr isochrons are furnished in Figures 3-8. Latitudinal and longitudinal coordinates are provided in Appendix II for each of the samples analyzed in this study.

Figure 3 is the Rb-Sr isochron for the Chelmsford granite. The nine whole rock samples of the granite yield an isochron with an age of 399±6 Ma and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7102±0.0009. This age correlates well with the U-Pb zircon age of 389±5 Ma obtained by Zartman and Naylor (1984) for the same body. The samples analyzed for this study do not appear petrographically to have undergone recrystallization, and the age obtained is therefore interpreted as the age of crystallization of the Chelmsford granite.

The Effingham granite yielded a Rb-Sr age of 319±13 Ma and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7048±0.0006 (Fig. 4a), in striking similarity to the age and initial ratio data obtained from the Lyman and Sebago plutons of southwestern Maine (see below). Mineral separates yielded a Rb-Sr age of 263±4 Ma (Fig. 4b).

Figure 5a is the Rb-Sr whole rock isochron for the Fitzwilliam granite of New Hampshire. The age derived is well constrained at 383±6 Ma, with an initial ratio of 0.7090±0.0005. The Fitzwilliam granite crosscuts the Spaulding quartz diorite which has been dated at 394±5 Ma (Lyons and Livingston, 1977). The isotopic results for these bodies are therefore consistent with relationships observed in the field. The mineral isochrons for two samples of the Fitzwilliam granite are displayed in Figures 5b and 5c. The mineral ages obtained for the two samples, 326±15 Ma and 267±12 Ma, confirm a Carboniferous to Permian thermal event in southern New Hampshire as was seen also in the Effingham granite.

Eight whole rock samples of the Hooksett granite were analyzed for their Rb and Sr compositions and concentrations.

TABLE 2. Rb/Sr CONCENTRATIONS AND Rb/Sr, $^{87}Rb/^{86}Sr$, AND $^{87}Sr/^{86}Sr$ RATIOS FOR MINERAL SAMPLES.

Pluton		Rb (ppm)	Sr (ppm)	Rb/Sr	$^{87}Sr/^{86}Sr$	$^{87}Rb/^{86}Sr$
Effingham						
UNH 458	I	1328	21	62.96	1.4311	194.64
	II	903	104	8.65	0.7913	25.27
	III	377	200	1.89	0.7267	5.48
	IV	95	224	0.42	0.7121	1.23
Fitzwilliam						
NH/MND3	I	961	13	73.92	1.7119	226.37
	II	155	22	7.05	0.8591	28.63
	III	317	119	2.66	0.7487	7.71
	IV	109	142	0.77	0.7230	2.22
NH/MND9	I	958	18	53.22	1.3401	168.08
	II	148	16	9.25	0.8303	27.25
	III	186	333	1.79	0.7502	1.73
	IV	152	133	1.14	0.7299	3.32
Sebago						
UNH465	II	1273	42	30.31	1.0678	91.18
	III	1132	112	10.11	0.8200	29.48
	V	881	106	8.31	0.7939	24.29
	VI	610	234	2.61	0.7372	7.57
	I	75	115	0.65	0.7178	1.89

The isochron produced by these samples is shown in Figure 6. The samples form a distinct line yielding an age of 400±7 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7066±0.0013.

Since Hayward and Gaudette's (1984) study of the Effingham and Sebago plutons, additional Rb-Sr whole rock analyses of the Sebago granite have been made. The original data are displayed along with the more recent results (Fig. 7a). Considered together, the data yield an age of 335±6 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7048±0.0006. The additional data are in excellent agreement with those of the original study and with the U-Pb results of Aleinikoff et al. (1985). Figure 7b is the isochron for UNH 465 whole rock and mineral samples. The data yield a mineral age of 268±9 Ma and an initial Sr ratio of 0.7084±0.0015 (pers. commun., H. E. Gaudette, 1984).

Figure 8 displays the Rb-Sr whole rock isochron of the Sunapee granite. The age is well constrained at 364±5 Ma, with an initial $^{87}Sr/^{86}Sr$ ratio of 0.7133±0.0006. The Sunapee granite crosscuts the Kinsman quartz monzonite and the Bethlehem gneiss, which have been dated by the Rb-Sr method at 402±19 Ma and 396±78 Ma respectively (Lyons and Livingston, 1977). The Kinsman quartz monzonite has also yielded a Sm-Nd age of 413±5 Ma that has been interpreted as the age of crystallization (Barreiro and Aleinikoff, 1985).

Oxygen isotope analyses have been made for some samples of the Chelmsford, Fitzwilliam and Sebago granites and surrounding country rock. These data are included in Table 3. In general, the granites exhibit an increase in delta ^{18}O value with increasing age and initial $^{87}Sr/^{86}Sr$ ratio. The average delta ^{18}O values for the plutons are ≥10‰, the lower limit reported by Miller (1985) for peraluminous granites. Remarkably, the

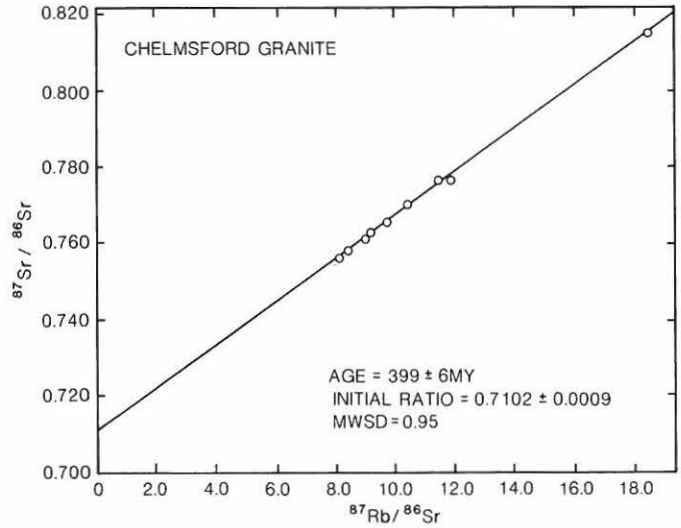


Figure 3. Rb-Sr whole rock isochron for the Chelmsford granite.

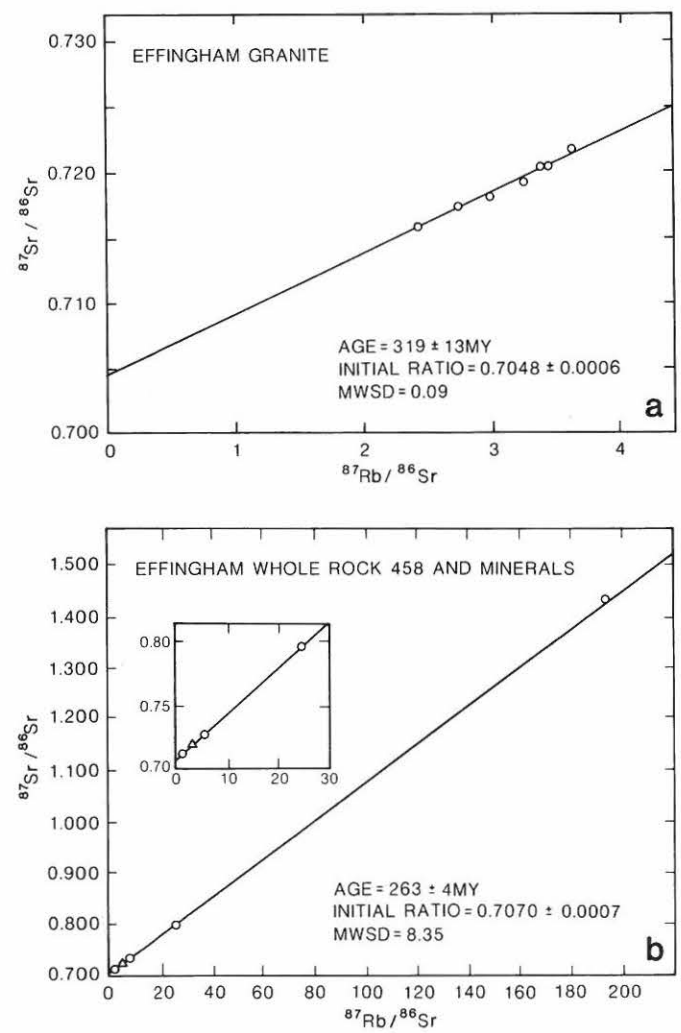
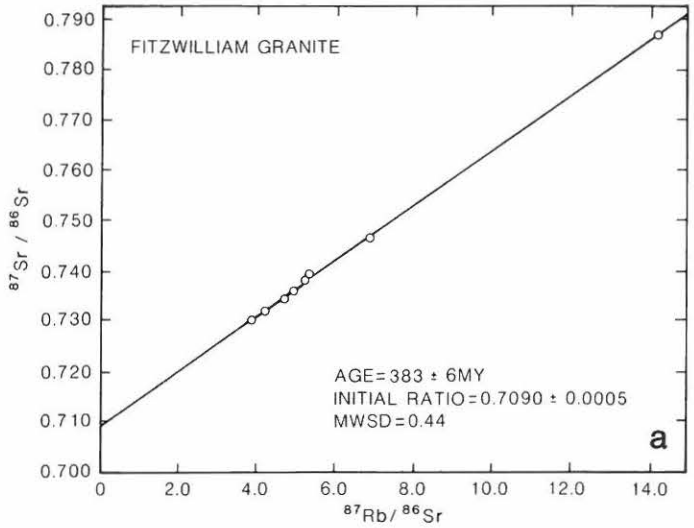
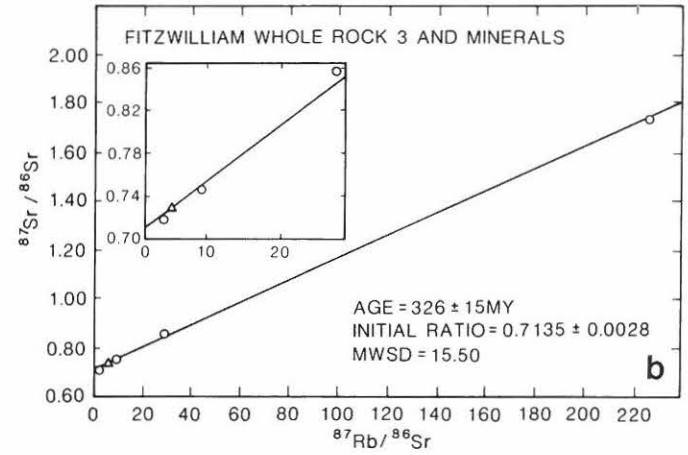


Figure 4. Rb-Sr isochron for the Effingham granite. (a) Whole rock, all samples. (b) Sample UNH 458; whole rock (*) and minerals (+).





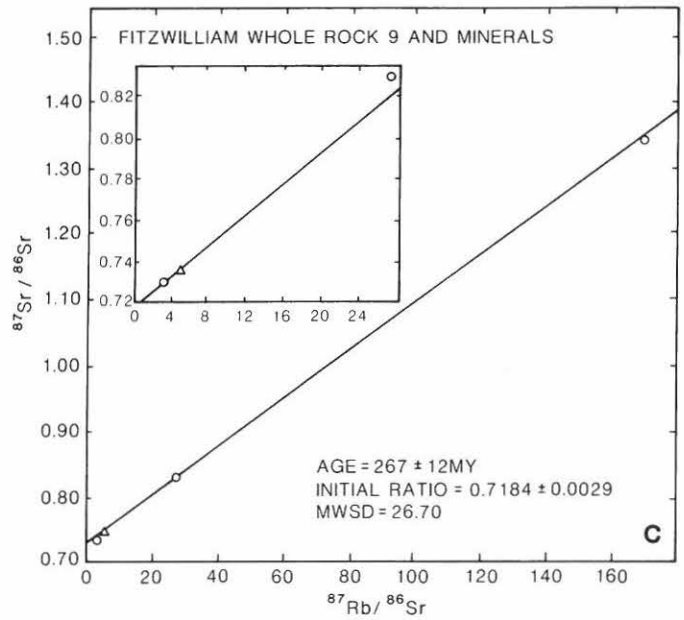


Figure 5. Rb-Sr isochron for the Fitzwilliam granite. (a) Whole rock, all samples. (b) Sample NH/MND 3; whole rock (*) and minerals (+). (c) Sample NH/MND 9; whole rock (*) and minerals (+).

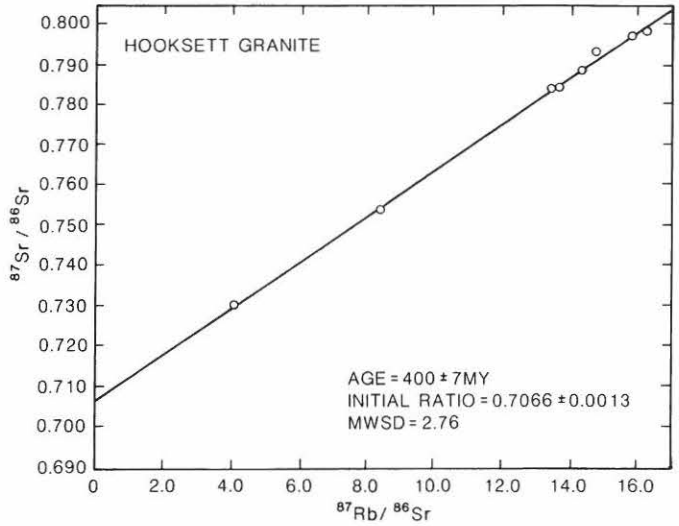
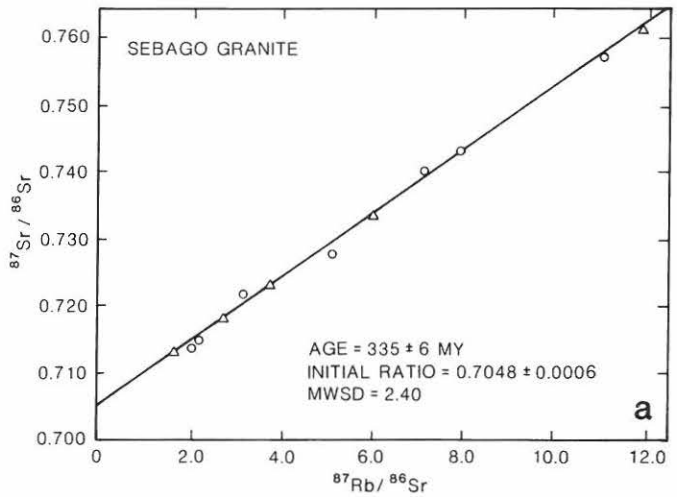


Figure 6. Rb-Sr whole rock isochron for the Hooksett granite.



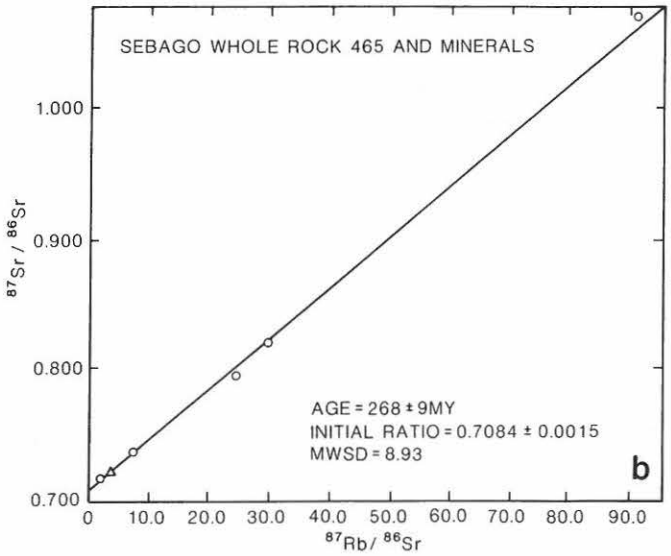


Figure 7. Rb-Sr whole rock isochron for the Sebago granite. (a) Reported in Hayward and Gaudette (1984) (*); this study (+). (b) Sample UNH 465; whole rock (*) and minerals (+).

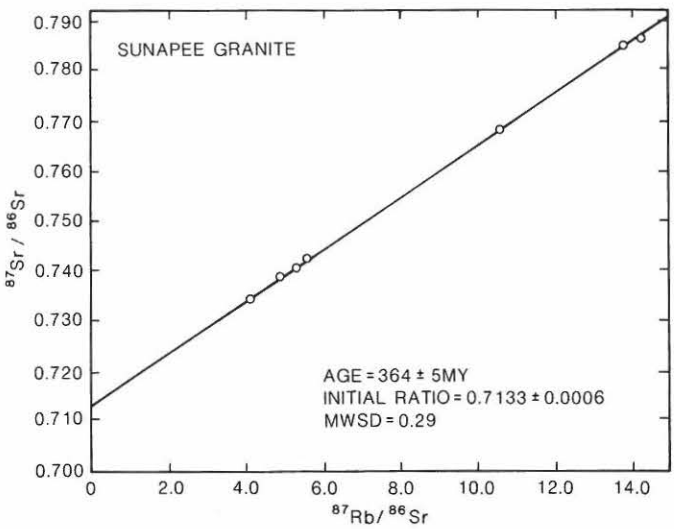


Figure 8. Rb-Sr whole rock isochron for the Sunapee granite.

TABLE 3. OXYGEN ISOTOPE ANALYSES.

Pluton	Sample #	delta ^{18}O (‰)	Calculated Sr initial ratio
Chelmsford	JAH8521	+12.5	0.7099
	JAH8524	+12.6	0.7103
	JAH8525	+11.6	0.7089
	JAH8526	+10.5	0.7108
Fitzwilliam	NH/MND 4	+ 8.6	0.7084
	NH/MND 7	+11.1	0.7085
	NH/MND 8	+10.6	0.7093
	NH/MND10	+10.0	0.7084
Sebago	JAH85161	+10.4	0.7074
	JAH85164	+11.1	0.7069
	JAH85165	+11.1	0.7078
	JAH85168	+ 7.5	0.7044

Sebago granite yields relatively high delta ^{18}O values despite its low whole rock initial $^{87}Sr/^{86}Sr$ ratio.

DISCUSSION

Whole Rock Ages

The six bodies studied can be characterized as generally undeformed, with only local areas where syn- to post-emplacement effects are apparent. Thus the Rb-Sr whole rock ages obtained probably represent crystallization ages and are in agreement with previously published data for members of the New Hampshire Plutonic Series and correlated plutons in southwestern Maine and eastern Vermont (Naylor, 1971; Lyons and Livingston, 1977; Gaudette et al., 1982; Aleinikoff et al., 1985).

The whole rock ages obtained herein for the Chelmsford (399±6 Ma), Fitzwilliam (383±6 Ma), Hooksett (400±7 Ma), and Sunapee (364±5 Ma) plutons are Devonian ages, while the ages obtained for the Effingham (319±13 Ma) and Sebago (335±6 Ma) plutons, are Mississippian (Palmer, 1983). In light of the data presented in this study and in those of Naylor (1971), Lyons and Livingston (1977), Aleinikoff et al. (1979), Gaudette et al. (1982), and Aleinikoff et al. (1985), the Concord intrusive suite appears to have been intruded in northern New England over a length of time that spans both the Devonian and Carboniferous periods. As suggested by Gilluly (1973) and discussed by Lyons et al. (1982), tectonic events and their associated magmatism may not be distinct episodes. Thus, members of the Concord intrusive suite may have been generated in response to tectonic activity that was manifested through the Devonian and Carboniferous.

Mineral Isochrons

The mineral ages obtained for the Effingham (263±4 Ma), Sebago (268±9 Ma), and Fitzwilliam (326±15 Ma and 267±12 Ma) granites support the notion of an event of Alleghenian affinity as seen in the southern Appalachians (e.g., Sinha and Zeitz, 1982; Fullagar and Butler, 1979), Rhode Island (e.g., Kocis et al., 1978), northern New England (e.g., Zartman et al., 1970; Dallmeyer and Van Breeman, 1981), and throughout western New England (e.g., Michard-Vitrac et al., 1980; Van Breeman et al., 1983). Those of the Effingham and Sebago granites and the younger of the ages obtained from the Fitzwilliam body are typical of the mineral ages obtained through Rb-Sr and K-Ar analyses for many plutons in the New England region (e.g., Faul et al., 1963; Zartman et al., 1970; Dallmeyer and Van Breeman, 1981). The extent of this event in northern New England had been thought to be limited to reset ages associated with erosion and uplift (e.g., Zartman et al., 1970). However, four two-mica granite plutons in northern New England have now been identified as Carboniferous in age (Effingham, Lyman, Milford, and Sebago). The significance of the fact that these four plutons were emplaced within the central portion of the Kearsarge-central Maine synclinorium will be addressed in greater detail below.

Oxygen Isotopic Data

In Figure 9, delta ^{18}O values are plotted against calculated initial Sr ratios for the three plutons. The fields of several "typical" source rock types are also delimited. These fields are after Taylor (1980). As can be seen in this diagram, a distinction can be made between the field of the Sebago granite and those of the Chelmsford and Fitzwilliam granites. The data arrays of the Chelmsford and Fitzwilliam granites lie along a trend that extends from an altered oceanic basaltic source toward a continental-derived sedimentary contaminant. The three plutons appear to have been produced through the anatexis of an altered oceanic basaltic end-member with isotopically distinct contaminant input of varied amounts. The O and Sr isotopic data of the Chelmsford and Fitzwilliam granites reflect the incorpora-

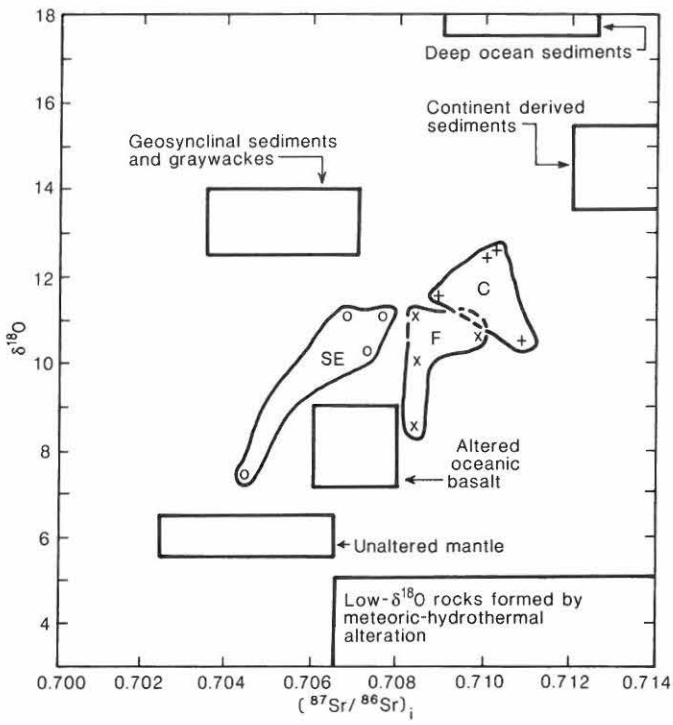


Figure 9. Graph of delta ^{18}O vs calculated initial $^{87}Sr/^{86}Sr$ ratios for the Chelmsford (C), Fitzwilliam (F), and Sebago (SE) granites.

tion of an ^{18}O- and ^{87}Sr-enriched contaminant. The O and Sr isotopic data of the Sebago granite, however, suggest the importance of a component such as a geosynclinal sediment in the genesis of this pluton. It appears, therefore, that the source of the Sebago granite in particular is distinct in isotopic character from that of the Chelmsford and Fitzwilliam granites.

Two-mica Granites

The generation of S- and I-type granites has been the subject of much recent petrological and geochemical study, discussion, and debate (e.g., White et al., 1986; Miller, 1985, 1986; Dickson, 1986; Pitcher, 1979). Granites, in general, have proven to be excellent reservoirs of information with regard to the nature of their source material (e.g., Ayuso, 1986; Andrew et al., 1983; James, 1981), but the S- and I-type classification has proved, in practice, to be unsatisfactory in consideration of two-mica granites in northern New England. Petrographically, the granites of this study have some characteristics like those reported for S-type granites in the Lachlan fold belt by Chappell and White (1974). Muscovite and biotite, as indicated previously, are present as primary minerals and thus, in terms of mineralogy alone these plutons could be considered S-type granites. Major element and CIPW analyses demonstrate that most of the granites are corundum-normative and have fairly high, restricted SiO_2 percentages. Most significantly, however, the Sr isotopic data for some of the two-mica granites of this study are inconsistent with an S-type designation. In fact, only three of the granites, the Chelmsford, Fitzwilliam, and Sunapee, have initial $^{87}Sr/^{86}Sr$ ratios that are greater than 0.708, the lower limit reported for S-type granites in the Lachlan fold belt of Australia (Chappell and White, 1974).

A clear geographic trend in initial $^{87}Sr/^{86}Sr$ ratios is exhibited by the plutons. The relatively higher ratios (>0.7085) are derived from the westernmost plutons (Fitzwilliam and Sunapee) and the southeasternmost (Chelmsford) of the plutons, while the lowest initial $^{87}Sr/^{86}Sr$ ratios (≤0.7066) are derived from those plutons emplaced in the central portion of the Kearsarge-central Maine synclinorium (Effingham, Hooksett, Lyman, and Sebago). The initial $^{87}Sr/^{86}Sr$ ratios of the latter group cluster about a mean of 0.7052 (0.7048, 0.7066, 0.7044, and 0.7048), while the westernmost plutons display a greater degree of diversity (0.7090-0.7133). Ayuso (1986) reported a similar tendency toward isotopic uniformity in Pb compositions derived from feldspars of granites emplaced in the Kearsarge-central Maine synclinorium in Maine versus the wider range in composition yielded by the feldspars from granites of the Connecticut Valley-Gaspé synclinorium and coastal lithotectonic belt of Maine.

Previously published Rb-Sr whole rock studies of granites of the Kearsarge-central Maine synclinorium of Maine generally support the trend seen in initial $^{87}Sr/^{86}Sr$ ratios in New Hampshire and Maine. The only pluton of the Kearsarge-central Maine synclinorium region in Maine from which an initial Sr ratio >0.0708 has been obtained to date is the Katahdin granite (0.7083±0.0006) (Andrew et al., 1983) and this pluton, like the Fitzwilliam and Sunapee granites of New Hampshire, is located in the western portion of the synclinorium. Other plutons of the Kearsarge-central Maine synclinorium range in initial Sr ratio from a low of 0.7041±0.0003 (Passadumkeag River member of the Bottle Lake Complex; Ayuso et al., 1984), to a high of 0.7066±0.0028 (Mattamiscontis member of the Seboeis Complex; Andrew et al., 1983). These plutons and those having initial Sr ratios within this range were emplaced in the central portion of the Kearsarge-central Maine synclinorium in Maine and along the northwestern edge of the Norumbega-Nonesuch River-Flint Hill fault. These data are in apparent agreement with the geographic trend in initial Sr ratio determined in this study.

Rb-Sr whole rock analyses of granites emplaced southeast of the Nonesuch River-Norumbega fault are not as clear-cut in this regard, however. Knight and Gaudette (1987) have determined a Rb-Sr whole rock age of 367±4 Ma and an initial $^{87}Sr/^{86}Sr$ ratio of 0.7123±0.0008 for the Waldoboro granite of the coastal lithotectonic belt of Maine. This ratio is greater than 0.708, as is that determined for the Chelmsford pluton which lies approximately along strike to the southwest. Other initial $^{87}Sr/^{86}Sr$ data such as those obtained from the Lead Mountain granite (0.7041; Andrew et al., 1983), the Lucerne granite (0.7077; Andrew et al., 1983), the Webhannet granite

(0.7046±0.0007; Gaudette et al., 1982), and the Biddeford granite (0.7067±0.0008; Gaudette et al., 1982), however, do not support the suggested trend of initial $^{87}Sr/^{86}Sr$ ratios >0.7085 for the region southeast of the Nonesuch River-Norumbega fault proposed in this study. It may be that the proposed coastal lithotectonic belt can be characterized as a zone of variable initial $^{87}Sr/^{86}Sr$ ratio, implying a more heterogeneous basement(s) underlying this zone, in contrast to the more homogeneous basement that appears to underlie the central portion of the Kearsarge-central Maine synclinorium. Alternatively, the variation in initial $^{87}Sr/^{86}Sr$ ratios may be the result of differential contamination in the generation or emplacement of the granites.

The initial $^{87}Sr/^{86}Sr$ ratio and O data suggest that the tectonic zone southeast of the Nonesuch River-Norumbega fault and that immediately east of the Bronson Hill anticlinorium are terranes distinct in isotopic character from the central portion of the Kearsarge-central Maine synclinorium. Interpretation of the initial Sr ratio data allows that the Chelmsford, Fitzwilliam, and Sunapee granite source(s) be characterized as enriched in radiogenic ^{87}Sr, hence having continental crust affinities. The oxygen isotope data available for the Fitzwilliam (+8.6 to +11.1) and the Chelmsford (+10.5 to +12.6) granites support this interpretation. The plutons of the central part of the synclinorium, however, do not appear to have been derived from a source enriched in radiogenic ^{87}Sr. The low initial $^{87}Sr/^{86}Sr$ ratios (<0.7066) suggest rather that the source(s) of these granites should be characterized as having a relatively low $^{87}Sr/^{86}Sr$ ratio. Interestingly, the oxygen isotopic data from the Sebago granite (+7.5 to +11.1) imply at least the incorporation of some continental crust in the magma. The highest initial $^{87}Sr/^{86}Sr$ ratio of the central synclinorium group of plutons is that of the Hooksett granite, the oldest of these plutons. It is possible therefore that the lowest initial $^{87}Sr/^{86}Sr$ ratios, those of the Effingham, Lyman, and Sebago granites, have some causal relationship to the Carboniferous timing of their emplacement. Such a relationship might result from the generation of the Carboniferous magmas from a more Rb-depleted source and/or different crustal level source.

Finally, trace element, rare earth element, more major element, and other isotopic data (Pb, Sm-Nd, and more O) for the two-mica plutons of New England could greatly enhance the characterization of source(s) based on the present data set. Assimilation-fractional crystallization models of these combined data should permit refinement of the basement terrane characteristics that have been identified to date and of the the tectonic models that have been developed for the northern Appalachians.

CONCLUSIONS

(1) Rb-Sr analyses of six two-mica granite plutons in northern New England yield Devonian (Chelmsford, Fitzwilliam, Hooksett, and Sunapee) and Carboniferous (Effingham and Sebago) crystallization ages.

(2) Rb-Sr mineral ages of the Effingham, Fitzwilliam, and Sebago plutons confirm a Carboniferous to Permian resetting event in northern New England as observed by others (e.g., Faul et al., 1963; Zartman et al., 1970).

(3) Plutons emplaced in the western and eastern portions of the Kearsarge-central Maine synclinorium yield high initial $^{87}Sr/^{86}Sr$ ratios (0.7090 to 0.7133), whereas those emplaced in the central portion of the Kearsarge-central Maine synclinorium yield low initial Sr ratios (≤0.7066). Plutons in the central portion of the synclinorium exhibit much less variability in initial $^{87}Sr/^{86}Sr$ ratio than do those emplaced in the western and southeastern portions of the synclinorium.

(4) The available oxygen isotope data suggests that the Chelmsford, Fitzwilliam, and Sebago granites were generated by the anatexis of an altered oceanic basaltic source with isotopically distinct contaminant input of varied volume.

ACKNOWLEDGMENTS

Parts of this research were funded by two grants from the Geological Society of America, #2999-82 and #3440-85, and a research grant from the Central University Research Fund of the University of New Hampshire. Oxygen isotope analyses were provided by Geochron Laboratories of Cambridge, Massachusetts under their Stable Isotope Ratio Analyses competition in 1985. The assistance of these organizations and that of the University of New Hampshire in providing assistantships and fellowships to the author are gratefully recognized. I am also indebted to H. E. Gaudette, W. J. Olszewski, E. L. Boudette, D. B. Stewart, D. A. Gust, C. J. Schneer, J. Laird, and W. A. Bothner, who supplied guidance, support, and encouragement throughout the course of this study. Thoughtful reviews by Calvin Miller, Malcolm Hill, and David Stewart resulted in substantial improvement of the manuscript.

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APPENDIX I. MAJOR ELEMENTS AND CIPW NORMS (WEIGHT PERCENT).

	SU 9†	SU 10	SU 11	SU 12	SU 13	CG 1‡	NLG 1	HILG1
SiO_2	74.70	72.60	73.80	73.60	75.10	74.47	70.20	72.47
TiO_2	0.07	0.15	0.07	0.03	0.01	–	0.44	–
Al_2O_3	14.70	15.20	15.20	15.30	14.90	14.15	16.00	16.17
Fe_2O_3	0.28	0.54	0.29	0.22	0.11	1.16	1.10	–
FeO	0.64	0.86	0.68	0.60	0.24	1.21	1.00	0.41
MnO	0.05	0.03	0.03	0.02	0.04	–	0.02	0.39
MgO	0.09	0.29	0.15	0.11	0.04	0.63	0.66	0.14
CaO	0.48	0.94	0.65	0.57	0.24	1.70	1.60	1.65
Na_2O	4.20	3.00	3.50	3.90	4.30	1.97	4.00	3.43
K_2O	3.90	4.70	4.70	4.10	4.40	4.14	4.60	4.83
H_2O^+	1.00	1.20	0.72	1.10	0.72	0.20	0.43	–
H_2O^-	0.14	0.16	0.14	0.02	0.02	0.06	0.43	–
P_2O_5	0.15	0.18	0.19	0.16	0.13	–	0.23	–
TOTAL	100.40	99.85	100.12	99.73	100.25	99.69	100.28	99.49
Ap	0.35	0.42	0.44	0.37	0.30	–	0.53	–
Il	0.13	0.28	0.13	0.06	0.02	–	0.84	–
Mt	0.41	0.78	0.42	0.32	0.16	1.68	1.59	–
Or	23.05	27.78	27.78	24.23	26.00	24.47	27.18	28.54
Ab	35.54	25.39	29.62	33.00	36.39	16.67	33.85	29.02
An	1.40	3.49	1.98	1.78	0.34	8.43	6.44	8.19
Hy	1.15	1.66	1.32	1.18	0.51	2.83	1.88	1.83
C	3.06	3.90	3.63	3.79	2.94	3.34	2.08	2.30
Q	34.18	34.79	33.94	33.87	32.85	42.01	25.45	29.61
TOTAL	99.27	98.49	99.26	98.60	99.51	99.43	99.84	99.49
Molar A/CNK	1.14	1.13	1.15	1.19	1.17	1.02	0.92	0.96

†Courtesy E. L. Boudette, 1983 (U. S. G. S. analyses); SU 9-13 = Sunapee granite

‡M. P. Billings and J. R. Wilson, 1964; CG 1 = Concord granite; NLG 1 = Newfound Lake granite; HILG 1 = Hillsboro granite; WMG 1 = Whiteface Mountain granite

APPENDIX I. (CONTINUED)

	WMG1	E290*	E291	E460	LY-1	LY-2	LY-3	LY-4
SiO_2	69.76	71.77	71.25	71.71	71.82	70.32	70.49	71.39
TiO_2	0.36	0.29	0.28	0.32	0.16	0.27	0.28	0.19
Al_2O_3	18.22	15.00	15.52	14.66	15.63	16.14	16.20	15.95
Fe_2O_3	0.25	2.18	1.97	2.08	2.10	2.64	2.78	2.44
FeO	1.59	–	–	–	–	–	–	–
MnO	0.25	0.05	0.04	0.04	0.03	0.04	0.04	0.05
MgO	0.40	0.41	0.37	0.42	0.54	0.62	0.67	0.46
CaO	2.68	1.33	1.38	1.62	1.41	1.21	1.18	0.79
Na_2O	4.06	4.17	4.24	4.49	3.90	3.52	3.84	3.44
K_2O	2.06	4.79	4.83	4.85	4.39	5.15	4.44	5.23
H_2O^+	0.50	–	–	–	–	–	–	–
H_2O^-	0.15	–	–	–	–	–	–	–
P_2O_5	0.50	0.01	0.14	0.01	0.02	0.08	0.08	0.06
TOTAL	100.03	99.93	100.02	100.00	100.00	99.99	100.00	100.00
Ap	1.16	0.02	0.32	0.02	0.05	0.19	0.19	0.14
Il	0.68	0.11	0.09	0.09	0.06	0.09	0.09	0.11
Mt	0.36	–	–	–	–	–	–	–
Or	12.17	28.31	28.54	28.66	25.94	30.44	26.24	30.91
Ab	34.35	35.29	35.88	37.99	33.00	29.79	32.49	29.11
An	10.03	6.53	5.93	5.52	6.86	5.48	5.33	3.53
Di	–	–	–	1.16	–	–	–	–
Ru	–	0.23	0.23	–	0.13	0.22	0.23	0.13
He	–	2.18	1.97	2.08	2.10	2.64	2.78	2.44
Ti	–	–	–	0.67	–	–	–	–
Hy	3.58	1.02	0.92	0.51	1.34	1.54	1.67	1.15
C	5.64	0.56	1.14	–	1.95	2.77	3.12	3.34
Q	32.06	25.75	24.99	23.49	28.56	26.84	27.86	29.15
TOTAL	100.03	100.00	100.01	100.19	99.99	100.00	100.00	100.01
Molar A/CNK	0.98	0.99	1.01	0.93	1.10	1.10	1.16	1.13

	LY-5	UNH464	UNH465	UNH466	UNH469
SiO_2	68.89	71.31	72.55	72.55	72.29
TiO_2	0.38	0.20	0.20	0.15	0.19
Al_2O_3	16.11	14.86	14.63	15.27	15.15
Fe_2O_3	3.21	2.54	2.04	1.29	1.81
FeO	–	–	–	–	–
MnO	0.04	0.04	0.06	0.03	0.05
MgO	1.12	0.43	0.41	0.23	0.36
CaO	1.54	1.54	1.50	1.10	1.35
Na_2O	3.62	3.25	3.64	3.63	3.83
K_2O	4.97	5.84	4.90	5.73	4.95
H_2O^+	–	–	–	–	–
H_2O^-	–	–	–	–	–
P_2O_5	0.12	0.01	0.06	0.01	0.01
TOTAL	100.00	100.02	99.99	99.99	99.99
Ap	0.28	0.02	0.14	0.02	0.02
Il	0.09	0.09	0.13	0.06	0.11
Or	29.37	34.51	28.96	33.86	29.25
Ab	30.63	27.50	30.80	30.72	32.41
An	6.86	7.57	7.05	5.39	6.63
Hy	2.79	1.07	1.02	0.57	0.90
Ru	0.33	0.15	0.13	0.12	0.13
He	3.21	2.54	2.04	1.29	1.81
C	2.26	0.42	0.76	1.12	1.06
Q	24.18	26.14	28.97	26.83	27.66
TOTAL	100.00	100.01	100.00	99.98	99.98
Molar A/CNK	1.08	0.94	0.99	0.98	1.01

*Courtesy H. E. Gaudette and J. Wilband, 1984; E 290, 291 and 460 = Effingham granite; LY-1 - 5 = Lyman granite; UNH 464, 466, 469 = Sebago granite

APPENDIX II. LATITUDE AND LONGITUDE COORDINATES FOR SAMPLES ANALYZED IN THIS STUDY.

Sample #	Latitude	Longitude	Pluton
JAH8520	42 38 07	71 24 28	Chelmsford
JAH8521	42 37 57	71 25 03	
JAH8522	42 38 51	71 25 13	
JAH8523	42 38 48	71 25 17	
JAH8524	42 38 48	71 25 17	
JAH8525	42 39 00	71 25 54	
JAH8526	42 35 59	71 28 22	
JAH8527	42 35 59	71 28 23	
JAH8528	42 36 00	71 28 24	
JAH8529	42 36 01	71 28 23	
JAH8530	42 36 01	71 28 23	
UNH 289	43 44 28	71 07 30	Effingham
UNH 290	43 44 28	71 07 30	
UNH 291	43 44 28	71 07 30	
UNH 292	43 44 28	71 07 30	
UNH 458	43 44 27	71 07 33	
UNH 459	43 44 27	71 07 33	
UNH 460	43 44 27	71 07 33	
NH/MND 1	42 45 16	72 08 41	Fitzwilliam
NH/MND 2	42 45 16	72 08 41	
NH/MND 3	42 45 16	72 08 41	
NH/MND 4	42 45 31	72 08 56	
NH/MND 5	42 45 31	72 08 56	
NH/MND 6	42 47 56	72 09 10	
NH/MND 7	42 47 56	72 09 10	
NH/MND 8	42 46 38	72 07 59	
NH/MND 9	42 45 47	72 09 42	
NH/MND 10	42 45 47	72 09 42	
NH/SUNC 1	43 02 48	71 28 20	Hooksett
NH/SUNC 2	43 02 48	71 28 20	
NH/SUNC 3	43 02 48	71 28 20	
NH/SUNC 4	43 03 14	71 29 14	
NH/SUNC 5	43 03 14	71 29 14	
NH/SUNC 6	43 03 14	71 29 14	
NH/SUNC 7	43 03 25	71 29 01	
NH/SUNC 8	43 03 25	71 29 01	
UNH 464	44 04 00	70 42 06	Sebago
UNH 465	44 01 19	70 42 51	
UNH 466	44 48 54	70 35 00	
UNH 467	44 00 00	70 45 43	
UNH 469	44 01 19	70 42 51	
JAH85160	43 55 48	70 31 00	
JAH85161	43 56 29	70 31 54	
JAH85162	43 54 03	70 19 43	
JAH85163	43 50 18	70 35 57	
JAH85164	43 50 14	70 36 03	
JAH85165	43 50 35	70 36 17	
JAH85166	43 51 12	70 37 11	
JAH85167	43 51 32	70 35 57	
JAH85168	44 03 18	70 41 54	
JAH85169	44 02 26	70 41 54	
NH/SUNP 1	43 26 25	72 04 30	Sunapee
NH/SUNP 2	43 26 25	72 04 30	
NH/SUNP 3	43 26 25	72 04 30	
NH/SUNP 4	43 24 52	72 02 58	
NH/SUNP 5	43 24 53	72 02 58	
NH/SUNP 6	43 21 35	71 59 46	
NH/SUNP 7	43 21 32	71 59 46	
NH/SUNP 8	43 21 31	71 59 47	

Maine Geological Survey
Studies in Maine Geology: Volume 3
1989

The Commingling of Diverse Magma Types in the Flagstaff Lake Igneous Complex

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ABSTRACT

The Flagstaff Lake Igneous Complex is a member of an array of Acadian plutons which extend from the Katahdin batholith in the northeast to New Hampshire in the southwest. It is characterized by four major rock types: (1) gabbro, (2) granite, (3) garnet tonalite, and (4) trondhjemite. Field, petrographic, and geochemical evidence support the hypothesis that these diverse rocks existed as contemporaneous liquids, and that extensive mixing occurred at the time of emplacement.

The dominant mafic mineral pairs in the gabbros are olivine-clinopyroxene, two pyroxenes, and pyroxene-hornblende. The high Fe, low Ni and Cr contents of most of the gabbro sampled imply that it underwent substantial fractional crystallization prior to and during emplacement. The granitic rocks of the Flagstaff Lake Igneous Complex are heterogeneous, ranging from two-mica granites with equal amounts (~30%) of orthoclase, plagioclase, and quartz, to mafic diorites. These K-feldspar-bearing silicic rocks are exposed over ~40% of the areal extent of the complex.

Trondhjemite containing biotite, plagioclase, cordierite, and quartz, with minor muscovite and K-feldspar, is located along the margins of the intrusion, particularly in areas where the contact between the gabbro and granite intersects aluminous metapelitic wall rocks.

The garnet tonalite is generally located in sections of the complex where the wall rocks are sulfidic, graphitic metasedimentary rocks. The tonalite is characterized by up to 60% almandine garnet, and oscillatory-zoned plagioclase, biotite, quartz, apatite, and zircon. The bodies of garnet tonalite are commonly, but not exclusively, located near the contacts between gabbro and country rock, and truncate the local strike of the country rocks.

Irregular, fine-grained mafic enclaves with biotite-rich rims are common throughout the granite, trondhjemite, and tonalite. This is especially true near the edge of the intrusion, in contrast to many occurrences of gabbro "pillows" (Wiebe, 1973; Taylor et al., 1980; Furman and Spera, 1985) where the mafic enclaves are generally restricted to dike-like chains. Other observations, such as phenocryst trains, schlieren, and large-scale chemical and mineralogical heterogeneity of the igneous body are common along the contacts between the four observed rock types.

INTRODUCTION AND REGIONAL GEOLOGY

The Flagstaff Lake Igneous Complex is one of a series of Acadian plutons that extends across western Maine from New Hampshire northeast to the Katahdin batholith in northern Maine. Along this array of plutons, the depth of emplacement increases from northeast (1-3 km at Katahdin -- Loiselle et al., 1983) to southwest (7-10 km at Pierce Pond -- Lyttle, 1976; and Flagstaff Lake -- Boudette, 1973, 1978). A crosscutting relationship with Acadian isoclinal folds, and isotopic dating of the nearby slightly younger Lexington pluton (Gaudette and Boone, 1985) fix the age of the complex at approximately 400 Ma, after the major tectonic events of the Acadian orogeny.

The Flagstaff Lake Igneous Complex is situated at the southern end of the Lobster Mountain anticlinorium, flanked on the southeast and northwest by the Kearsarge-central Maine and Moose River synclinoria, respectively (Osberg et al., 1985). Major metamorphic units surrounding the complex are the lower Paleozoic Dead River, Carrabassett and Seboomook Formations around the northeastern portion of the complex (Vehrs, 1975), and the Ordovician Quimby and Greenvale Cove Formations around the southwestern part. Regional Acadian chlorite-grade metamorphism is overprinted by a narrow (2 km) contact aureole. Geobarometry on minerals from the contact zone yields pressures in the range of 2-3 kbar (Vehrs, 1975; Lyttle, 1976). The igneous rocks show little evidence of deformation or metamorphism.

The complex can be divided into four main rock types: gabbro, granite, trondhjemite, and garnet tonalite. Gabbroic rocks are concentrated in the northern and southern ends of the complex, covering approximately 60% of the areal extent (Fig. 1), while granitic rocks and trondhjemites dominate the middle third.

Contacts between gabbroic and granitic rocks are gradational and irregular. The granite contains mafic enclaves and gabbro-cored biotite schlieren, while the gabbros contain large (2 cm) xenocrysts of sodic plagioclase feldspar (An_{25-40}) and an overall increase in amphibole and biotite near the contact. Trondhjemite is found exclusively where the gabbro-granite contact intersects the edge of the intrusion, in contact with aluminous metamorphic rocks of the Dead River, Seboomook and Carrabassett Formations.

The garnet tonalites are also generally located at gabbro-country rock contacts, particularly at the extreme northern (Pierce Pond) and southern ends of the complex. These are areas where the gabbro is in contact with Fe-rich, sulfidic, graphitic metasiltstones and metagraywackes of the Quimby and lower Dead River Formations.

From Figure 1, one can see that the complex is crudely zoned. There are many processes which have been called upon to explain individual cases of zonation such as marginal accretion/assimilation (Holmes, 1932; DePaolo, 1981; McBirney, 1980), fractional crystallization and accumulation, autointrusion by a magma of different composition (Ayuso, 1984), and thermo-gravitational diffusion (Hildreth, 1981). Each of these processes place chemical and textural signatures on the suite of rocks produced. In this paper we shall present field, petrographic, and geochemical evidence that shows several of these processes were active in the formation of the present form of the Flagstaff Lake Igneous Complex.

FIELD RELATIONS

In general, the Flagstaff Lake Igneous Complex is concordant with the northeast-southwest regional strike of the metamorphosed sedimentary rocks (Moench, 1971). However, detailed mapping shows that parts of the intrusion truncate the strike of the country rock (Boone, 1973). In those areas, the shape of the contacts appear irregular on a scale of tens to hundreds of meters. An example from the southern end of the complex is illustrated in Figure 2. The scale of the irregularity of the contact varies greatly, from none, to that shown in Figure 2. However, the largest bodies of both tonalite (Fig. 2) and trondhjemite (Boone, 1973) are found in areas with the most convolute contacts.

Within a few tens of meters of the contact, the metamorphic rocks show progressive development of migmatites together with veins of granite or trondhjemite. The proportion of leucosome to melanosome, and the incidence of veins, increases steadily as one approaches the complex. At approximately 40-60% leucosome, the rock becomes a breccia and grades into the igneous body (Fig. 3a,e). Stoped blocks of metamorphic rocks, generally less than 4 m in size, are common within a few hundred meters of the contact. However, they are almost totally absent from the center of the complex. In addition, unlike many of the surrounding intrusions (Boone et al., 1970; Vehrs, 1975; Lyttle, 1976) the Flagstaff Lake Igneous Complex contains few large roof pendants or large stoped blocks.

Mafic enclaves are pervasive in the complex, particularly near (100-200 m) where the gabbro-granite, gabbro-

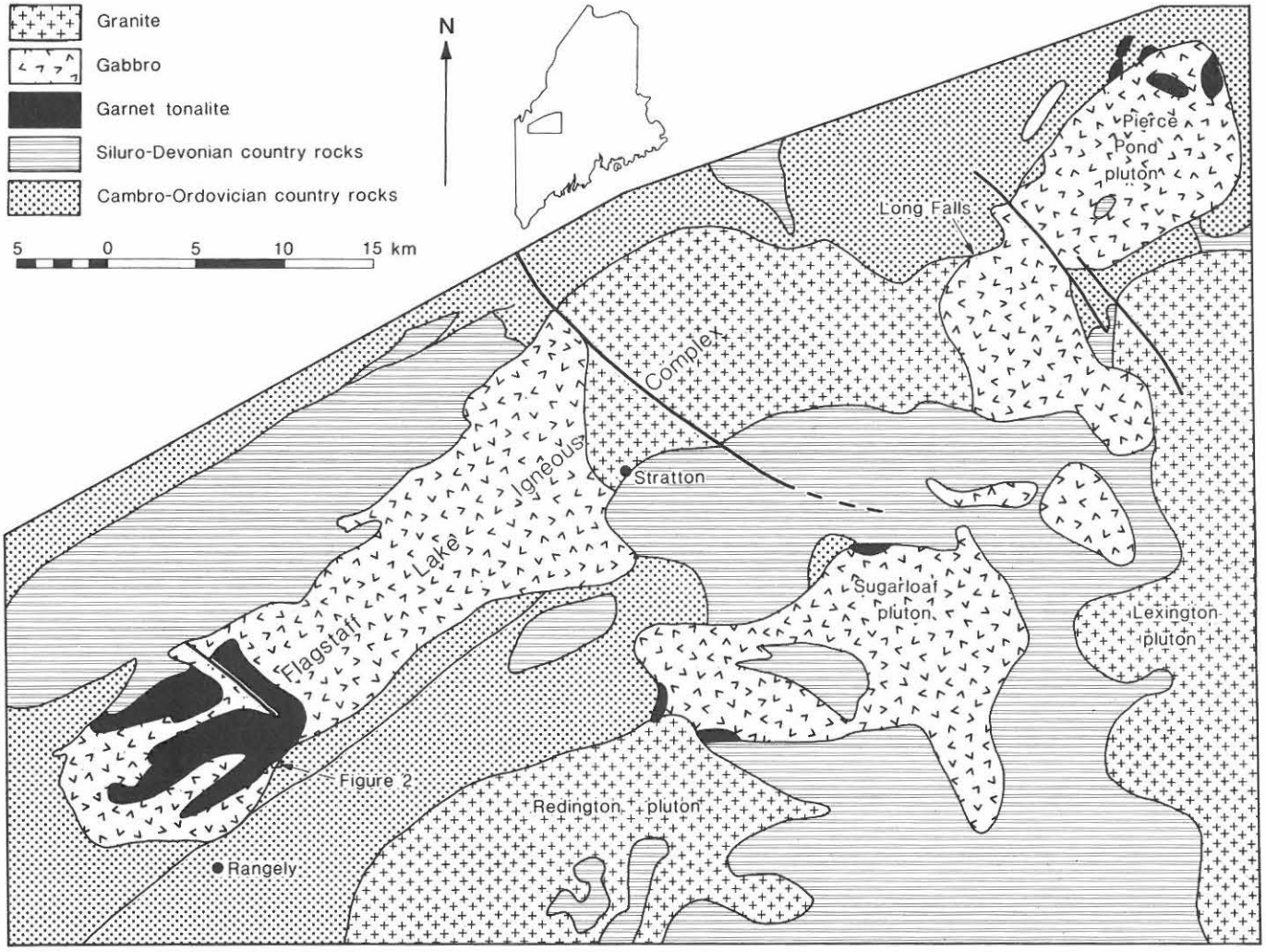

Figure 1. Location and geologic maps of the Flagstaff Lake Igneous Complex (modified after Moench et al., 1982).

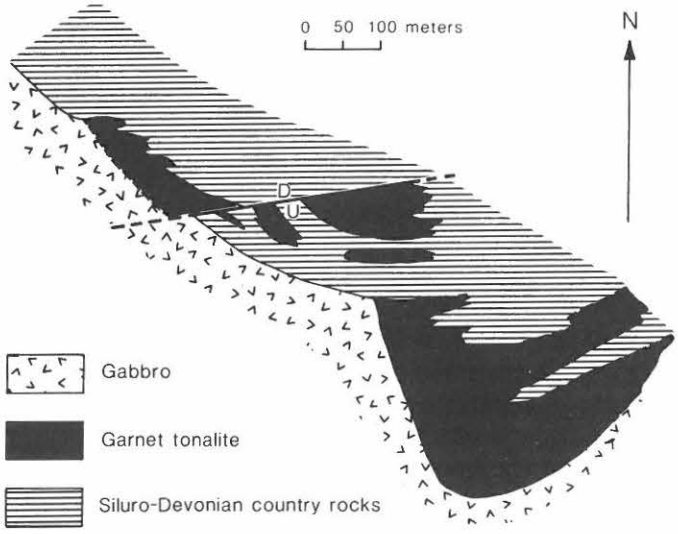

Figure 2. Map of the details of the field relationship of the gabbro, garnet tonalite, and metamorphic country rock.

Figure 3. Field evidence of magma mixing and partial melting. (a) Migmatites and contact breccias, Dead River Formation at Long Falls, 2 m from metasediment-trondhjemite contact.

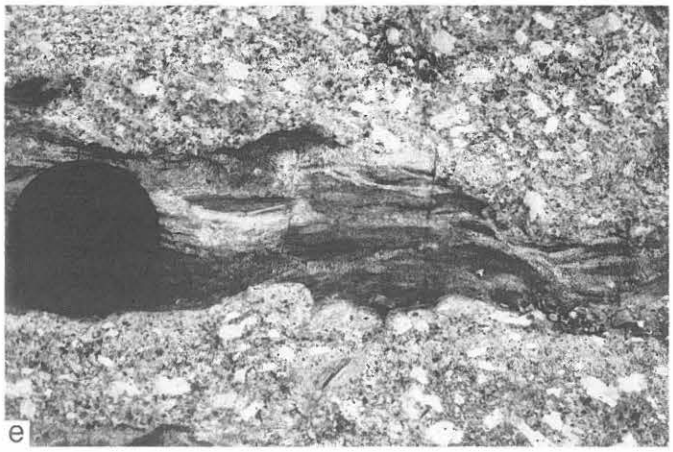

Figure 3 (continued). Field evidence of magma mixing and partial melting. (b) Xenoliths (quartz-cordierite hornfels) in garnet tonalite near Rangeley. (c) Biotite-rimmed gabbro inclusions, Long Falls. (d) Mixing interface between quartz monzonite porphyry and fine-grained diorite showing plagioclase phenocryst trains in diorite. (e) Biotite schlieren (metasediment derived) in quartz monzonite porphyry, near Stratton. (f) Biotite-rimmed gabbro inclusion in garnet tonalite, Pierce Pond. (g) Cuspate gabbro enclave-granite contact, Long Falls.

trondhjemite, or gabbro-tonalite contacts intersect the edge of the intrusion (e.g. Long Falls - Fig. 1). These enclaves have irregular morphologies (Fig. 3c,f,g), are fine grained, and have biotite-rich rims. Angular blocks and breccia, evidence of brittle behavior, are rarely observed away from the immediate vicinity (<10 m) of the edge of the intrusion. The thickness of the biotite rims increases roughly as a function of the distance from the edge of the complex. The rims of the enclaves are often crenulate (Fig. 3g). As one proceeds toward the center of the complex, the mafic enclaves become more ellipsoidal in shape and dioritic in composition (Fig. 3d). Eventually, in the center of the intrusion, the gabbro-granite and gabbro-tonalite contacts become gradational. There, the only remaining evidence of the mafic enclaves is biotite schlieren.

In other reported cases of mafic enclaves in granitic bodies, the enclaves are largely restricted to dike-like clusters (Taylor et al., 1981; Furman and Spera, 1985). In most recent investigations this has been attributed to the injection of gabbro into a granitic body which was largely crystalline. In the Flagstaff Lake Igneous Complex, the mafic enclaves commonly occur in a wide band along the boundary separating the granitic and gabbroic sections of the intrusion. This observation suggests that the enclaves may have originated in a boundary layer between two different magma types. The dip or orientation of this boundary is too irregular for any determination of its nature.

In the southern part of the complex, weakly metamorphosed sulfidic, graphitic, Fe-rich metasiltstones and metagraywackes are the dominant lithologies near occurrences of garnet tonalite. At Pierce Pond, in the far northern end of the intrusion, similar metamorphic rocks are present in the middle and lower sections of the Dead River Formation (Lyttle, 1976; Moench et al., 1982). These rocks are also associated with garnet tonalites (Boone, 1985). Randomly oriented blocks of metamorphic rocks and biotite-rimmed gabbro enclaves are present in both locations (Fig. 3b,c,f,g). Garnet tonalites do not occur in contact with either trondhjemite or granite. The garnet tonalite-gabbro contact ranges from sharp (4 m wide) to gradational (1 km wide). A full suite of intermediate rock types can be observed in the field, from garnet tonalites with 35-70% garnet to garnet-free gabbro. The garnet tonalite exhibits no foliation or mineral layering, on either a large or small scale.

ANALYTICAL METHODS

Bulk major and trace element chemistry of the samples from the Flagstaff Lake Igneous Complex was determined by XRF analysis using fused pellets and pressed powder preparations (Johnson, 1984). Standards included both USGS standards, BHVO-1, W-2, GSP-1, RGM-1, and a set of synthetic standards for Fe, Y, and Zr which were calibrated by ICP and XRF analysis at the USGS. Error on the major elements is approximately 2.2% relative and ±5% for the trace elements.

Mineral chemistry was determined by electron microprobe at the facilities of the Mineral Sciences Department, Smithsonian Institution (ARL SEMQ), and the Department of Engineering, University of Maryland (JEOL 840a). Precision is approximately 1% for the major components.

PETROGRAPHY AND MINERAL CHEMISTRY

The four major igneous rock types present in the Flagstaff Lake Igneous Complex can be readily distinguished on the basis of their mineralogy and modal proportions (Table 1). All rock types were named using the IUGS (Streckheisen, 1976) classification. The two-mica granite (Fig. 4a) is medium to coarse grained and contains approximately equal amounts of orthoclase, quartz, and plagioclase (An_{20-40}). In several locations, the granite is characterized by large (>5 cm) orthoclase phenocrysts. Biotite is the major mafic phase, although minor garnet and cordierite are also present (Boone, 1973). The modal proportion of muscovite is variable, but is always less than that of biotite. Diorites, both north and south of the central granite body, are finer grained than the granites (Fig. 4b) and contain local orthoclase with abundant biotite and amphibole (Table 1). These intermediate composition rocks are often heterogeneous, grading continuously from mafic diorite to granite in a space of 100 meters. They also contain large (>1 cm) resorbed sodic feldspar (An_{20-40}) and quartz crystals, many with biotite halos. Biotite is also present as discrete clusters in both the diorites and in some of the more silicic rocks.

The trondhjemites are characterized by a medium to fine grained texture, and plagioclase, quartz, and biotite with occasional trace amounts of cordierite, and garnet. Muscovite and K-feldspar are rare. Most of the plagioclase phenocrysts exhibit patchy zoning, particularly those in trondhjemite near the mafic enclaves. Samples near the metamorphic contact contain the most cordierite and garnet, both of which commonly show textural evidence of resorption and/or re-equilibration (rounded crystal forms, biotite reaction rims). A large fraction (>1/2) of the biotite in the trondhjemite is present in schlieren or clusters. It is therefore difficult to determine the modal percentage of biotite that is representative of the trondhjemite itself. Thus, in this case it is clear that the bulk composition of the trondhjemite does not represent a liquid composition.

The rims of the mafic enclaves are fine grained (Fig. 4c) but coarsen toward the center of the enclaves. As one proceeds inward from the rim of an enclave (Fig. 4c,d) the proportion of amphibole increases and biotite decreases. Texturally, biotite appears to be the product (Fig. 4e) of a reaction involving amphibole (and melt?). The centers of the larger enclaves (>3 m) are mineralogically similar to the olivine-pyroxene-amphibole-bearing gabbros found elsewhere in the complex, away from contacts with granitic rocks (Table 1). Many of the mafic enclaves contain feldspar phenocrysts, some of which have patchy zoning, resorbed rims, and sodic compositions (An_{25-40}) compared to the groundmass plagioclase compositions (An_{55-70}), and are thus probably xenocrystic.

Biotite schlieren fall into two mineralogic groups. The first is characterized by the almost total absence of any mineral but biotite. The biotite is medium grained (1-2 mm) and devoid of inclusions. The other type of schlieren contain quartz and cordierite in addition to biotite. These schlieren are finer grained (~200 microns) and biotite frequently contains abundant inclusions of monazite and zircon. Toward the center of the intrusion, the schlieren appear to break down into smaller and smaller units. The end products are clusters of biotite, 1-5 mm in diameter, made up of 5-50 individual crystals of biotite. Some of these clusters contain monazite and zircon inclusions, while others contain few or none. A few clusters are characterized by a single grain of biotite containing inclusions, surrounded by inclusion-free biotite. There is no consistent textural or color difference between the two biotite types. Some of these clusters contain inclusions characteristic of the biotite in the metamorphic rocks. Others are devoid of inclusions, as is the biotite associated with the mafic enclaves.

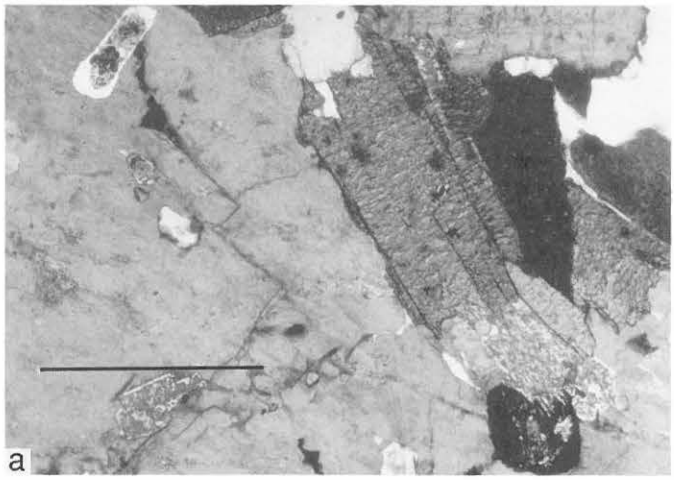

Figure 4. (a) Two-mica granite from Hedgehog Hill, 1 mi south of town of Stratton. Bar scale 1 mm.

TABLE 1: MODES OF REPRESENTATIVE SAMPLES FROM THE FLAGSTAFF LAKE IGNEOUS COMPLEX AND ITS AUREOLE

	86081 White granite	86020 Pink granite	EW-53 Diorite	86001A Hornfels	86087A Gabbro enclave	86087B Gabbro bt. rim	86087B Host granite
alk.fld	30-40	20-30	5	–	–	–	–
plag.	20-30	40-50	40	40-50	35-40	50	55
quartz	10-20	10-20	5	5	–	0	35
biotite	10	5-10	5	10	5-10	45	10
opaques	Tr	Tr	5	–	5-10	5	1
amphibole	–	–	50	40-50	35-40	–	–
muscovite	5	5	–	–	–	–	–
zircon	Tr	Tr	–	–	–	–	–
apatite	Tr	Tr	–	–	–	–	–
sphene	–	–	–	Tr	–	–	–
epidote	–	–	–	–	–	–	Tr
monazite	Tr	Tr	–	–	–	–	Tr
mean grain size	1-2mm pheno-3mm	1-3cm	2mm	400-600 m	200 m-1mm	50 m	1-2mm

	18 Garnet tonalite	146 Garnet tonalite	145 Garnet gabbro	226 Garnet gabbro	2 Gabbro	10 Gabbro	183 Hornfels incl.	193 Mafic incl.
plag	20	5	45	35	50	44	60	50
quartz	15	20	5	5	–	–	33	–
biotite	5	10	5	3	1	1	4	1
opaques	3	Tr	3	5	5	10	2	10
amphibole	–	–	–	–	15	–	–	–
garnet	45	65	5	10	–	–	–	–
pyroxene	–	Tr	35	40	–	45	–	35-40
cordierite	15	–	–	–	–	–	–	–
olivine	–	–	–	–	30	–	–	–
chlorite	Tr	–	–	–	Tr	–	–	–
apatite	Tr	Tr	Tr	Tr	–	–	Tr	–
zircon	Tr	Tr	Tr	Tr	–	–	–	–
mean grain size	0.8mm	1.9mm	0.1-2.5mm	1mm	1.3mm	1mm	0.3mm	0.5mm

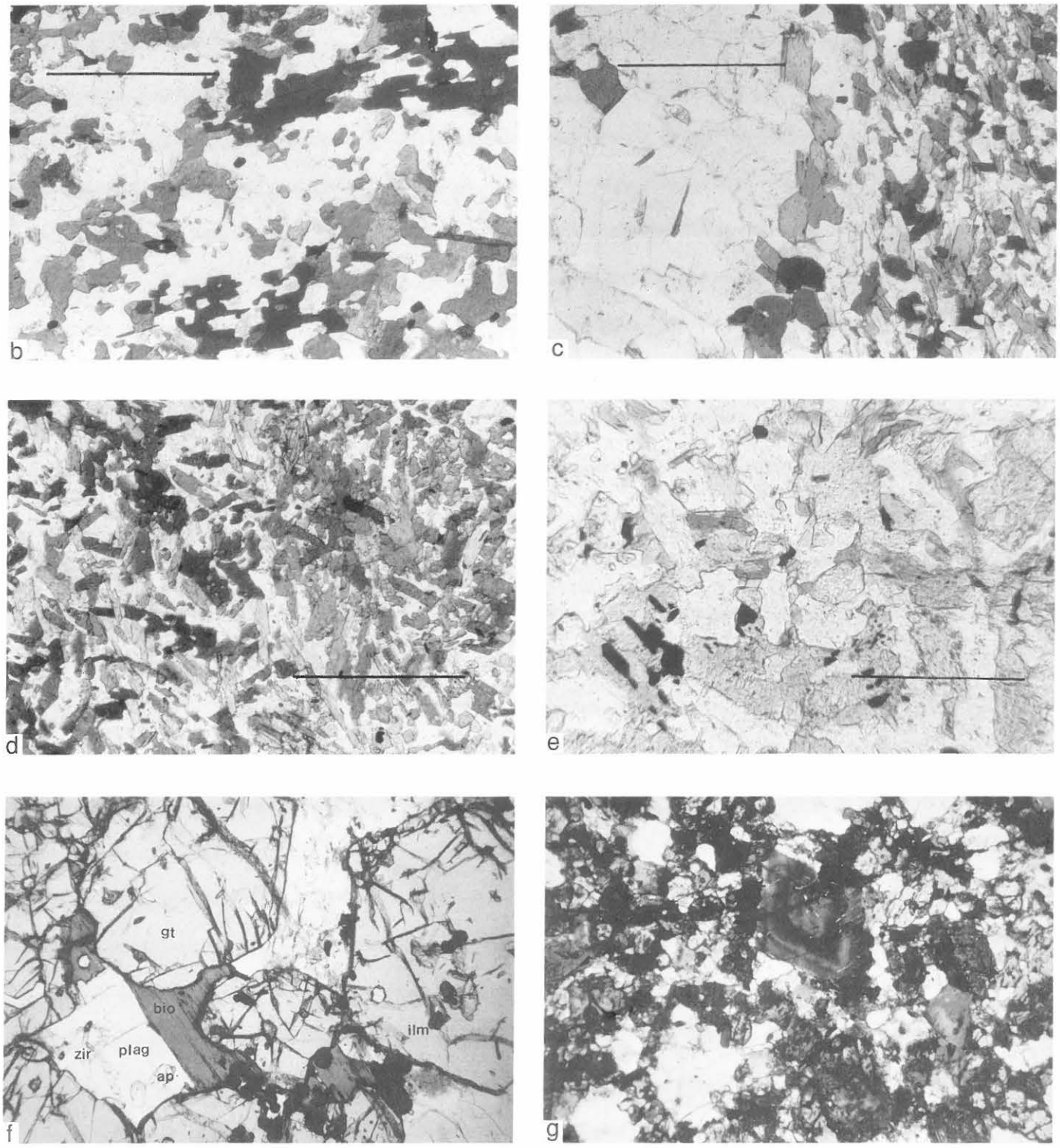

Figure 4 (continued). (b) Diorite from center of pluton on north shore of Flagstaff Lake. Note lack of zircon or monazite inclusions in biotite. Bar scale 1 mm. (c) Biotite rim on mafic enclave, Long Falls, 20 meters from granite-metamorphic contact. Bar scale 1 mm. Note fine grain size and absence of any mafic material except biotite in the rim. (d) Mafic enclave 2 cm from rim - same enclave as 4c. Bar scale 1 mm. Note acicular form of opaques (ilmenite), increase in amphibole content and grain size. (e) Mafic enclave 10 cm from biotite rim. Note reaction relationship between biotite and amphibole, equant habit of oxides, acicular apatites. Bar scale 200 microns. (f) Garnet tonalite - Bar scale 1 mm. gt - garnet; plag - plagioclase; zir - zircon; ap - apatite; bio - biotite; ilm - ilmenite. (g) Oscillatory zoned plagioclase in garnet gabbro (garnet excluded from field of view), xpl. Bar scale 1 mm.

This observation suggests the possibility that a significant proportion of biotite in the Flagstaff Lake Igneous Complex was inherited and does not represent a primary crystallization product of a granitic melt. This hypothesis is currently being investigated in detail in order to gain the information necessary to constrain the chemical, modal, and textural variation of biotite in the complex. If a significant proportion of biotite is the inherited reaction product of assimilation or reaction of granite with the mafic enclaves, then the granites do not represent liquid compositions. This could be of special importance when considering the evolution of the Flagstaff Lake Igneous Complex using the systematics of components added by the inclusions in the biotite (e.g. REE, Th, Zr, U, Pb). Two possibilities are currently being considered as working hypotheses. The first is that biotite was included as an already re-equilibrated reaction product. In this case, the rock composition would lie on a straight line between the biotite composition (plus inclusions) and the granitic magma plus crystals. The second possibility is that biotite was included as xenolithic or enclave material that reacted with the granitic magma producing the biotite and an altered granitic magma. In this case, the observed composition of the rock would reflect the biotite composition, plus the original granite magma composition, plus the added components derived from the xenoliths or enclaves. The observed suite of rocks would then exhibit a complex pattern of compositions, dependent on the nature and amount of the contaminant.

Rocks at the southern end of the Flagstaff Lake Igneous Complex form a continuous suite of compositions ranging from olivine gabbro to garnet tonalite. The garnet tonalites are equigranular rocks composed of 25-70% unzoned garnet (Table 1,2), oscillatory zoned plagioclase (~An_{50}, Fig. 4g), quartz, biotite, ilmenite, pyrrhotite, apatite, and zircon (Fig. 4f). Mineral grains tend to be euhedral, and the garnets in particular contain inclusions of the other phases present in the rock. They exhibit no foliation or lineation in thin section or hand specimen.

While the garnet tonalites are mineralogically similar to one another, percentages of phases (most notably garnet) are variable. With the exception of the xenoliths and mafic enclaves, there are no sharp contacts within the garnet tonalite body. The transition from garnet tonalite containing 25-30% garnet to that containing up to 70% garnet is gradational. The tonalites with the highest garnet contents (~50%) have a cumulate texture, but a mappable cumulate pile or zone has not been identified.

The garnet tonalite-gabbro contact is transitional, marked by an "intermediate" garnet gabbro. Since the series is continuous, the boundary between garnet tonalite and garnet gabbro is arbitrarily set at 25% garnet. Rocks with less garnet generally contain pyroxene, and those with more garnet generally contain quartz.

The garnet gabbros are characterized by orthopyroxene, oscillatory zoned plagioclase, and local cordierite. The garnet in these "intermediates" tends to be larger (up to 1 cm), less euhedral, and more magnesian as the modal percentage declines. The decrease in percent garnet is directly proportional to distance from the garnet tonalite toward the gabbro. This transition zone in the Rangeley area ranges in width from several to hundreds of meters. In addition, garnet sporadically appears as an accessory mineral throughout the gabbros of the southern part of the Flagstaff Lake Igneous Complex. The fact that garnets present in the transitional rocks are compositionally different from those in the garnet tonalite suggests the possibility that the 1-cm garnets are not xenocrysts from the tonalite. The major phases present as groundmass minerals in the garnet gabbros also occur as compositionally similar garnet inclusions, suggesting simultaneous crystallization (Table 2).

TABLE 2. MINERAL CHEMISTRY OF REPRESENTATIVE SAMPLES FROM THE RANGELEY AREA.

	Ferro-gabbro			*Diorite*	
	Ortho-pyroxene	Clino-pyroxene	Ilmenite	Biotite	Plagioclase
SiO_2	46.32	50.33	–	34.78	62.04
Al_2O_3	1.15	5.72	–	16.85	24.06
FeO	40.42	15.42	46.13	22.52	–
MgO	9.20	12.52	0.07	4.34	–
CaO	0.56	13.27	–	0.01	5.32
K_2O	–	–	–	9.49	0.09
Na_2O	–	0.77	–	–	6.94
TiO_2	0.20	1.07	50.33	3.54	–
MnO	0.97	0.34	0.78	0.11	–
H_2O	–	–	–	3.81	–
Total	98.82	99.55	98.39		

	Garnet tonalite (18)				
	Garnet grain traverse				
	Rim	Core	Rim	Biotite	Ilmenite
SiO_2	37.48	37.00	36.86	33.57	–
Al_2O_3	20.86	20.55	21.27	16.71	–
FeO	37.29	36.95	37.86	26.71	46.82
MgO	2.54	3.30	2.77	6.34	0.08
CaO	1.34	1.38	1.44	0.21	–
K_2O	.04	.02	.01	8.93	–
Na_2O	.01	.02	.01	0.07	–
TiO_2	.08	.11	.03	2.64	52.60
MnO	1.65	1.32	1.54	0.10	0.43
H_2O	–	–	–	3.73	–
Total	101.30	100.65	101.78	98.98	100.04

	Garnet gabbro (145)				
				Pyroxene	
	Garnet	Cordierite	Biotite	in garnet	gdmass
SiO_2	37.74	47.23	34.13	48.04	48.28
Al_2O_3	21.37	32.00	15.75	2.54	2.75
FeO	36.26	10.95	25.28	35.89	36.74
MgO	3.84	6.52	7.24	12.02	11.91
CaO	1.21	0.00	0.06	0.31	0.31
K_2O	.02	0.00	8.57	–	0.02
Na_2O	.01	.76	0.16	0.04	0.04
TiO_2	.07	0.00	4.20	0.19	0.04
H_2O	–	2.23	3.79	–	–
Total	101.87	99.07	99.22	99.95	100.67

TABLE 3: MAJOR AND TRACE ELEMENT ANALYSES ON REPRESENTATIVE ROCKS FROM THE RANGELEY AREA.

	14 Gabbro main body	145 Garnet gabbro	18 Garnet tonalite	140 Gabbro	LF14a Gabbro	PP 13 Gabbro enclave	FL 4 Grano-diorite
SiO_2	51.2	46.0	41.5	49.5	49.76	47.99	65.82
TiO_2	2.22	2.65	3.5	3.2	1.86	1.78	0.56
Al_2O_3	18.25	17.25	18.0	16.0	16.87	14.50	16.66
FeO	14.4	21.3	28.8	15.1	11.17	12.65	4.78
MnO	0.2	0.25	0.7	0.23	0.20	0.20	0.08
MgO	3.8	4.20	1.9	5.66	7.81	9.41	0.97
CaO	5.6	3.6	2.67	8.5	10.12	9.98	2.60
K_2O	0.65	0.39	0.63	0.19	0.99	1.10	4.98
Na_2O	2.61	2.51	1.61	2.35	2.25	2.39	3.66
Total	98.97	98.15	99.31	100.63	101.03	100.01	100.11
Cr	–	–	–	–	–	301	250
Ni	19	50	11	31	67	65	–
Cu	16	30	27	7	51	42	17
Zn	116	118	106	88	115	102	76
Rb	25	14	5	4	27	36	134
Sr	308	230	84	226	185	160	121
Y	270	490	1538	125	51	14	30
Zr	813	1058	1750	145	131	127	186
Nb	28	35	40	19	4	7	11

GEOCHEMISTRY

The geochemistry of the gabbros of the complex indicates that they have been significantly fractionated. They characteristically contain low amounts of Ni and Cr (Table 3) and high amounts of Fe. The gabbros with the highest Ni and Cr and lowest Y and Zr are still fractionated compared to liquids in equilibrium with the mantle. Absence of cumulates in the complex suggests that crystal fractionation probably occurred prior to emplacement.

The most distinctive rocks from the complex are the garnet tonalites. They are characterized by low Si, Ca, and K, and very high Fe, Y, and Zr (Table 3). Trace element abundances (especially Y and Zr) are quite variable and directly correlate with the amount of zircon and garnet in each sample.

Preliminary microprobe mapping reveals that Y is concentrated in a separate (as yet unidentified) phase within fractures in garnet. This could be caused by low-grade metamorphism or perhaps hydrothermal alterations immediately following the emplacement of the complex. Since the Y phase is not dispersed throughout the rocks, but is localized within garnets, we feel that the whole-rock analyses represent original Y concentration, which has been locally redistributed.

The continuity of the gabbro-tonalite suite can be demonstrated by considering the compositional range of rocks collected from the area shown in Figure 2. The distribution of Ca and Fe (Fig. 5) in the suite indicates that not only is the suite continuous, but it is characterized by a wide range of Ca contents at a specific Fe concentration. This cannot easily be explained as the product of any single set of processes.

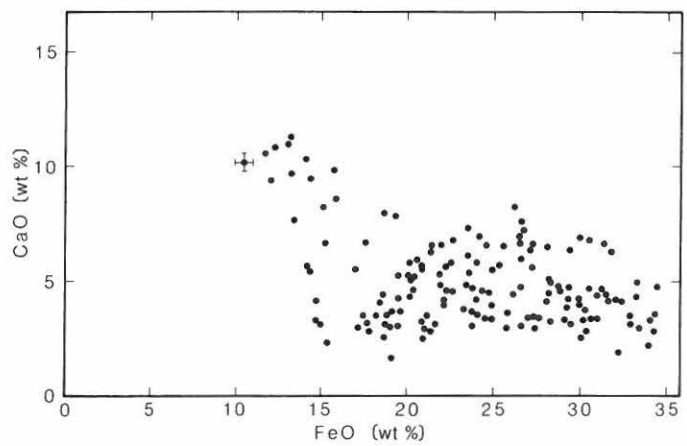

Figure 5. Ca vs. Fe for gabbros and garnet tonalites from the southwest third of the Flagstaff Lake Igneous Complex.

The incompatible elements (e.g. Y vs. Zr, Fig. 6) also exhibit a wide range of concentrations. In the case of Y and Zr, the gabbro-garnet tonalite suite exhibits a 20-500 fold increase in concentration. As with Ca and Fe, the data set exhibits a wide range of Y values at any given Zr value.

DISCUSSION AND CONCLUSIONS

From the information presented above, it is evident that many different, interacting processes are responsible for the phenomena observed in the Flagstaff Lake Igneous Complex.

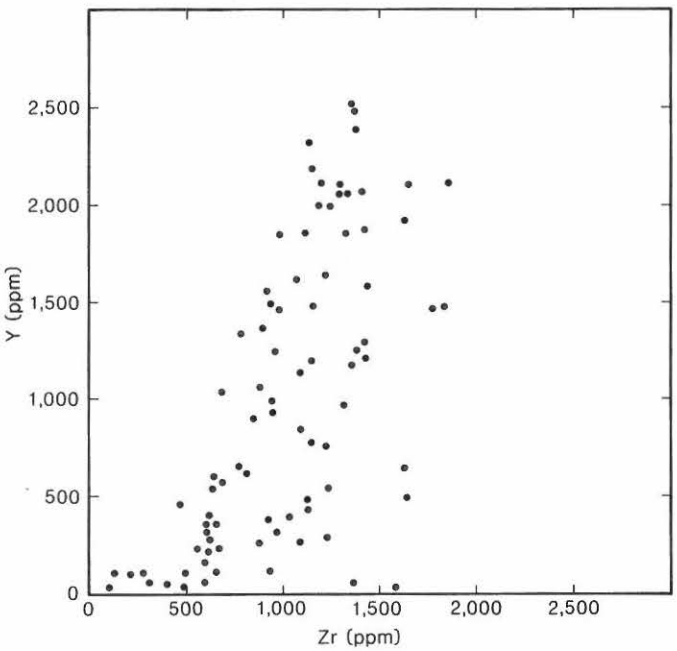

Figure 6. Y vs. Zr for gabbros and garnet tonalites from the southwest third of the Flagstaff Lake Igneous Complex.

These include the processes responsible for (1) the origin of the mafic enclaves; (2) the origin of the garnet tonalite; (3) the observed continuous chemical and mineralogic series between garnet tonalite and gabbro; and (4) the significance of the monazite and zircon inclusions in biotite in the granite.

There are two potential explanations for the origin of the mafic enclaves found in the granite, tonalite, and trondhjemite. They either originated as blocks of gabbro, broken off and swept up as the granite intruded a previously solidified gabbroic pluton, or they represent quenched "pillows" of gabbro, injected as a liquid into a partially solidified granite body (Taylor et al., 1980; Furman and Spera, 1985). It is important to distinguish the two hypotheses in order to establish the relative timing of the emplacement of the major rock types. The irregular, ameboid shape of the enclaves, the inclusion of patchy zoned sodic feldspar, the fine grained rims and inward-coarsening texture, and the cuspate interface with the host rock all point to the enclaves originating as liquids injected into a low temperature crystal mush. Therefore we feel that granitic and gabbroic magmas coexisted during the major emplacement event of the complex. The rough zoning pattern of the intrusion itself suggests the possibility that the magma chamber was zoned, with granite lying above a larger body of gabbro. With the information presently available, two intrusion histories seem possible. The first scenario is that granite originally filled the entire magma chamber and was then partially pushed out by an influx of gabbroic magma. Alternatively, the magma system rose through the crust as a unit with the granitic magma being produced from melting of the wall rock as the composite pluton rose through the crust (McBirney, 1980).

There are several hypotheses that could explain the origin of the garnet tonalite. The first point to be addressed is whether the rocks are igneous or metamorphic in origin. Previous workers have called them (1) restites of partial melting (Boone, 1973; Vehrs, 1975); (2) reaction hornfels (Boudette, 1978); (3) products of igneous differentiation (Lyttle, 1976); or (4) restites mixed with their own partial melts (e.g. similiar to cases described by Clark and Lyons, 1986; Plank, 1987).

Restites of partially melted metapelite could produce a rock mineralogically similar to the garnet tonalite (Gribble, 1970), but the body would not cut the regional strike as the tonalites invariably do. Neither would one find oscillatory zoned plagioclase, or the equigranular texture exhibited by the tonalites. If the partial melt were not removed, but remained to mix with the refractory phases, some of these problems would be eliminated (oscillatory plagioclase, crosscutting relationship). However, garnet of metamorphic origin is usually free of inclusions and often zoned (Clark and Lyons, 1986). The garnets from tonalite and gabbro samples are both full of inclusions, identical to the matrix assemblage, and are unzoned. This suggests that the garnet co-precipitated with the other phases in the rocks. We do not eliminate the possibility that the tonalite magma was partially crystalline at the time it mixed with the gabbro, but we feel the evidence is not compelling. This is in contrast to the pervasive evidence that the granitic magma was partially crystalline when it interacted with the gabbro.

The mobilization of Y, represented by its concentration in fractures in the garnet, indicates that some late-stage metasomatism did occur. Large bodies of apparently metasomatically generated garnet-rich rock (skarns) are associated with the contacts of many plutons. However, garnets associated with skarns are different in most characteristics from those from the Flagstaff Lake Igneous Complex. Garnets from skarns are usually andradite-grossular rich, strongly zoned, inclusion-free, and the bodies occur in veins. In addition, skarns are characterized by rocks with a small number of phases and variable phase proportions. The garnet tonalite is characterized by a large number of major phases and relatively constant phase proportions. An origin by metasomatism also would not account for the angular, randomly oriented blocks of hornfels or mafic enclaves as observed in the garnet tonalite bodies. Finally, it would be difficult to explain the continuous series of rocks from gabbro to tonalite by metasomatism.

An igneous origin for the tonalites implies two possible source areas and processes: differentiation of mantle-derived magmas (gabbro) or anatexis of the country rock. To date, no source has been identified in the country rock to produce a Fe-rich, alkali- and silica-poor melt. The coincidence of garnet tonalite with sulfidic, graphitic, Fe-rich metapelite and the numerous xenoliths strongly suggests a petrogenetic relationship between them. If, however, the country rocks cannot be

demonstrated to be related to the tonalite-gabbro series, it does not rule out the possibility of assimilation or anatexis at depth.

The wide range of both major and trace element compositions in the tonalite-gabbro suite provides great difficulty in specifying a petrologic process or set of responsible processes. Fe-rich, Ca-poor magma compositions could potentially be produced from gabbro by fractional crystallization of augite. The low but variable Cr and Ni contents of the gabbros indicate that at least some fractional crystallization has occurred. However, the increasing amount of garnet in the suite as Fe increases negates fractional crystallization as a major process in the origin of the tonalites. This is a consequence of the high partition coefficient for Fe, Y, and Zr in garnet. Fractional crystallization of garnet would prevent the evolution of the magma system to the observed high Fe, Y, and Zr contents. In addition, even if the partition coefficient for those elements was 0, over 99% fractional crystallization would be needed to raise Y from 20 to 2500 ppm. Alternatively, the observed distribution could be explained as due to the accumulation of igneous garnet derived from an Fe-rich magma, assimilation of metamorphic garnet, or to the mixing of two liquids, one gabbroic and the other Fe-rich. The theory of garnet accumulation or assimilation is supported by two observations. First, in a few locations, tonalite contains almost 70% garnet and has a cumulate texture. Second, garnets in the garnet gabbro often have rims of biotite, suggesting that they have reacted with the gabbroic magma. On the other hand, garnets in tonalite and gabbro in general have mineral inclusions similar to those in the groundmass of the rock in which they are found. Also, there is a correlation of the Mg-content of the garnet with the Mg-content of the igneous rock. This would not necessarily be the case if the garnet was assimilated from a garnet-bearing metasedimentary rock or crystallized in an Fe-rich magma prior to magma mixing. The biotite rims on the garnets could be a result of a change in the crystallizing assemblage, analogous to orthopyroxene rims on olivine. Therefore, it is our preliminary conclusion that the observed distribution is due to a series of interacting processes: (1) the fractional crystallization of the gabbro; (2) garnet accumulation, at least in the most garnet-rich tonalites; and (3) mixing of gabbroic and Fe-rich tonalite magmas.

The proportion of the biotite in the granites, trondhjemites, and tonalites that was inherited from either the mafic enclaves or metamorphic stope blocks is a critical point from several perspectives. All of these center around the determination of the extent to which these rock types represent liquid compositions. Inherited biotite could add Th, Zr, Pb, and U to the bulk rock from monazite and zircon inclusions, in addition to the exchange of Mg, Ti, and Fe from re-equilibration of biotite itself.

The proportion of schlieren in the granite, trondhjemite, and tonalite ranges from 0 to 20%, with the highest percentage being found in the trondhjemites. It is certain that at least some of the biotite crystallized from the magma and was not incorporated in the solid state as enclaves, xenoliths, or schlieren. It is not unreasonable to suggest that monazite was a crystallizing phase during a large part of the crystallization history of the Flagstaff Lake Igneous Complex granites and trondhjemites. Therefore, it is extremely difficult to distinguish the origin of individual biotite crystals in the granites. However, several approaches could provide useful information and are presently being applied to these rocks. First, a statistical evaluation could be made of the number of inclusion-bearing biotites vs. inclusion-free biotites in a given size range. Second, a detailed study is presently being conducted of the range of biotite compositions in the biotite rims, metamorphic stope blocks, and schlieren of both types. Third, the range in monazite and zircon compositions in biotite in the same rock types, in addition to those in granite, tonalite, and trondhjemite is being determined by microprobe to evaluate any possible correlations.

In summary, on the basis of the data presently available, we feel that the processes responsible for the observed suite in the Flagstaff Lake Igneous Complex were dominated by magma mixing of at least three discrete end members: fractionated mantle-derived gabbro, granite, and garnet tonalite. The specific mechanism driving the mixing is as yet unclear. However, the available evidence is consistent with autointrusion as suggested for the Bottle Lake Complex by Ayuso (1984) or with the creation of a granitic cap on a rising gabbroic magma system by anatexis of the wall rock (DePaolo, 1981; McBirney, 1979, 1980). Fractional crystallization played a relatively minor role after emplacement, although the gabbroic rocks show evidence of significant olivine and pyroxene fractionation.

ACKNOWLEDGMENTS

Many individuals have been of great help in our work. We thank the Maine Geological Survey for providing logistical support to R. L. Nielsen and E. S. Landis. The initial stages of this investigation were funded though a summer Graduate Research Board award from the University of Maryland. Mike Doughton of the USGS at Reston provided invaluable help in calibrating XRF standards needed to analyze some of the more unusual rock compositions we encountered. Finally and most importantly, the Rangeley Mining Company and Industrial Garnet Extractors graciously provided us with access, logistical and financial support, and drill core from some of the most critical areas. Most of all we thank Frenchy -- for being Frenchy.

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Maine Geological Survey
Studies in Maine Geology: Volume 3
1989

Mesozoic Dikes of Southern Coastal Maine: An Historical Perspective and Update on Research, 1838-1988

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ABSTRACT

The Mesozoic dikes of southern coastal Maine have experienced 150 years of research interest that began with C. T. Jackson's original survey of the state in 1838 in which he hypothesized three phases of dike intrusion. Investigation in the 1800's culminated with Kemp's (1890) detailed mapping and petrographic examination of dikes in the Bald Head exposures near Ogunquit. Research in the early 1900's included an extensive regional survey of petrographic dike types by Wandke (1922a) and a study of the detailed intrusive relations in multiple dikes by Haff (1939) at Cape Neddick. The regional geologic survey of York County by Hussey (1962) provided a solid background in which an older and younger group of dikes were thought to be intermediate in age between the Early Mesozoic Agamenticus alkaline igneous complex and the younger Cape Neddick gabbroic complex.

Recent orientation studies of dikes in coastal exposures (Swanson, 1982) revealed a progressive northward change in dike trend from N25^{o}E at Gerrish Island through N45^{o}E at Cape Neddick to N60^{o}E at Bald Head. This dike trend variation is interpreted to reflect increasing control on dike intrusion by preexisting structures. These preexisting structures include a near-vertical N60^{o}E bedding-anisotropy as the vertical limbs of regional F_2 isoclinal folds and a N45^{o}E F_3-related cleavage-anisotropy in the host Kittery Formation.

A detailed mapping program (1:240) focusing on these near-continuous coastal exposures (Swanson, 1982, 1983, 1984, 1988, and in progress) has included most of the 12 kilometer coastline about the Agamenticus and Cape Neddick igneous complexes and south to Gerrish Island. The resulting detailed mosaic map sheets (1:480) of large, continuous sections of outcrop allow the recognition of a complex mafic-felsic dike intrusion sequence based on the detailed crosscutting map relations and the observed compositional and textural varieties. These relationships suggest the following intrusive history: a sequence of early Agamenticus minor (mafic and felsic) intrusions associated with peripheral host rock deformation; intrusion of the main phases of northeast-trending mafic dikes; a late Agamenticus sequence of mafic and felsic dikes largely postdating the main phase dikes; and lastly, a complex, Cape Neddick mafic-felsic sequence of intrusions.

INTRODUCTION

C. T. Jackson's initial survey of the state in 1838 began a long history of geologic research on the dikes of southern coastal Maine. These coastal dike exposures are part of an intense coast-parallel swarm complex of Mesozoic age related to rifting and crustal extension in the early stages of development of the Atlantic Ocean. Sporadic reports were published throughout the late 1800's and the early to middle 1900's with Kemp (1890) attempting the first detailed mapping and petrographic study of these dikes. Later, Wandke (1922a) described the various petrographic dike types in these exposures, while Haff (1939) presented a detailed description of the locally complex intrusive relations. Hussey (1962) summarized the observed dike characteristics in light of the geology of York County. Most recently, Swanson (1982, 1983, 1984, 1986, 1988, and in progress), by

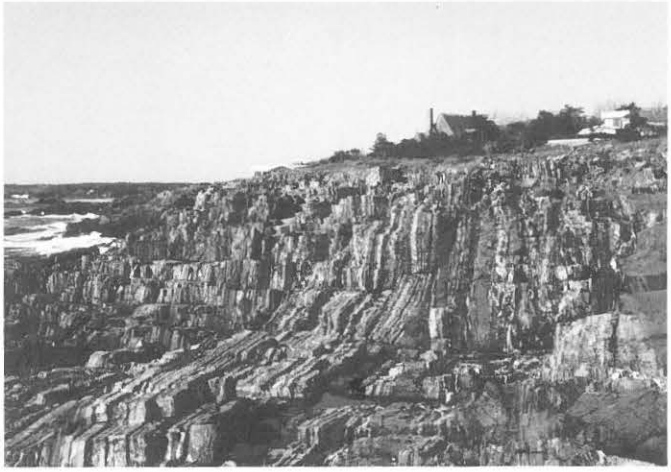

Figure 1. Photograph of the coastal exposures just south of Bald Head Cliff showing the prominent near-vertical bedding anisotropy in the host Kittery Formation and several concordant Mesozoic dikes.

very detailed mapping, has begun to unravel the complex structural and intrusive history of these remarkable exposures (see Fig. 1).

The Mesozoic dikes and associated intrusive complexes of Agamenticus and Cape Neddick are found in what are some of the best exposures anywhere in the world, capturing the attention of geologists for 150 years. The descriptions of the earlier research on these dikes is presented in an historical perspective from 1838-1988 that is divided into four distinct periods of focus: the 1800's, the early 1900's, the 1960's, and current research in the 1980's. Figure 2 will serve as a location guide for the map areas and as a guide to geographic locations and general geology of the southern Maine area.

EARLY WORK IN THE 1800's

C. T. Jackson's (1838) description of the trap rocks found at Bald Head in York begins "There are many singularly obscure rocks in this town, and every variety of metamorphosis which slate rocks can undergo, may be observed. Sienite veins and trap dykes, of various ages, occur in great numbers, cutting across the strata, and furnishing an index to their history." The host rock to these intrusions was described as a "striped novaculite or hone-slate" which has been "cut through by an infinity of dykes."

Three distinct dike intersections were apparent to Jackson in these exposures. Jackson's sketches are presented here as Figure 3, redrafted in the style of the original engravings. The oldest intrusions were large N55^{o}E-trending dikes of porphyritic greenstone trap that included a large trap dike, 17 m wide, just below Bald Head Cliff. A second series of smaller northeast-trending irregular trap dikes were found to crosscut the first. A brown scoriaceous trap represented a "third cutting" by yet another series of dikes. After examining numerous exposures in this general area, Jackson was able to "confidently confirm 4 or 5 distinct eruptions of molten trap rocks have burst through the strata of Maine."

C. H. Hitchcock in 1861 made a visit to the Bald Head exposures to examine the "trappean rocks in the southern part of Maine." Hitchcock described these dikes as "frequent, but generally uninteresting." The dikes were thought to occupy former fissures where "some general cause has produced fractures over a large area of essentially the same direction." Thus, it was possible for Hitchcock to "distinguish the ages of the trap dikes as well by their direction as their difference in mineral structure."

As previously noted by Jackson (1838), Hitchcock found the dikes in York to be of three different ages. The porphyritic trappean rocks in very large dikes were thought to be the "oldest series of dikes off the coast." The second series of northeasterly dikes that cut the porphyritic traps were found to be "the most common of all the dikes in southern Maine." The third series of brown scoriaceous trap was found to cut across the "common trap" thus reaffirming Jackson's earlier observations and interpretations.

J. F. Kemp in 1890 reported that "the southern coast of Maine is very generally seamed with dikes." Kemp conducted the first detailed mapping of intrusive relations in the coastal exposures at Bald Head and Kennebunkport as well as reporting on the first petrographic examination of these dikes. The exposures at "Bald Cliff" were described as "finely stratified quartzytes, slates and jaspery layers...that belong to the Merrimack group of the New Hampshire survey." It was noted that "sometimes a slatey cleavage runs obliquely across the bed, even when but a few inches thick and between two quartzyte layers," a fact that has proved significant in recent structural studies nearly 100 years later.

Kemp had mapped 18 dikes at the Bald Head Cliff exposures and his sketch, redrafted in his original style, is presented as Figure 4. The crosscutting relations reported by Jackson (1838) and Hitchcock (1861) were verified by Kemp's mapping and interpretations.

The petrographic examination of these dikes led Kemp to propose a "type" example as a holocrystalline olivine diabase. All other dike types were thought of as "structural deviations" from the "type" olivine diabase. Three different petrographic varieties were recognized in the Bald Head Cliff exposures. These include the olivine diabase, a melaphyre (a dark feldspar phyric basalt with fine-grained mesostasis) and an augite porphyrite. The first chemical analysis of these dike rocks was also reported by Kemp (1890). The analyzed sample was from the large augite-porphyrite just below Bald Head Cliff that had earlier caught the attention of Jackson (1838).

THE EARLY 1900's

F. J. Keeley (1914) stated that "it would probably be difficult to find a more remarkable display of igneous rocks than that along the coast of Maine, south of Ogunquit." Keeley's work

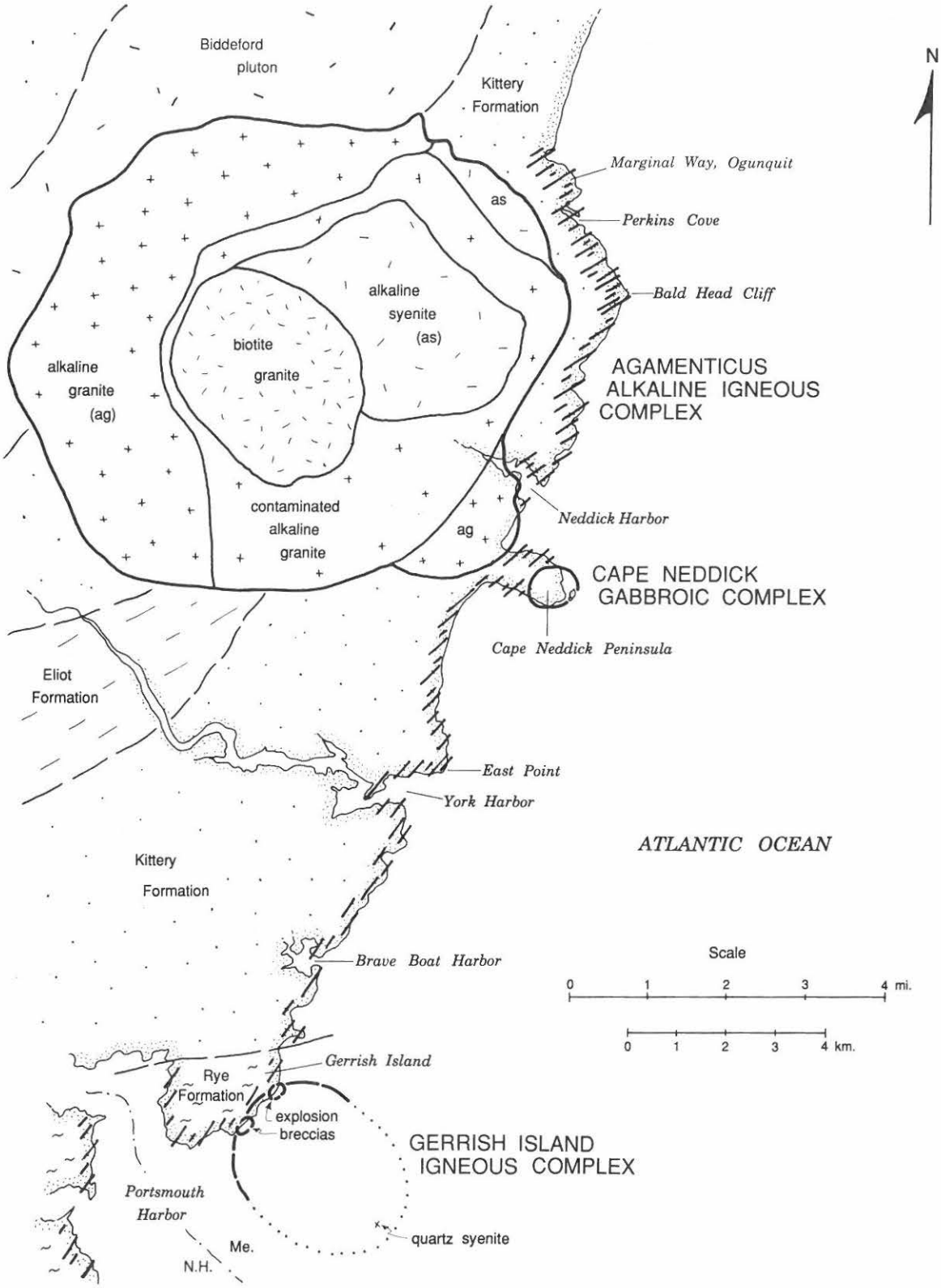

Figure 2. General geology of the southern Maine area (modified after Hussey, 1962) that includes the major Mesozoic intrusive complexes and the coast-parallel dike swarm. Dikes are shown schematically as short bold lines in coastal exposures only and do not include dikes in inland exposures.

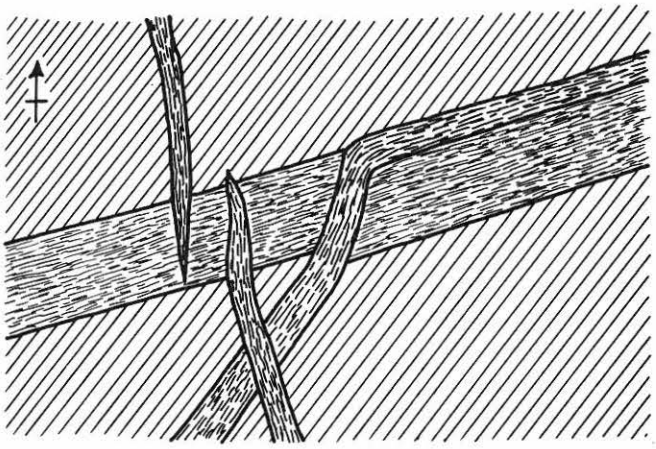

Figure 3. Dike intrusion relations depicted in Jackson (1838) from observations at Bald Head. No scale included in the original diagrams.

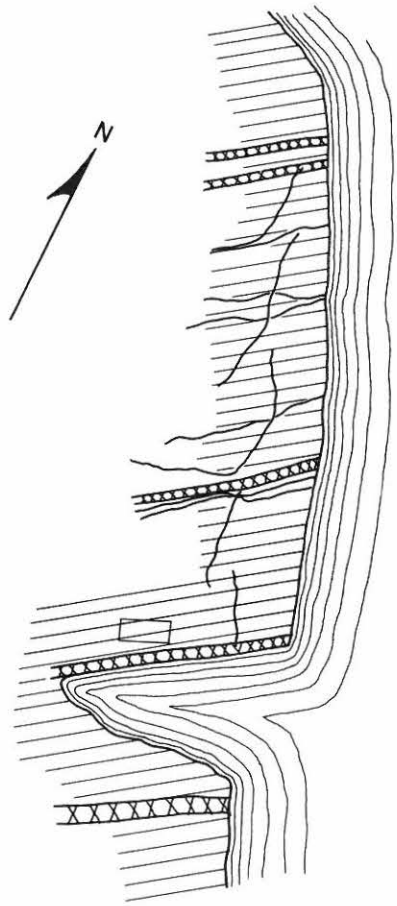

Figure 4. Dike intrusion relations at Bald Head as depicted by Kemp (1890). No scale or north arrow included in the original diagram. Refer to Figure 6 for comparison.

was primarily a petrographic study of dike rocks found near Ogunquit, Maine and Pigeon Cove, Massachusetts. The localities examined by Keeley included outcrops around a small cove on Israels Head and exposures south of Perkins Cove at Spouting Rocks in Ogunquit. Keeley was decidedly impressed with the diverse array of dike types in these exposures. To quote Keeley (1923) "These dikes form an interesting and puzzling group. In less than 100 ft. we have a half dozen rocks ranging from acidic to basic and differing not only among themselves but from any others in the district." The dike types reported by Keeley included diabase (plagioclase-porphyrite, coarse-grained, pyritic, and calcite-filled vesicular varieties), trachyte (with kaolinitized orthoclase), rhyolite, and granitic quartz porphyry.

A xenolithic dike south of Perkins Cove caught Keeley's attention, one of several in these exposures later studied by Powers (1915) and featured as a photograph in Hussey's (1962) report. The dike was a diabase porphyrite (plagioclase phenocrysts) with altered syenite xenoliths (orthoclase phenocrysts with riebeckite) related to the nearby Agamenticus alkaline intrusive complex.

S. Powers (1915) reported on investigations into "the origin of the inclusions in dikes." Powers' investigations involved dozens of dike localities from around the world including a stop at Ogunquit, Maine, as one of several examples of "rising inclusions." The locality visited by Powers is halfway between Perkins Cove and Bald Head. These exposures were described as "thick-bedded slates...intruded by vertical trap sills of two or three generations, the older being porphyritic." The xenolithic dikes were described as "inclusion-bearing sills."

The particular dike examined by Powers is of the older porphyritic type, an augite-biotite kersantite. The composition of the numerous xenoliths observed includes granite, coarse- and fine-grained syenite, aplite, slate, and vein quartz. Powers described the nature of the fused contacts and resorption of xenoliths in relation to the development of chill margins along exposed sections of the dike.

By this time, largely owing to the work of F. J. Katz (1917) with the U.S. Geological Survey, the intrusive rocks (dikes and other Mesozoic intrusions) were being included in the interpreted geologic history of southern Maine. A. Wandke (1922a) conducted perhaps the first extensive study of these igneous rocks in his paper titled "The intrusive rocks of the Portsmouth Basin." Wandke's description begins "the dikes, because of their number, their wide range in composition and the manner and order of their intrusion form a group of rocks well-deserving of more detailed study." He reiterates that "the best display of dikes occurs along the coastal section from Perkins Cove (south) to Brave Boat Harbor." In addition to descriptions of the dike rocks, Wandke's investigations provided the first mapping and lithologic-petrographic descriptions of the Agamenticus alkaline igneous complex (1922a) and the Cape Neddick gabbroic complex (1922b).

Wandke (1922a) found the dike compositions to range from olivine-bearing lamprophyre through diabase, diorite, and granite porphyry to paisanite and aplite. The dikes were subdivided into two groups generally considered to be earlier and later than the "stocks and batholiths." The earlier group consisted of diabase (as the oldest) and variations to quartz diorite (with lighter flow layering) and lighter dikes of granite porphyry (with feldspar and quartz phenocrysts). The second group of dikes consists of two subgroups of "diaschistic dikes" that include paisanites (a single exposure), camptonites (with poikilitic hornblende) and tinguaites (with aegirite needles) associated with the alkaline Agamenticus complex and a still later series of aplites and late diabases that crosscut all other rocks.

J. C. Haff (1939) presented the first detailed structural and petrographic investigation of the "multiple dikes of Cape Neddick." The dikes were described as "extraordinarily numerous", in places comprising "more than 25% of the rocks exposed." Haff (1939) was primarily interested in the multiple dike intrusions (with each dike developing its own chill margins) exposed along the southern shoreline of the Cape Neddick Peninsula. The composite dikes (mafic dikes with unchilled felsic interiors) and a single alkaline vitrophyre dike were the subjects of later papers (Haff, 1941, 1943). Haff (1939) introduced terms such as suite, complex, component, member, epoch, stage, and group in his descriptions of the dike intrusion. Perhaps his most useful terms in describing the intrusive relations were intrasectate (reintrusion within an earlier dike) and extrasectate (reintrusion along the margin of an earlier dike). Variations in these relations led to the development of braided multiple dikes.

Haff collected specimens of 338 "minor intrusions" along the south shore of the Cape Neddick peninsula. The petrographic varieties found in these exposures included diabase (olivine, plagioclase-porphyritic, augite-porphyritic, acicular plagioclase, and modified varieties); porphyrites (hornblende-, augite- and diorite-); basalt (olivine); lamprophyre (augite, modified); basic (modified, meta-); and felsic (granophyre, felsophyre, hornblende granite, aplite, quartz porphyry, bostonite, and augite andesite).

In reference to the association of dikes and the soda granite at Agamenticus and the Cape Neddick gabbro, Haff states "each of these intrusions contains its own suite of dikes whose members are not, however, involved in multiple relationships with one another." In addition to the two distinct suites of post-alkaline granite and post-gabbro dikes, Haff recognized at least two intrusive phases of dikes that were generally older than the intrusive complexes. These dominantly basic dikes consisted of the multiple dike suite with both mafic and felsic members oriented at N40-45°E and a later set of basic dikes without multiple relations that transect the earlier dikes at large angles (northwest-trending). Several stages of intrusion were apparent in the "highly involved complexes" in the multiple earlier dike set.

THE 1960's

Hussey's (1962) report was the first comprehensive description of the bedrock lithologies, their distributions, structure, and the geologic history within the southern York County area (see Fig. 2) which serves as host to the intrusion of the coast-parallel dike swarm. Field and petrographic studies helped to subdivide the Mesozoic igneous rocks into the alkaline complex at Agamenticus (correlated with the White Mountain plutonic-volcanic series), small circular to elliptical basic complexes, and the basaltic to rhyolitic dikes.

Hussey's description of the major group of dikes noted that "they are most abundant in the coastal areas and tend to become scarcer inland, in the vicinity of Berwick and Sanford, they are rare," suggesting the presence of a well-defined coast-parallel dike swarm. Two general age groups of dikes were distinguished based on observations along exposures from Bald Head Cliff, York, to Perkins Cove, Ogunquit. Dikes of the older group were found to be more numerous, from 15 cm to 15 m in width and intruded parallel to the upright bedding in the Kittery Formation. Dikes of this group were also "more diversified as to type and texture" ranging from diabase and basalt to rhyolite. Several subgroups could be separated locally based on crosscutting relations. This group would include the first (older porphyritic) and second series of dikes reported by Jackson (1838) and Hitchcock (1861). Dikes of the younger group were found to be "rather narrow and transect the bedding and the dikes of the older group at nearly right angles" (northwest-trending) and consist only of "deep rusty red basalt." This younger group would correlate with the third series of scoriaceous dikes of Jackson (1838) and Hitchcock (1861).

The interrelationships between the intrusive complexes and the major group of dikes within the coast-parallel dike swarm was "of paramount importance" in determining the age relations of the basic complexes with respect to the Agamenticus complex. In this regard, Hussey (1962) states that "numerous dikes cut the alkaline granite and nowhere is the reverse observed" whereas all diabase dikes were found to be cut by the normal gabbro of the Cape Neddick gabbroic complex. The few dikes within the basic complexes were thought "to be related to the complexes and not to be part of the major group of dikes." The major group of dikes was thought to be generally of the same age, being intermediate between the Agamenticus and Cape Neddick complexes.

A few years later, Gaudette and Chapman (1964) focused on the late structural and intrusive history of the Cape Neddick gabbro complex. Their study included the minor mafic dikes within the complex referred to by Hussey (1962) as well as a system of radial and concentric joints. There were 18 small basic dikes found within the intrusive complex, generally less than a foot in width. All of these dikes showed extensive recrystallization and alteration resulting in a hornfelsic texture. The mafic dikes were found to be distributed in a radial pattern and closely

associated with intermediate to granitic phases. This assemblage of intrusions and structures was interpreted by Gaudette and Chapman (1964) as "a later minor explosive phase...a sudden doming action accompanied by the injection of only a few thin mafic dikes." This late structural and intrusive phase was thought to have displaced the central portion of the complex upwards as a "single conical block isolated by cone fractures."

Additional investigations in the 1960's included a study of wall-rock alteration at the Cape Neddick gabbro (Woodard, 1968a) and two unpublished studies directly related to host rock structure and sequence of minor intrusions. Woodard (1968b) attempted to decipher the complex deformational history of the Kittery Formation host rocks and its effects on dike intrusion and recognized host rock deformation peripheral to the Agamenticus and Cape Neddick intrusive centers. Woodard (1969) suggested the development of three spatially overlapping, sequential felsic-mafic intrusive cycles in these exposures related to the central intrusive complexes at Agamenticus, Cape Neddick, and an hypothesized center near York Harbor.

Two age studies in the 1960's included a whole rock Rb/Sr determination of 227±3 Ma for the Agamenticus complex (Hoefs, 1967) and a fission track age of 119±9 Ma for apatites in the Cape Neddick gabbro by Christopher (1969). Additional isotopic age determinations (K/Ar, biotite) were reported through the 1970's that included ages of 216±4 Ma (Foland et al., 1971) and 228±5 Ma (Foland and Faul, 1977) for Agamenticus and 116±2 Ma (Foland and Faul, 1977) for Cape Neddick. These reported ages appear to confirm the earlier ages of Hoefs (1967) and Christopher (1969) for these intrusive complexes.

CURRENT RESEARCH IN THE 1980'S

A study by McHone and Trygstad (1982) on the Mesozoic dikes of southern Maine provided a brief regional survey of the orientations and mineralogies of approximately 50 dikes throughout the southern Maine area, recognizing Early Jurassic dolerites and Jurassic to Cretaceous lamprophyres. Major oxide compositions of 9 of these dikes indicate that alkalic and tholeiitic (quartz and olivine normative) dolerites are represented (McHone and Trygstad, 1982), but did not include any dikes from the specific area of this report. Orientations were dominantly N30°E with a second maximum at N60°E. These dike trends are similar to those within the coastal exposures discussed in more detail below. McHone and Butler (1984) in a broader study of igneous provinces throughout the northeast included the coast-parallel dike swarm in their coastal New England igneous province. McHone (1984) has compiled dike ages for Maine and the rest of New England. Dolerite dike ages were found to range from 235-165 Ma. Ages for dikes within the coast-parallel dike swarm of this report are few, with ages of 216 and 224 Ma reported (McHone, 1984) for dikes near Seabrook, New Hampshire.

Hussey et al. (1984) focused primarily on the structural and sedimentologic characteristics of the Kittery Formation but included a detailed map of the structure and dike intrusions in the Marginal Way exposures of Ogunquit. Several phases of dike intrusion are apparent in these exposures including a few "pre-S_3" trachytic intrusions related to Agamenticus.

The remaining sections of this report are based on Ph.D. investigations and post-doctoral research that began in 1980. This research has covered four summer field seasons (1980, 1981, 1984, and 1986) and is still in progress at this time. Initial studies were focused on the structural and intrusive relationships within the abundant Mesozoic dikes that are found in the near-continuous coastal exposures in southern Maine. The areas of study which afforded the best exposures (see Fig. 2) were between Perkins Cove, Ogunquit and south to Gerrish Island, Kittery in southernmost Maine.

A continuous outcrop mapping technique was developed as an accurate method of recording field data on a series of detailed base maps. The construction of the detailed base maps begins with the preparation of individual field maps by outcrop grid mapping using intersecting 100 foot measuring tapes at a scale of 1:240. The individual field sheets are reduced into a combined mosaic for each subarea at a scale of 1:480. At present, mosaic map sheets for 14 subareas have been completed or are nearing completion covering over 12 kilometers of coastal exposure. The resulting mosaic map sheets at a scale of 1:480 are being used to study the intricate intrusive relations in terms of the varying orientations, structural styles of intrusion, crosscutting relations, and dike types preserved in these exposures in an effort to establish a composite sequence of minor intrusion for the entire map area.

Dike Swarm Structure

Characteristic dike swarm parameters for the coastal dike exposures include a regional average for dike thickness of 1.12 m/dike; dike intensity of 137 dikes/km; and crustal extension of 16.6% (Swanson, 1982). Maximum observed dike thickness was 24 meters, while a maximum crustal extension of 22% was calculated for the exposures along the Cape Neddick Peninsula.

Dike orientations within the coast-parallel dike swarm were studied in relation to possible preexisting structural control on the pattern of dike intrusion. Average dike orientations from over 1600 dikes (Swanson, 1982) in six subareas throughout the coastal exposures were found to vary systematically from N25°E at Gerrish Island through N45°E at Cape Neddick to N60°E at Bald Head (Fig. 5). This systematic dike trend variation is interpreted to represent a northward increase in the degree of preexisting structural control by the N60°E-trending upright beds of the Kittery Formation. At Bald Head, where structural control is most prominent, there is a bimodal distribution of dike orientations of N45°E and N60°E. This is shown clearly in the dike trend histogram for Bald Head (Fig. 5) and in the recently mapped pattern of intrusions in these exposures (Fig. 6). These are the same exposures visited by Jackson (1838), Hitchcock

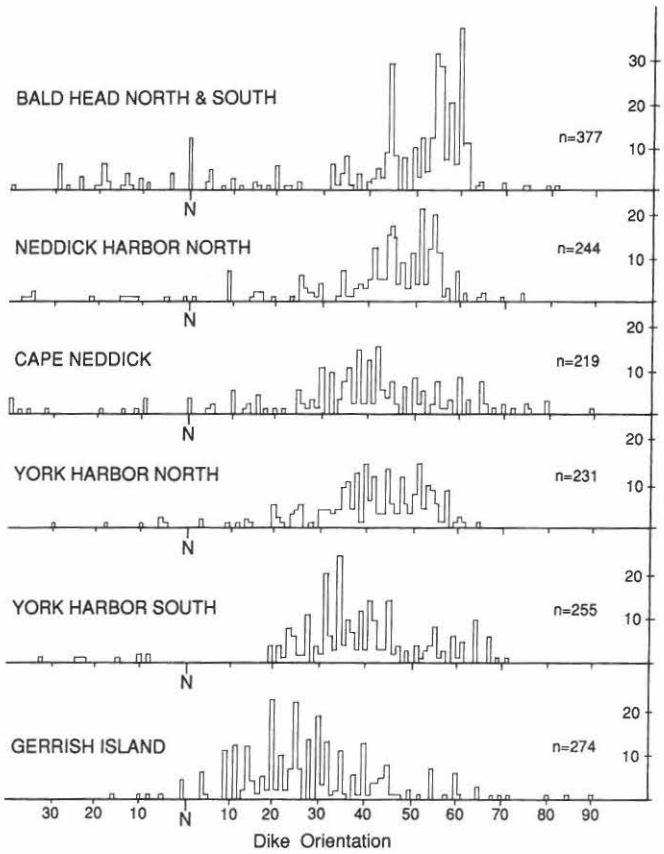

Figure 5. Comparative histogram of dike segment orientation for six subareas along the southern Maine coast. The systematic dike trend variation is interpreted to reflect an increasingly predominant preexisting structural control toward the northern Bald Head exposures.

(1861), and Kemp (1890). Kemp's dike map is also included as Figure 4 in this report.

The bimodal dike trend distribution for the Bald Head vicinity reflects a dominant control on the pattern of dike intrusion by the N60°E bedding-anisotropy as well as a N45°E oblique-cleavage anisotropy in the pelitic schistose units of the host Kittery Formation. The N60°E bedding-anisotropy represents the upright limbs of gently-plunging F_2 regional isoclinal folds originally defined by Hussey (1962) and discussed in more detail by Hussey et al. (1984). The uniformity of fold limb orientation and the alternating quartzite and metapelitic units of the turbiditic Kittery Formation creates a pervasive N60°E-trending vertical planar anisotropy (see Fig. 1) throughout these exposures. This anisotropy strongly controls the dike segment orientation by dilational reactivation during intrusion.

The N45°E cleavage-anisotropy represents a slip cleavage within the schistose beds of the Kittery Formation suggesting layer-parallel dextral shearing of the host rocks to these intrusions (Swanson, 1982, 1984). It was this "oblique slatey cleavage" that was first noted by Kemp (1890). This anisotropy is also developed as an axial plane cleavage associated with steep to moderately plunging asymmetric z-shaped F_3 folds (Swanson, 1984). F_3 folds become more inclined northward in the Marginal Way exposures (Hussey et al., 1984), but the cause of this relationship is not clear at this time.

As a result of the prominent control on dike orientation by preexisting structures, the measured directions of horizontal extension (opening directions) become a more accurate representation of the overall dike swarm trend and the orientation of minimum principal stresses than simply the average orientation of any exposed, structurally-controlled dike segments. A brief survey of opening directions for intruded fractures suggests that many of the N60°E bedding-parallel dikes in the Bald Head area have a sinistral component to their dilation, suggesting a more northerly oriented regional, extensional dike system. Detailed study of opening directions for dike-intruded fractures and the overall pattern and distribution of extensional strain is currently in progress.

The structure of the dike terminations was of particular interest in the early phases of this study (Swanson, 1982, 1983, 1988) as these structures are exquisitely developed in the exposures. Dike terminations in en echelon pairs were found to vary from simple tapers to interconnected transform-faulted segments as well as rotated slab, rotated block, and a rare kink-type dike termination structure (Swanson, 1988). Several examples of en echelon dike terminations can be seen in Figures 6 and 7. These dike terminations were interpreted to have developed at the onset of mechanical interaction between the en echelon fracture tips during dilation. This mechanical interaction results in tip-to-tip and tip-to-plane styles of structural development. These structures and observations helped lead to the interpretion of these dikes as having intruded into upward-flaring propagating fracture systems.

Felsic Intrusive Features

Detailed mapping during these earlier investigations also revealed a subtle vertical zone of concentric extensional fracture fabric development within the host rock about the Cape Neddick gabbro complex (Swanson, 1981, 1982). A late phase of rhyolitic magma was injected into this fracture zone. The fracture fabric zone may vary from a subtle fracture cleavage in the Kittery Formation and earlier basalt host rocks to veined fabrics and localized breccia zones upon injection. This zone of near-vertical fracturing may be related to the radial mafic dikes within the gabbro itself and the late doming pulse described earlier by Gaudette and Chapman (1964) as a late phase of steep vertical cone fracturing. Earlier interpretations (Swanson, 1981) considered these structures as vertical fabric zones that would lead, through brecciation and felsic intrusion, to the development of vertical felsic ring dikes and associated xenolithic screens.

Mapping of igneous explosion breccia bodies (Swanson, 1982) at Gerrish Island (first described by Hussey, 1962) also showed an association with felsic rhyolitic melts as sheath structures about the nearly matrix-free vent breccias and as

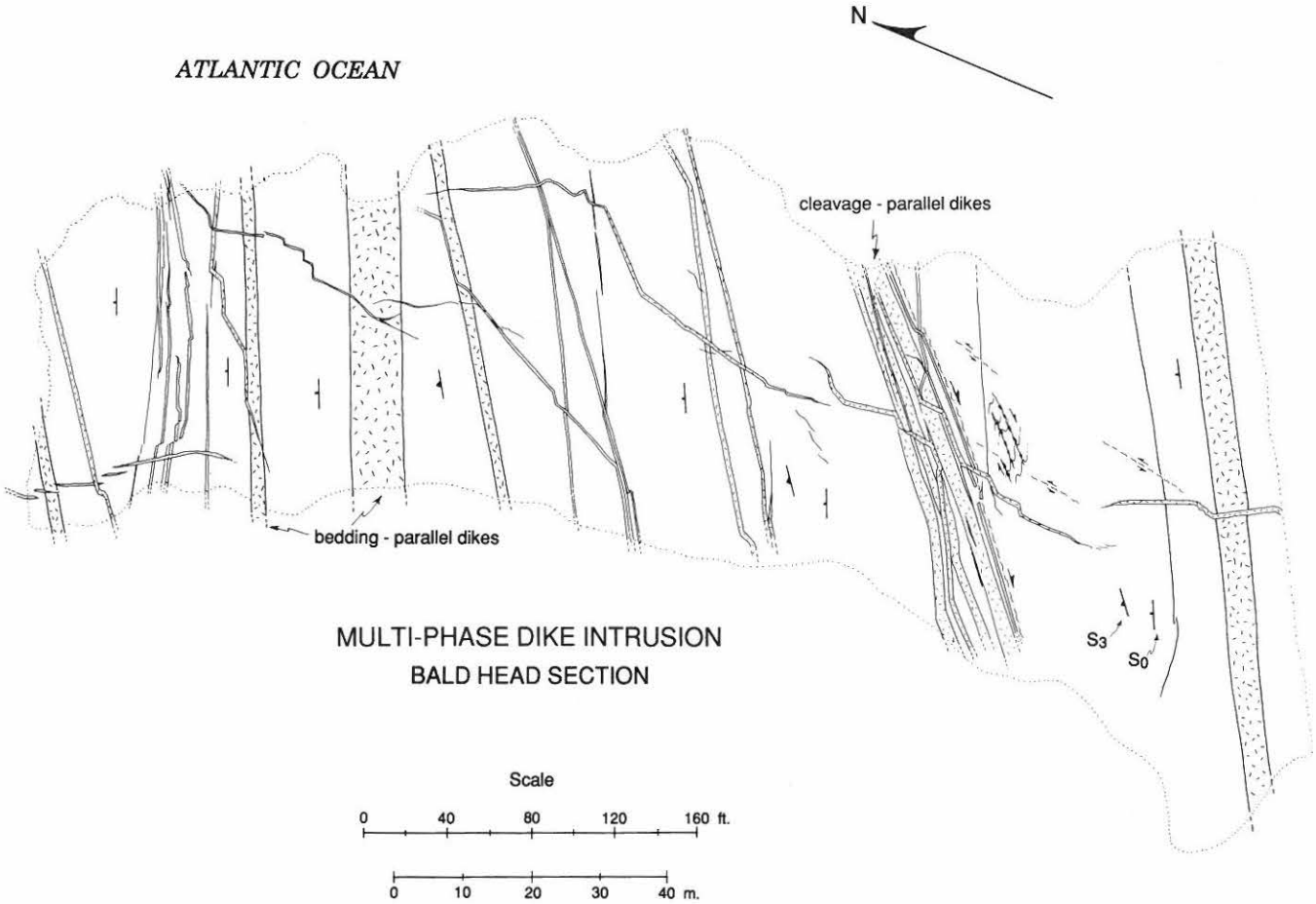

Figure 6. Detailed dike intrusion relations at Bald Head indicating a preexisting structural control on dike segment orientation by a near-vertical N60°E (S_0) bedding-anisotropy and a near-vertical N45°E (S_3) cleavage-anisotropy. Gently-plunging F_2 isoclinal folds are responsible for the bedding-anisotropy. Early dextral shear localized along early mafic(?) dike, moderately-plunging z-shaped asymmetric folds, and reoriented quartz boudins are associated with the slip and axial plane cleavage. Several phases of northeast-trending dikes are present which are crosscut by the younger north-northwest-trending dike phase. All dikes, bedding (S_0), and cleavage (S_3) in the Kittery Formation are steeply dipping to vertical in orientation

individual dikes. Basalt dikes occur as fragments in the breccias and as younger crosscutting dikes. Locally, lamprophyric dikes cut the breccias and associated basaltic dikes and were found to carry an amazing collection of plutonic xenoliths. These plutonic xenoliths range in composition from gabbro and hornblendite to syenite. This assemblage of minor intrusions and structures was interpreted (Swanson, 1982) to represent the near surface expression of a White Mountain-type central intrusive complex (Fig. 2) developed at some depth below the present erosion surface. Both the coast-parallel positive gravity anomaly of Bothner and Harrower (1973) and the recently discovered submarine exposures of quartz syenite, 3.5 kilometers offshore of Gerrish Island reported by Brooks (1986), may be expressions of this largely subsurface central intrusive complex. The coast-parallel gravity anomaly may in part be due to the intensity of dike injection within the coast-parallel dike swarm.

The minor structural and intrusive development in the exposures immediately adjacent to the Agamenticus alkaline syenite complex has been interpreted in terms of host rock deformation related to the intrusive emplacement of the various phases of the Agamenticus complex (Swanson, 1984). An early ductile phase of host rock deformation appears to have been dominated by layer-parallel dextral shear localized within the pelitic layers of the Kittery Formation. This resulted in the development of steeply-plunging, z-shaped asymmetric folds, the N45°E near-vertical oblique cleavage, the reorientation and boudinage of preexisting en echelon quartz veins, and vertical N45°E dextral shear zones, a few of which appear to deform early mafic dikes. Near the more southeasterly contact of the Agamenticus complex, outwardly directed stresses would be oriented perpendicular to the bedding anisotropy where at least one northwest-dipping mylonitic thrust surface (approximately 0.5 meters thick) can be observed along the Cape Neddick

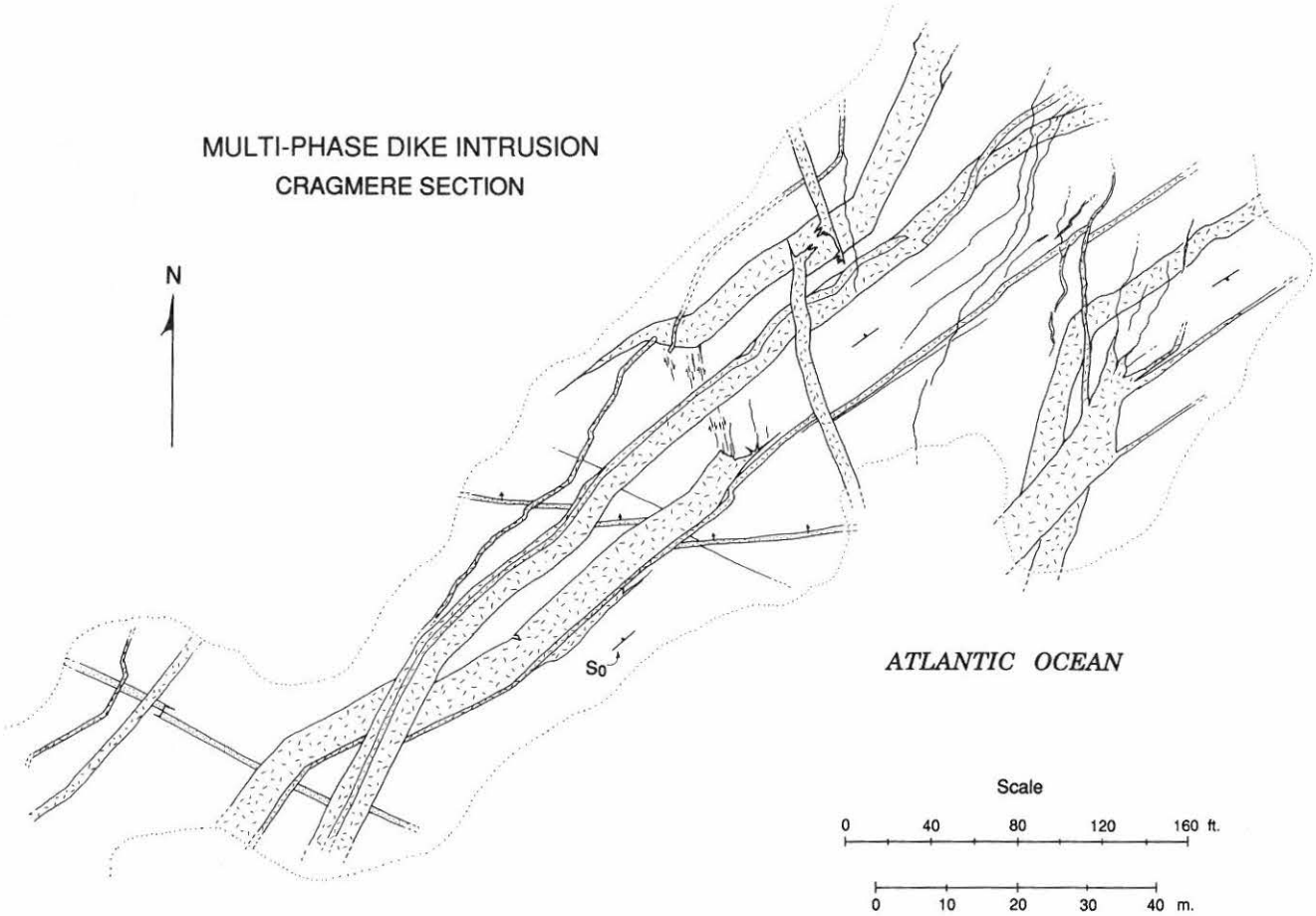

Figure 7. Detailed dike intrusion relations along exposures near Cragmere (1.5 km south of Bald Head) indicating an early aplitic phase of intrusion (stippled with arrows in dip direction), several phases of northeast-trending basalt intrusion and a later phase of north-northwest-trending basalt intrusion (hachured and as single lines). All dikes and bedding (S_0) in the Kittery Formation are steeply dipping to vertical in orientation.

peninsula. Early inclined trachytic sheets and dikes, some of which have been involved in this phase of host rock deformation, suggest that the forceful intrusion and lateral expansion of the central syenite stock at Agamenticus may have been associated with this early phase of deformation. A later brittle phase was dominated by layer-parallel dike intrusion along the preexisting N60°E bedding-anisotropy as well as the N45°E cleavage-anisotropy. Dike intrusion was accompanied by conjugate kink development due to sub-horizontal, layer-parallel compression outward from the central intrusive complex. The development of numerous northeast-trending composite dikes with rhyolitic centers would suggest that this late brittle phase of structural-intrusive development was contemporaneous with the later granitic and quartz syenitic intrusive phases at Agamenticus.

Intrusive Model

A general interpretive model for the style and sequence of intrusion within these exposures was formulated following the linear dike swarm - central intrusive complex model of Sloan et al. (1969) and Speight et al. (1982) and a composite magma chamber model for central intrusive complexes after Chapman (1976) and Walker (1975). The linear dike swarm model of Sloan et al. (1969) and Speight et al. (1982) was based on studies in the Tertiary igneous province of northwest Scotland.

An intricate and complex relationship is thought to exist between the multi-phase dike intrusions of the prominent coast-parallel linear dike swarm and the mafic-felsic minor intrusions associated with the central intrusive complexes at Agamenticus, Cape Neddick, and at Gerrish Island. Some of these relationships are illustrated in the dike map of Figure 7. The expected complexities in dike intrusion relations as a result of this model would include overlapping mafic-felsic injection cycles, multi-phase dike intrusion histories, complex crosscutting relations between dikes of the prominent linear northeast-trending swarm and radial and concentric dikes directly related to the central intrusive complexes, and a complex relationship between host rock deformation, metamorphism, and the history of dike in-

trusion about the central intrusive complexes. The observed trend variations for dikes of the northeast-trending linear dike swarms had been interpreted earlier (Swanson, 1982) as possibly separate linear swarms centered about each of the central intrusive complexes. However, a more continuous swarm structure is now envisioned with trend variations due to preexisting structural controls in the host rocks.

Recent investigations during the 1984 and 1986 field seasons have expanded the detailed base map area. The resulting base maps can now be used for continuing detailed structural analysis and to begin a comprehensive petrographic study to characterize the different intrusive phases and in particular to discriminate amongst the various mafic dikes found in these exposures.

Preliminary results (Swanson, 1986) indicate a sequence of minor intrusions that includes two (or more) main phases of mafic dike intrusion (here referred to as basalt) within the northeast-trending linear dike swarm; an older, gray-weathered, epidotized, feldspar-phyric basalt and a younger, red-weathered, pitted, feldspar and pyroxene-phyric basalt. These main phase dikes are typically 3-8 m in width but many thinner dike segments are exposed. The main phase basalts were preceded by an early peripheral Agamenticus sequence of intrusions that consists of pre- to syn-contact, boudined quartz veins, early dextrally sheared mafic (?) dikes up to 20 cm in width and early blue-weathered trachyte dikes, up to 50 cm in width. All of these early intrusions show evidence of involvement in the local development of early ductile to early brittle structural assemblages related to the host rock deformation described above. Early post-contact intrusions include additional blue-weathered trachytes up to 1.0 m in width, thin low-angle felsic aplitic sheets, generally 1.0 cm in width (but up to 1.0 meter), and early thin, (up to 1.0 meter) fine-grained gray-weathered, finely feldspar-phyric to aphyric basalts, all with a variety of orientations.

The northeast-trending main phase mafic dikes are also followed (with some overlap) by a late peripheral Agamenticus sequence of intrusions that consists of trachytic rhomb-feldspar porphyries up to 2.0 m in width, orange-weathered rhyolite felsites as 1.0 meter-wide dikes and larger irregular inclined sheets, sporadic plutonic-xenolith bearing lamprophyric dikes up to 0.5 m in width, felsic breccia dikes up to 0.5 m in width and a late phase of northwest-trending, rusty-weathered, sulfidic, feldspar-pyroxene-phyric basalts up to 1.0 m in width. The late phase basalts are associated with a final phase of rusty-weathered calcite mineralization in fractures.

Peripheral Cape Neddick intrusions include inclined mafic and felsic (cone) sheets and a late phase of felsic melt that intrudes the concentric, extensional fracture fabric around the intrusion and forms the matrix for some local breccias. Additional dike phases associated with the Cape Neddick complex include xenolithic hornblende lamprophyre dikes (as described by Haff, 1939) and associated intrusive bodies that form the agglomerate phase at Cape Neddick as described by Hussey (1962). Minor intrusions of the Cape Neddick complex clearly crosscut the members of the main northeast-trending basalt dikes.

Minor intrusive phases above and adjacent to the Gerrish Island igneous complex include the explosion breccias and associated felsic rhyolitic melts, felsic breccia dikes and plutonic xenolith-bearing lamprophyres. Breccias of the Gerrish Island igneous complex appear to be contemporaneous with the main northeast-trending linear basalt dike swarm.

Research in progress is concentrating on adding structural details to the subarea mosaic map sheets, detailed field descriptions of compositions and textures for the dike phases, particularly the mafic dikes, sampling of the characteristic dike types, and beginning a comprehensive petrographic and geochemical study. This detailed approach is necessary in order to fully characterize these intrusive phases and their complex interrelationships.

CONCLUSIONS

The Mesozoic dikes of southern coastal Maine, as perhaps one of the best series of dike exposures anywhere in the world, have attracted the attention of geologists for 150 years (Jackson, 1838; Hitchcock, 1861; Kemp, 1890; Keeley, 1914; Powers, 1915; Wandke, 1922a; Haff, 1939, 1941, 1943; Hussey, 1962; Gaudette and Chapman, 1964; Woodard, 1969; Swanson, 1982, 1983, 1984, 1986, 1988, and in progress). The exposures at Bald Head and Cape Neddick, in particular, are truly classic geologic localities serving as the focus of several geologic investigations. As a result of these concerted research efforts the dike types, their orientations and intrusive sequences in these exposures can now be interpreted to represent a complex interaction between the main northeast-trending linear dike swarm, minor intrusions related to the Agamenticus and Cape Neddick central intrusive complexes and preexisting structures in the host rocks. Field relations suggest at least two main phases of intrusion within the northeast-trending mafic dike swarm as well as a variety of minor mafic and felsic intrusions related to each of the central intrusive complexes.

ACKNOWLEDGMENTS

Initial field work in southern Maine was undertaken as part of the author's Ph.D. dissertation at the State University of New York at Albany and was partially supported by a Grant-in-Aid from the Geological Society of America. More recent field work was supported by a faculty fellowship and research grant from the University of Southern Maine. In addition, acknowledgment is made to the donors of the Petroleum Research Fund, administered by the American Chemical Society for support of the current field investigations. Thanks to reviewers Bob Marvinney, Greg McHone, and Marty Ross for their constructive criticism of this manuscript and to H. H. Woodard for use of his unpublished manuscripts in this report.

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Maine Geological Survey
Studies in Maine Geology: Volume 3
1989

Geochemical Aspects of Volcanic Rocks on Islands in East Penobscot Bay, Maine

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ABSTRACT

The volcanic rocks on the Great Spruce Head-Eagle group of islands in East Penobscot Bay, Maine, belong to two different suites. The two suites are differentiated in field mapping on the basis of differences in structural style and metamorphic character and these differences are also brought out by their chemistry. These differences suggest that the two suites are unrelated to each other, have different ages, and have different tectonic affinities.

Rocks of the older suite belong to the North Haven Formation that has an ill-defined age of Ordovician or older. They are dominated by basalts and basaltic andesites with only minor amounts of felsic volcaniclastic rocks and epiclastic sedimentary rocks. Most of their major and trace element data do not distinguish unequivocally among the possible tectono-magmatic settings. The TiO_2 contents of the basalts are in the range for typical ocean-floor basalts. Negative Ta-Nb anomaly in the basalts suggests that they may be part of an immature oceanic arc.

Rocks of the younger suite belong to the Silurian-Lower Devonian Castine Formation. Most of these rocks are flows and tephra of rhyolite and dacite with subordinate pillowed basalt and basaltic andesites. The TiO_2 concentrations of the basalts are intermediate between those of typical ocean floor basalts and basalts from continental rifts. Trace element diagrams that most successfully classify the tectonic setting of the basalts suggest a within-plate origin. The large volume of felsic rocks present and the inferred continental basement for the region are compatible with an ensialic rift setting. However, the large volume of agglomerate in the Castine Formation and the existence of andesites in outlying areas may suggest that at least some of these volcanic rocks could have been formed in an Andean-type plate margin.

INTRODUCTION

The area of this study is situated in the coastal volcanic belt that extends from Massachusetts northeastward into Maine and New Brunswick (Fig. 1). A suite of Silurian and Lower Devonian volcanic and associated sedimentary rocks have spectacular development within this belt (Brookins et al., 1973; Gates, 1975; Brookins, 1976; Shride, 1976; Gates and Moench, 1981; Hon and Thirlwall, 1985). In East Penobscot Bay, Maine (Fig. 2), the Silurian-Lower Devonian volcanic rocks include the Castine Formation, the Thorofare andesite, and the Vinalhaven rhyolite (Smith et al., 1907). These formations lie above the less extensive North Haven Formation (Smith et al., 1907), which also has a volcanic character. Within the area of study, the Castine Formation contains abundant agglomerates, tuffs, rhyolite and dacite, and less abundant basalts, and the North Haven Formation contains abundant basalts and minor dacite, rhyolite, and tephra. This paper focuses on the geologic and

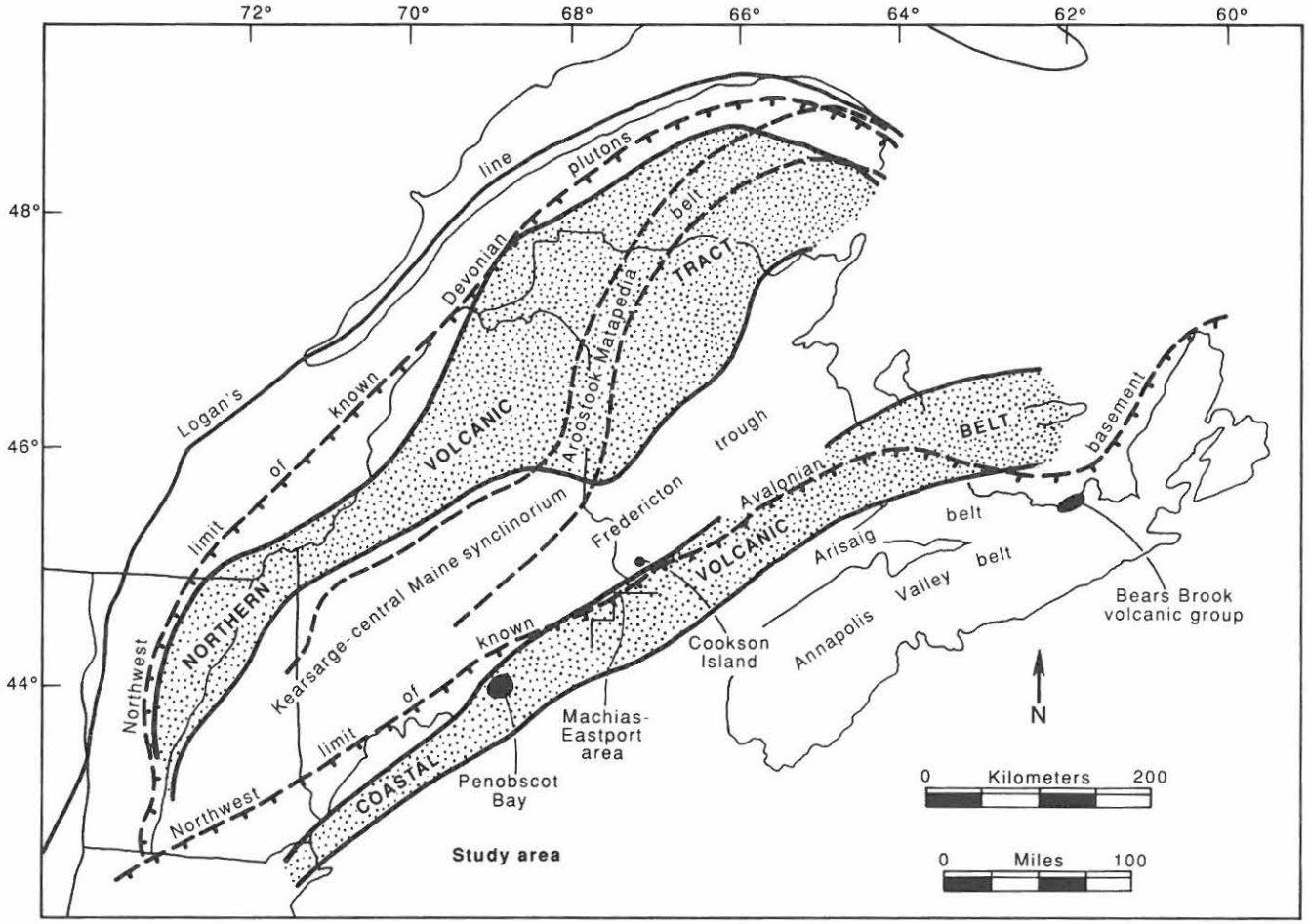


Figure 1. Map of northern New England and Maritime Provinces showing Silurian and Lower Ordovician volcanic tracts and principal paleotectonic features. Adapted from Gates and Moench (1981).

chemical nature of the Castine Formation and the underlying North Haven Formation.

Rocks of both the Castine Formation and the North Haven Formation are metamorphosed to regional and low greenschist facies, but higher grades of thermal metamorphism occur locally around the granitic plutons. Outside of contact aureoles, the Castine Formation exhibits only a single metamorphic event, whereas the North Haven Formation may be polymetamorphic. Volcanic rocks of the Castine Formation carry a metamorphic fabric which is barely perceptible to imperceptible, and they are deformed by a single, prominent fold set. In contradistinction, the volcanic rocks of the North Haven Formation display a well developed foliation and exhibit several sets of minor folds. Both volcanic suites are cut by numerous high-angle faults, and local zones of high strain develop along some of these.

The North Haven Formation is here interpreted to unconformably underlie the Castine Formation. This conclusion is predicated on: (1) the existence of an unconformity at Ames Knob on the island of North Haven, where the Thorofare andesite (equal to Castine Formation) overlies the fossiliferous Ames Knob Formation (Smith et al., 1907), and where a basal conglomerate in the Ames Knob Formation contains clasts of the underlying North Haven Formation; and, (2) on observations that the North Haven Formation is polydeformed and possibly polymetamorphic when compared to the Silurian-Lower Devonian section. The North Haven Formation has no firm age, but permissible ages range from Precambrian to Ordovician. Because of the above relationships, we believe that the North Haven Formation is an earlier volcanic suite, possibly unrelated to the Castine Formation. However, it must be noted that Bouley et al. (1981) interpreted the unconformity at Ames Knob as a local feature, and they included the rocks of both formations in the same magmatic suite.

Basement to these volcanic suites is not exposed in the area of study. Although the North Haven Formation could be a fragment of oceanic basalt, regional considerations suggest that basement with continental affinities underlies both the North Haven Formation and the Castine Formation. In southern New Brunswick, McCutcheon and Ruitenberg (1987) interpret the Silurian section, which includes volcanic rocks similar to the Castine Formation, to rest unconformably on Precambrian granite. In East Penobscot Bay, Maine, the volcanic rocks of the

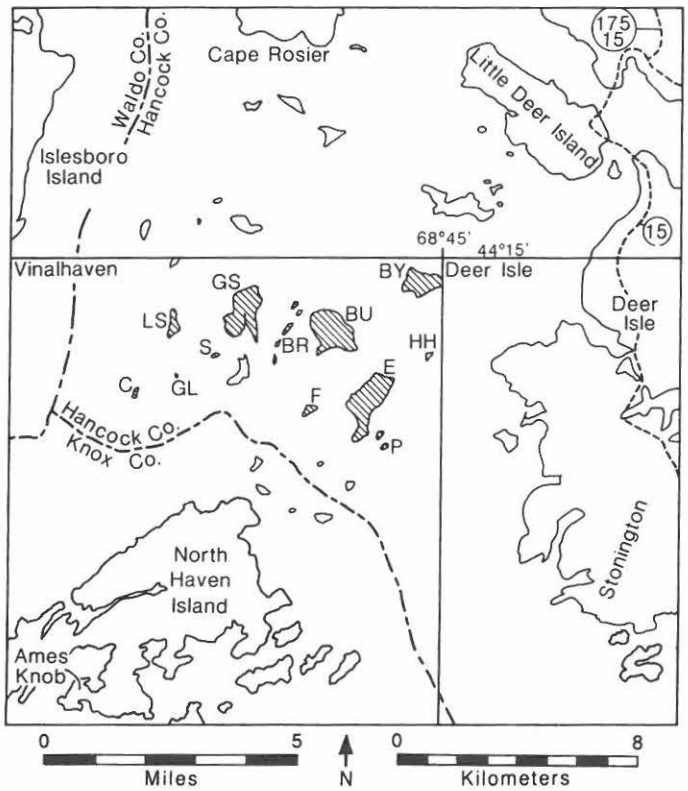


Figure 2. Location map of East Penobscot Bay, Maine. The islands included in this study are diagonally ruled and labelled as follows: Barred Islands (BR), Bradbury (BY), Butter (BU), Compass (C), Eagle (E), Fling (F), Grass Ledge (GL), Great Spruce Head (GS), Hard Head (HH), Little Spruce Head (LS), Porcupines (P), Scragg (S).

Castine and North Haven Formations contact the Islesboro sequence along steeply dipping faults (Stewart, 1974). David B. Stewart (pers. commun., 1985) has reported dikes cutting the Islesboro sequence that he believes contain rocks related to the Castine Formation. These relationships suggest that the rocks of the Castine Formation had stratigraphically overlain the Islesboro sequence before their erosion, and further suggest that the Islesboro sequence underlies the current outcrop of the Castine Formation. The Islesboro sequence contains metamorphosed volcanic rocks, schist, and marble that are cut by pegmatitic dikes dated at 600±15 Ma (Brookins, 1982). Younger sequences contain quartzite, quartzose phyllite, marble, and conglomerate. These are undated, but may have ages ranging from Late Precambrian into the Cambrian. These rocks represent a cover sequence that almost certainly has continental affinities.

The rocks of the coastal volcanic belt are thought to represent the western margin of the Avalon plate prior to the closing of an ocean basin (Osberg, 1974; see also Berry and Osberg, in press). Pridolian through Gedinnian fossils having Rhenish-Bohemian-Baltic provinciality (Johnson, 1979; Berdan, 1983) support this idea. However, there is little consensus amongst researchers regarding the nature of the suture separating Avalonia from North America.

The origin of the Silurian-Devonian volcanic rocks in the coastal belt must be considered to be equivocal, perhaps because of an insufficient data base or the presence of more than a single suite of volcanic rocks. McKerrow and Ziegler (1971), Osberg (1978), and Bradley (1983) interpreted the Silurian-Lower Devonian volcanic rocks as having been generated near a subduction plate margin. On the other hand, Ruitenberg and Ludman (1978) and Gates and Moench (1981) noted that the volcanic rocks located in the vicinity of Eastport and adjacent New Brunswick had a persistent and strongly bimodal character. On this basis, Gates and Moench (1981) suggest that these rocks are more characteristic of continental-rift volcanic rocks than they are of subduction-generated volcanic rocks. At the southern end of the coastal volcanic belt, Shride (1976) described a section that contains abundant andesite. Hon and Thirlwall (1985), on the basis of the chemistry of those rocks, interpreted them as having island-arc affinities.

This paper incorporates geological observations and data, laboratory analyses, and interpretations presented in Pinette (1983). The area of study includes the following islands located in East Penobscot Bay: Barred Islands, Bradbury, Butter, Compass, Eagle, Fling, Grass Ledge, Great Spruce Head, Hard Head, Porcupine Islands, Little Spruce Head, and Scragg (Fig. 2).

STRATIGRAPHY AND LITHIC DESCRIPTIONS FOR THE EAST PENOBSCOT BAY AREA

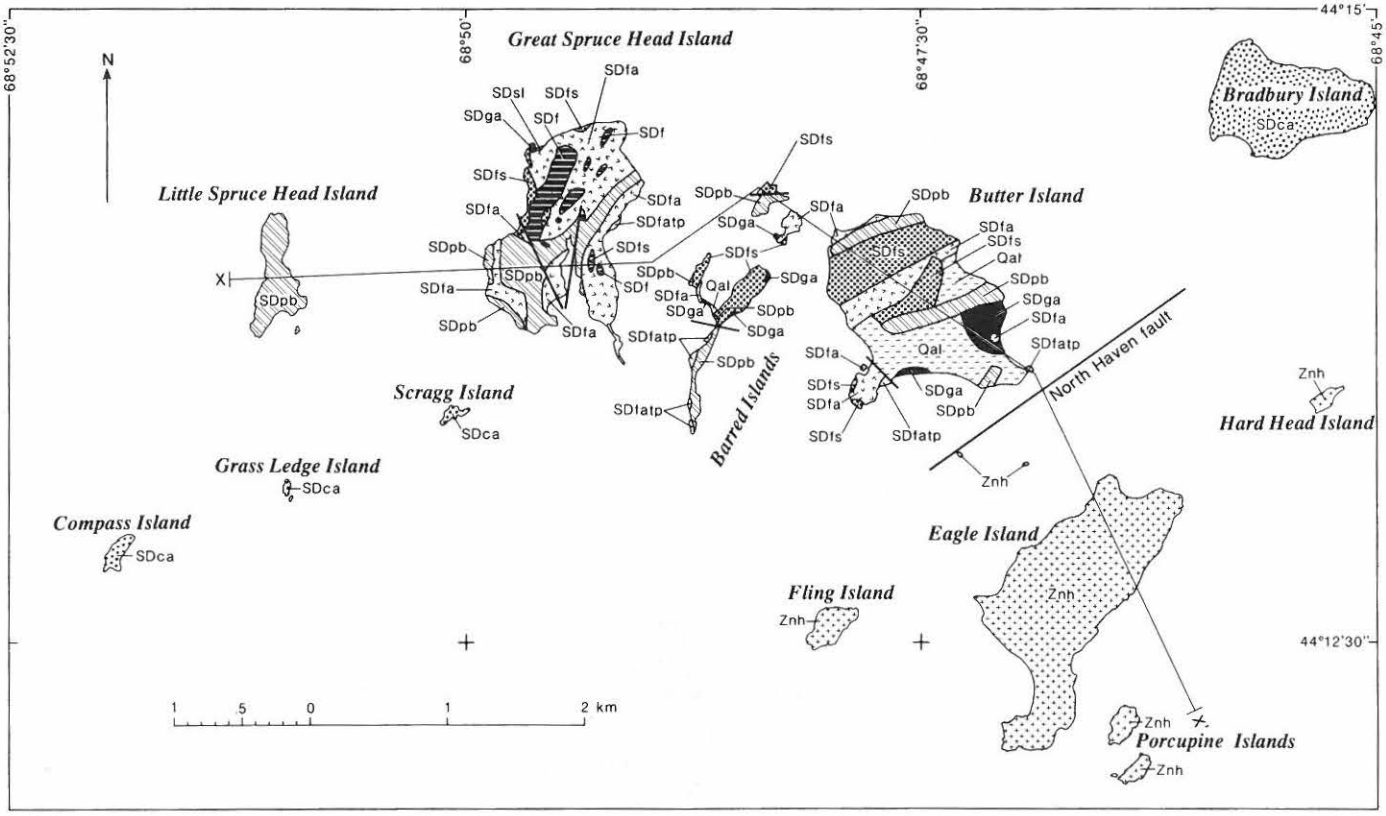
General Statement

Figure 3a shows the general geological features of this part of Penobscot Bay. The volcanic rocks exposed in this area belong to two suites: an older suite correlated with the North Haven Formation (Smith et al., 1907), and a younger suite included in the Castine Formation (Smith et al., 1907) as redefined by Stewart (1956) and Wingard (1961). Figure 4 presents a structural section for the rocks described in this report. A stratigraphic column for the study area is presented in Figure 5.

Displacements along high-angle faults disrupt the local stratigraphy, and in this region a major high-angle fault, the North Haven fault, is thought to separate the Castine and North Haven Formations. Upright folds on some of the outlying islands indicate that the Castine Formation has been deformed by large-scale folds. Observations of minor folds and schistosity suggest that the North Haven Formation is polydeformed.

North Haven Formation

Rocks belonging to the North Haven Formation crop out on Eagle, Fling, the Porcupines, and Hard Head Islands. These rocks are predominantly mafic (>90%), principally basalts and basaltic andesites. Relict pillows and mafic agglomerate are locally preserved, but much of the rock has been recrystallized



LITHOLOGIC SYMBOLS

Quaternary

- Qal — Undifferentiated alluvium and glacial deposits

High-angle fault

• E33 — Outcrop sampled for chemical analyses

CASTINE FORMATION

Silurian - Devonian

Eruptive Rocks

- SDca — Undifferentiated Castine formation lithology; reconnaissance mapping only
- SDpb — Dominantly pillowed basalts and basaltic andesites
- SDfatp — Felsic lapilli tuff-agglomerate with tuffaceous to phyllitic matrix, may be in part spherulitic
- SDfa — Felsic agglomerate with vague clasts
- SDfs — Felsic unit with abundant spherulites
- SDf — Felsic unit weathering white and lacking any field-evident distinguishing textural features

Intrusive Rocks

- SDga — Fine- to coarse-grained mafic (gabbroic ?) intrusion

Sedimentary Rocks

- SDsl — Laminated siltstone unit occurring with limestone unit

NORTH HAVEN FORMATION

Late Precambrian to Ordovician

- Znh — Undifferentiated sequence of metabasalts interbedded with relatively minor felsic volcaniclastics and very minor sedimentary rocks

Figure 3a. Generalized geologic map of study area.

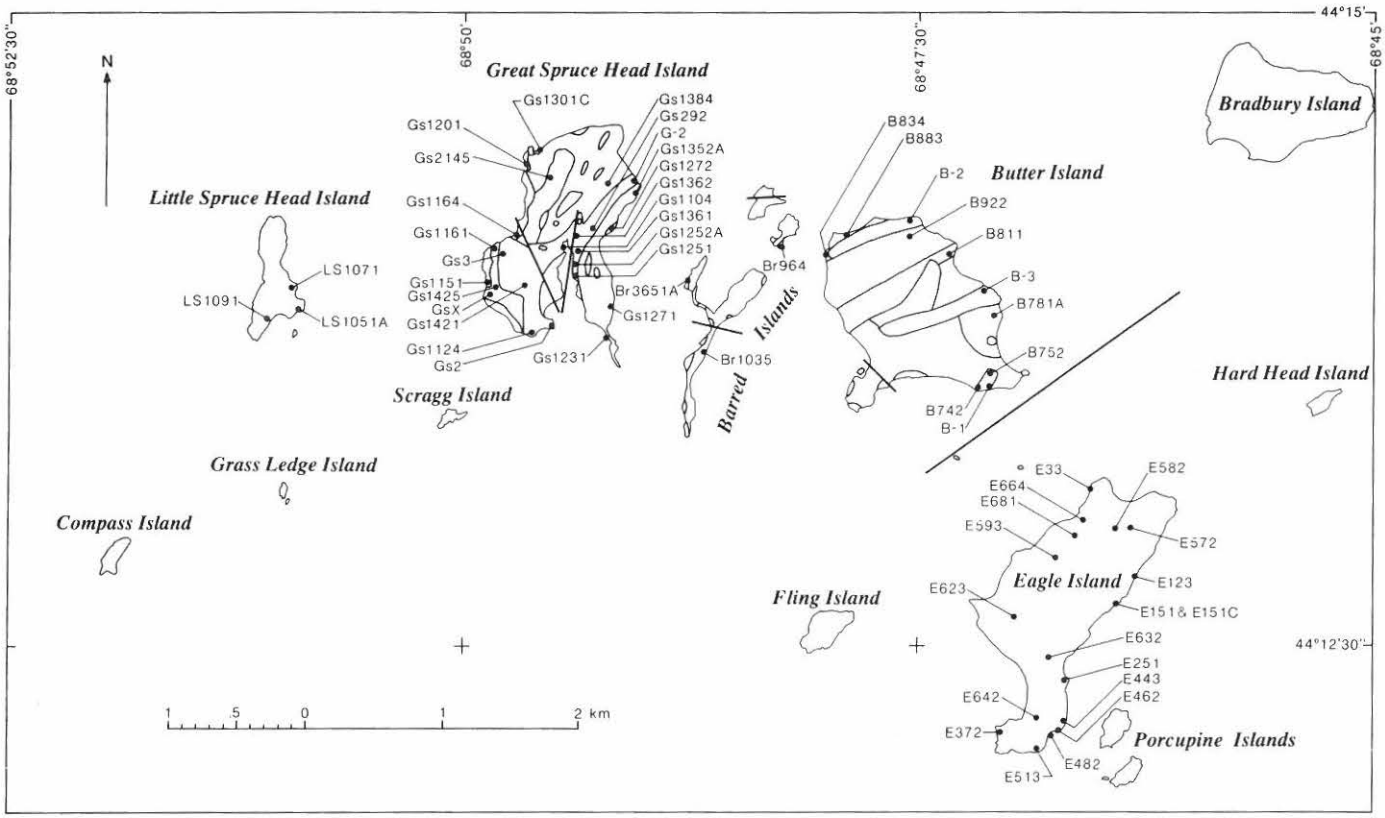


Figure 3b. Locations of chemically analyzed rock samples (see Table 1).

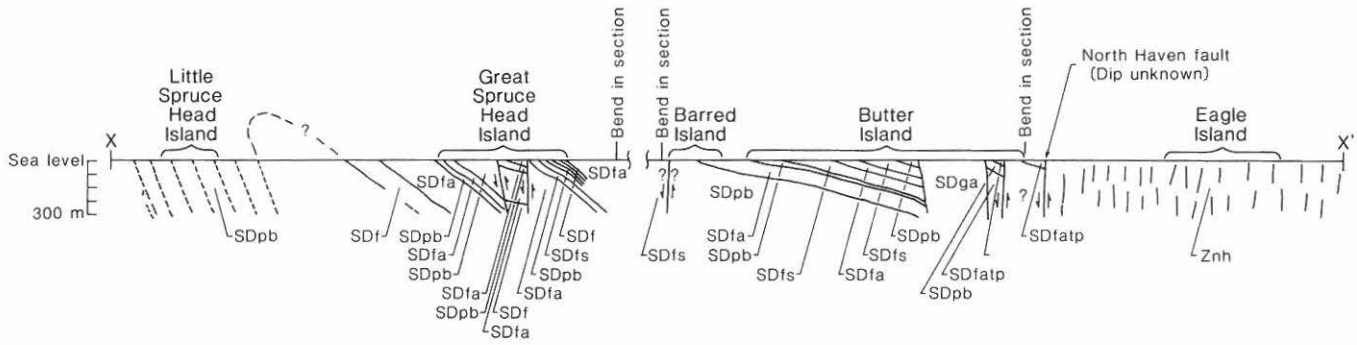


Figure 4. Stratigraphic cross-section of the study area. Symbols and scale are same as in Figure 3a.

to mafic phyllites. Felsic tuffs, felsic lapilli tuffs, and felsic lapilli tuff-agglomerates form layers one to tens of meters thick that are interbedded with the mafic volcanic rocks. These rocks, classified chemically as dacites and alkali rhyolites, make up a small volume of the formation (<10%). Minor amounts of slightly metamorphosed mudstone, graywacke, and quartzofeldspathic wacke are also present. Bedding in these rocks, where identified, is steeply dipping.

The rocks are thoroughly recrystallized and contain chlorite with two different compositions, possibly signifying polymetamorphism. In addition, they contain a pervasive foliation which, in outcrops displaying both bedding and foliation, is essentially parallel to bedding. Isoclinal upright folds with two different orientations deform both bedding and foliation. Small asymmetrical kink folds postdate the other structural features. The features taken together suggest a complicated deformational history for rocks of the North Haven Formation.

Castine Formation

The Castine Formation is exposed on Bradbury, Butter, Compass, Grass Ledge, Great Spruce Head, and Scragg Islands,

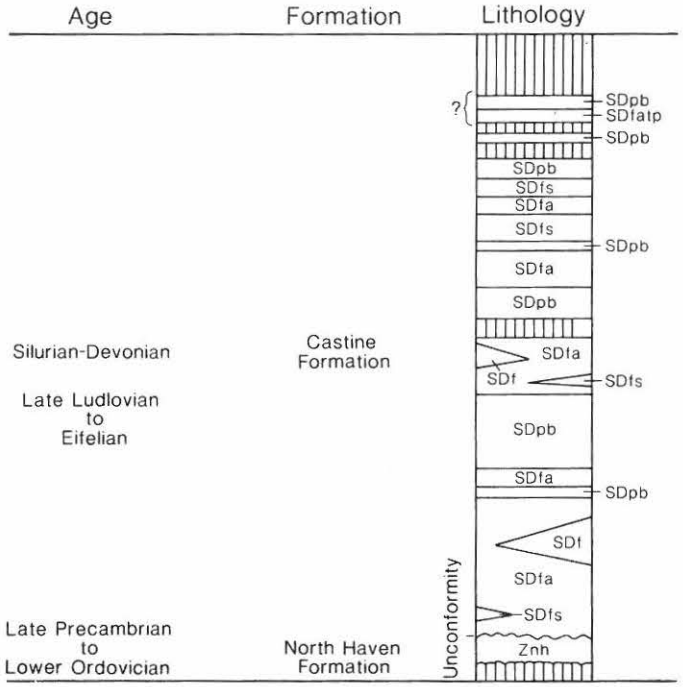


Figure 5. Stratigraphic column of the study area. Symbols same as in Figures 3a and 3b.

and on the Barred Islands. Unlike the North Haven Formation, the Castine Formation is dominated by felsic lavas and volcaniclastic rocks (~70%); mafic lavas are subordinate (~30%). Chemically, the felsic rocks are dacites and alkali rhyolites. These rocks vary spatially in their textural characteristics. Locally they are massive and featureless; at other locations they are flow-banded, spherulitic, or agglomeratic. Felsic tephra deposits include ash tuffs, lapilli tuffs, and lapilli tuff-agglomerates. Most felsic units, except for the fine tephra, occur as distinctive 1-3 meter thick beds. The mafic rocks are basalts and basaltic andesites. Mafic flows commonly are pillowed, although locally mafic pillow agglomerates and lapilli tuff-agglomerates are associated with them. Mafic units are from 2-4 meters thick. Minor amounts of siltstone and limestone are intercalated with the volcanic rocks. Rapid lateral and vertical lithofacies changes and great variability in the volcanic assemblage suggest several nearby sources and some topographic influence on the distribution of these rocks (Stewart and Wones, 1974). In general, the rocks of the Castine Formation are slightly recrystallized, but they do not carry a pervasive foliation.

Rocks of the Castine Formation as exposed on Great Spruce Head, Butter, and the Barred Islands have uniformly gentle dips toward the east. Observation of facing directions, such as pillow tops, lapilli tuff tops, and brecciated tops of flows, without exception indicate that this section is right-way up and essentially homoclinal. The local stratigraphic section shown in Figure 4 incorporates these observations.

Although the Castine and North Haven Formations are juxtaposed across the North Haven fault in the area of this study, regionally the Late Silurian-Early Devonian section, including the Castine Formation, is interpreted to lie unconformably on the North Haven Formation. The stratigraphic relationship at Ames Knob on the island of North Haven, where the Silurian-Early Devonian section rests unconformably on the North Haven Formation, supports this interpretation. Randomly oriented, foliated clasts of the underlying North Haven Formation have been incorporated into the basal conglomerate of the Early Silurian Ames Knob Formation. The randomly oriented, foliated clasts and the absence of a penetrative foliation in the matrix of this conglomerate indicate that the foliation developed in the North Haven Formation before erosion and burial of the Early Silurian rocks.

Fossil evidence found on nearby islands outside of and on Compass Island within the study area indicates that the Castine Formation spans at least Late Silurian through Early Devonian time. Fossil fauna in the Ames Knob Formation on the island of North Haven range from Late Llandoverian to Pridolian in age (Smith et al., 1907; Gates, 1969; Berdan, 1971; Brookins et al., 1973). Therefore, the maximum age for the overlying Thorofare andesite and the equivalent Castine Formation must be Pridolian. Brookins et al. (1973) found ostracodes in a limestone interbedded with pillowed andesite on the southwestern corner of Compass Island which they interpreted to range in age from Early Devonian to Eifelian. Thus, the age of the Castine Formation can be bracketed between Late Silurian (Late Ludlovian or Pridolian) and Eifelian.

Intrusive Rocks

Small bodies of gabbro intrude both formations, and two types of mafic dikes and sills are present in the study area. Gabbros, having idiomorphic texture, are massive and unfoliated. One type of mafic dike/sill intrudes only the Castine Formation. It has aphanitic to fine porphyritic texture with feldspar laths up to 1 mm long. Some of these dikes are multiple, and their spatial relations to pillow basalts in the Castine Formation suggest the possibility that they are feeder dikes for the basalts. The second type of mafic dike is similar to the first, but fresher. It also has an aphanitic texture, but in addition it possesses well developed columnar joints. D. B. Stewart (pers. commun., 1982) reports that one of these dikes located on Great Spruce Head Island contains analcime, and he believes it to be Mesozoic.

PETROGRAPHIC DESCRIPTION

General Statement

Petrographic descriptions of rocks in the North Haven and Castine Formations and in several cross-cutting bodies of gabbro

are given below. Compositions of feldspar are mainly determined by extinction angles.

North Haven Formation

Microscopically, the mafic phyllites and pillow basalts, locally amygdaloidal, contain phenocrysts of plagioclase with pilotaxitic texture and augite. These are set in a matrix which originally must have been finely crystalline or glassy. The plagioclase consists of albite microlites and laths as long as 3 millimeters. These crystals must have been completely recrystallized from more calcic feldspar. In thoroughly schistose rocks, albite has a resorbed outline and shows little evidence of twinning. In many rocks inclusions of calcite, epidote, and sericite in plagioclase occur as alteration products of the original feldspar. Locally, acicular mineral aggregates tentatively identified as pumpellyite also occur within the albite. In this regard, D. B. Stewart (pers. commun., 1982) has microprobe data to confirm the presence of pumpellyite in similar rocks just to the north of the study area. Augite crystals ($Z\wedge C = 43^{o}$, and 2V (+) = ~50°) with ophitic to subophitic and intergranular textures range to 3 mm in prismatic section. It may occur as relatively unaltered grains, or as barely perceptible cores which are largely replaced by actinolite, chlorite, and opaque pseudomorphs. Accessory minerals in the mafic volcanic rocks include opaques and sphene, commonly altered to leucoxene. Much of the groundmass consists of fine-grained albite, chlorite, and actinolite. Parallelism of chlorite and actinolite defines a pronounced bedding plane metamorphic foliation. Chlorites with two different compositions are present in some rocks, possibly indicative of polymetamorphism. In these rocks, chlorite with a berlin-blue interference color (Fe-rich; Albee, 1962) overprints chlorite with an anomalous brown interference color (Mg-rich; Albee, 1962).

The felsic tephra of the North Haven Formation consists of alkali feldspar fragments, commonly exhibiting trachytic alignment, set in a groundmass of microscopic to submicroscopic, interlocking, anhedral feldspar and quartz. Sericite is locally an alteration product of the larger feldspar crystals, and it also occurs along with sparse actinolite, chlorite, and opaques in the matrix. Both Fe-rich and Mg-rich chlorite may be present. Parallelism of sericite and actinolite defines a metamorphic foliation in these rocks.

Castine Formation

Pillow basalts, which locally carry amygdules, and basalts contain small phenocrysts and a fine-grained to submicroscopic groundmass. Laths of plagioclase, up to 2 millimeters long, have trachytic development. Plagioclase compositions vary between albite and andesine. Minor saussuritization of the plagioclase is evident from the presence of fine-grained inclusions of epidote, calcite, sericite, and possibly pumpellyite in plagioclase. Calcic augite (B. A. Bouley and R. L. Barnett, unpub. microprobe data), occurring in grains measuring as much as 2.5 mm in prismatic section, has ophitic to subophitic and intergranular texture. The augite is relatively unaltered, but locally grains are rimmed with thin alteration coronas of chlorite and actinolite. The groundmass consists of microscopic to submicroscopic grains of plagioclase and lesser amounts of actinolite and chlorite. Sphene and opaques are distributed throughout the groundmass. All of the chlorite in these mafic rocks is Fe-rich in contradistinction to those in the North Haven Formation. Most of the basalts do not possess a penetrative mineral foliation; an exception being the basalts exposed on Little Spruce Head Island (Fig. 2), where a well developed foliation along with rotated crystals of pyrite suggests that these structures formed in a localized zone of high strain.

The rhyolites and dacites consist of microporphyritic, partially resorbed albite crystals having a trachytic development and set in a fine-grained to submicroscopic felsitic matrix. The matrix consists of varying amounts of fine-grained felty albite with intergranular, microscopic to submicroscopic, interlocking, anhedral quartz and feldspar. Minor amounts of opaques, probably magnetite, are disseminated throughout the rock. Minor sericite is the only obvious alteration product. Felsic ash tuffs and lapilli tuffs have similar mineralogy to the rhyolites and dacites. The main difference is the presence of fragmental feldspars scattered throughout the rock. The alignment of sericite flakes produces a subtle foliation in all of the felsic rocks.

Intrusive Gabbroic Rocks

The gabbros have a fine- to medium-grained idiomorphic texture. Phenocrystic plagioclase and augite are set in a dense groundmass of fine-grained anhedral to subhedral plagioclase and pyroxene. Albite occurs as euhedral laths measuring up to 5 mm long. Sericite ranges from minor to important as an alteration product within the plagioclase. Augite ($Z\wedge C = 43^{o}$ and 2V (+) = 50°) occurs as salmon-colored euhedra measuring as much as 3 mm in prismatic section. Minor actinolite and Fe-rich chlorite are alteration products of the augite. Magnetite and apatite are ubiquitous accessory minerals. At one locality on Eagle Island (Fig. 2) basaltic hornblende appears as an additional phase.

ANALYTICAL TECHNIQUES

Sample preparation for the determination of seven major element oxides excluding silica was done by HF-aqua regia-boric acid aluminosilicate digestion modified from Buckley and Cranston (1971). Sample preparation for the determination of silica used a technique modified after Shapiro (1967). Major element oxide analyses for 57 rock samples were performed by one of the authors (S.R.P.) on a Perkin-Elmer 703 Atomic Absorption Spectrophotometer according to the techniques outlined in Perkin-Elmer (1976). Precision and accuracy determinations for the major element oxide concentrations in these samples were

performed by replicate analyses of rock standards GSP-1, AGV-1, and BCR-1 (Flanagan, 1967). Major element oxide and trace element analyses for 6 of these samples were done by the U.S. Geological Survey, and for these the major element oxides were determined according to the procedure described in Shapiro (1975). Four samples analyzed by the U.S. Geological Survey are duplicates of samples analyzed by Pinette, and they were used to compare the accuracy of the two major element oxide analytical techniques. Trace elements for the 6 basalt samples analyzed by the U.S. Geological Survey were done by either X-ray spectroscopy or instrumental neutron activation techniques. The rare-earth element concentrations in these samples were determined by instrumental neutron activation analysis.

CHEMICAL ALTERATION

Rocks of the North Haven and Castine Formations have been variably metamorphosed, developing new mineral assemblages from the original assemblages. Thin section study indicates that the mafic rocks are more susceptible to this recrystallization than are the felsic rocks. The compositional and textural characteristics of chlorite in greenstones of the North Haven Formation suggest that those rocks have been subjected to two metamorphic events, whereas the mafic rocks of the Castine Formation have been recrystallized in a single event. The mineral assemblages observed in all of these rocks are compatible with the pumpellyite-actinolite subfacies.

Cann (1969), Pearce and Cann (1973), Vallance (1974), Dostal and Strong (1983), and Murphy and Hynes (1986) have documented changes in the original chemistry of volcanic rocks by metamorphic processes. Almost all major and trace elements can migrate under different sets of conditions. The processes of albitization, chloritization, and epidotization indicate the complexity and the diversity of possible chemical changes.

A major problem in studying ancient volcanic rocks such as those in the coastal belt of Maine is to obtain chemical parameters that are unaffected by halmyrolysis, metamorphism, and/or weathering and serve as a guide to their original compositions. This problem can never be solved satisfactorily. However, to minimize the effects of weathering, we have chosen for chemical analysis only samples that showed no evidence of weathering. Samples were also examined in thin section to determine the extent to which the original mineralogy was recrystallized to metamorphic minerals. It must be noted that all of the samples exhibited a degree of new mineral growth, but only the least recrystallized were chemically analyzed. In addition, two criteria were employed to assess how well the chemical data reflects the original compositions. In one of these criteria, proposed by Irvine and Baragar (1971), rocks with $H_2O + CO_2 > 2.5$ weight percent are considered to be chemically altered. In our analyses the sum of H_2O and CO_2 was determined by loss on ignition, and, consequently, could be understated. In any case, many of the analyses do not meet this test. In the second criterion, the weight percent of CaO is compared to the sum of those of $Na_2O + K_2O$. Erratic and low concentrations of CaO relative to erratic and high concentrations of $Na_2O + K_2O$ suggest lime depletion and/or alkali enrichment (Pamic, 1974; Vallance, 1974). Many of our analyses show these relationships, again indicating that they have been chemically altered.

The major element oxide data for the volcanic rocks analyzed in this study are shown in Table 1, and the solid symbols in Figure 6 represent what we interpret to be the least altered compositions of basalt and basaltic andesite. Diagnostic criteria to identify chemical alteration in felsic rocks are less clear, and therefore, no attempt was made to identify the least altered felsic rocks in Figure 6.

CHEMICAL PETROLOGY OF THE VOLCANIC ROCKS

General Statement

Major element analyses for the least altered volcanic rocks are presented in Table 1. Eighteen analyses are from the North Haven Formation, and 45 are from the Castine Formation. The analyses represent the important lithologies that constitute these formations, but the number of analyses for each lithology should not be considered to be statistically related to their abundances. Sample locations are plotted in Figure 3b. North Haven Formation

Basalts, high alumina basalts, and basaltic andesites (nomenclature of Taylor, 1969) constitute >90% of the rocks of the North Haven Formation; most of the remainder are rhyolites and dacites. Cation abundances indicate that nearly half of these basalts are olivine and hypersthene normative, whereas the others are olivine and nepheline normative. According to the normative classification of Irvine and Baragar (1971) these basalts are undersaturated and alkaline, but because many samples show some evidence of alkali mobility during alteration, their normative mineralogies may be suspect. If spilitization was part of the alteration process, alkali enrichment would increase the amount of normative plagioclase and, thereby, lock additional silica up in the plagioclase. This depletion of the total silica tends to produce undersaturation in at least some of the remaining normative minerals.

Major element oxides are plotted on Harker-type variation diagrams (Fig. 6). The preponderance of basalts and basaltic andesites (SiO_2 <55%), the lack of intermediate rocks (SiO_2 = 55-62%) and the paucity of siliceous rocks (SiO_2 >62%) are evident from these diagrams. Many basalts have high concentrations of Al_2O_3, MgO, Na_2O, and K_2O, and low concentrations of CaO relative to the basaltic averages posted in Cox et al. (1979). These concentrations, along with the scatter of these oxides, suggest a degree of alteration of the samples, possibly involving spilitization. The concentration of major elements in the two felsic rocks are comparable to the average oxides from their Cenozoic counterparts, although SiO_2 is higher by nearly 3 weight percent in the rhyolite, and Al_2O_3 is lower by nearly

TABLE 1. WHOLE-ROCK MAJOR ELEMENT OXIDE ABUNDANCES.

$FeO^* = FeO + 0.9/Fe_2O_3$; $Mg\# = [MgO/(MgO+FeO^*)] \times 100$. Fe_2O_3/FeO standardized at 0.25; B, Br, E, Gs (G also), and LS represent Butter, Barred, Eagle, Great Spruce Head and Little Spruce Head, respectively.

North Haven Formation

	E33	E123	E151	E151C	E251	E372	E443	E462	E482
SiO_2	49.3	47.7	51.1	54.7	48.9	47.6	47.7	49.8	47.9
Al_2O_3	16.2	13.4	21.5	18.9	15.3	17.2	15.9	15.6	14.4
FeO*	9.1	10.6	6.3	5.6	10.1	8.7	10.0	9.7	10.0
MgO	7.3	14.1	2.8	3.9	8.0	8.3	9.7	6.9	10.0
CaO	6.2	4.4	4.7	3.1	8.0	6.6	3.9	6.4	7.6
Na_2O	4.6	3.3	5.2	8.3	3.6	4.6	3.8	3.0	2.8
K_2O	1.0	0.4	3.3	0.3	0.7	0.6	1.8	2.2	1.3
TiO_2	1.7	1.2	0.9	0.8	1.3	1.4	1.3	1.4	1.2
LOI	2.5	2.8	1.7	3.3	2.2	3.7	3.5	2.2	2.4
Total	97.9	98.2	97.5	98.9	98.1	98.7	97.6	97.2	97.6
FeO*/MgO	1.2	0.8	2.3	1.4	1.3	1.1	1.0	1.4	1.0
Mg#	44.5	57.1	30.7	41.1	44.2	48.8	49.2	41.5	50.0
Rock type	Hi-Al Bas.	Bas.	Hi-Al Bas.	Hi-Al Bas. Andes	Bas.	Hi-Al Bas.	Bas.	Bas.	Bas.

	E513	E572	E582	E593	E623†	E632	E642	E664	E681
SiO_2	48.9	48.6	47.5	48.3	48.1	48.7	66.8	74.6	48.0
Al_2O_3	16.0	15.6	17.0	17.5	16.6	16.0	13.5	13.3	18.4
FeO*	9.9	9.5	9.6	9.4	9.2	9.7	5.3	1.4	9.8
MgO	5.2	7.4	7.0	7.3	8.3	8.6	2.8	0.3	8.9
CaO	5.1	5.9	7.0	7.4	8.7	7.7	0.2	0.0	5.1
Na_2O	5.9	4.2	4.8	4.2	3.5	4.0	4.7	3.6	3.8
K_2O	1.0	0.7	0.5	0.7	0.4	0.3	2.3	3.2	0.6
TiO_2	1.9	1.7	1.8	1.5	1.3	1.1	0.2	0.1	1.5
LOI	2.9	3.3	2.4	2.8	2.6	2.1	2.1	0.9	3.3
Total	97.4	96.9	97.6	99.2	99.0	98.5	98.1	97.4	99.4
FeO*/MgO	1.9	1.3	1.4	1.3	1.1	1.1	1.9	4.7	1.1
Mg#	34.4	43.8	42.2	43.7	47.4	47.0	34.6	17.6	47.6
Rock type	Hi-Al Bas.	Bas.	Hi-Al Bas.	Hi-Al Bas.	Hi-Al Bas.	Hi-Al Bas.	Dac.	Rhyol.	Hi-Al Bas.

Castine Formation

	LS1051A	LS1071	LS1091	G-2†	Gs-2†	Gs-2	Gs3	Gs292	Gs1104	Gs1124	Gs1151	Gs1161
SiO_2	41.7	42.4	46.5	47.9	51.9	50.9	45.6	46.7	66.8	53.4	57.1	68.1
Al_2O_3	17.1	20.8	17.2	15.9	15.0	15.6	15.4	16.7	16.0	18.8	15.3	12.7
FeO*	14.0	11.1	10.9	10.0	9.2	9.5	10.8	10.7	2.5	9.4	7.7	5.5
MgO	1.9	4.2	9.1	6.3	4.8	5.0	6.5	5.3	0.4	4.1	6.2	2.4
CaO	9.0	5.5	4.5	9.3	7.9	6.5	9.3	8.0	0.2	1.6	1.6	2.4
Na_2O	4.8	4.6	3.9	3.6	5.7	4.6	2.8	5.4	5.4	6.7	5.1	1.7
K_2O	6.5	4.3	1.2	0.8	0.2	0.5	0.5	0.2	5.4	0.5	0.1	4.1
TiO_2	1.8	1.1	1.7	1.6	2.3	1.9	1.7	2.0	0.2	1.0	0.9	0.6
LOI	2.5	6.0	3.2	3.4	3.0	2.5	2.5	2.9	1.2	2.7	3.4	1.6
Total	99.3	100.0	98.2	100.0	101.0	97.0	97.7	97.9	98.1	98.2	97.8	99.1
FeO*/MgO	7.4	2.6	1.2	1.6	1.9	1.9	1.7	2.0	6.3	2.3	1.2	2.3
Mg#	11.9	27.5	45.5	38.6	34.0	34.3	37.6	33.1	13.8	30.3	44.6	30.3
Rock type	Hi-Al Bas.	Hi-Al Bas.	Hi-Al Bas.	Bas.	Bas.	Bas.	Bas.	Hi-Al Bas.	Dac.	Bas. Andes	Low-K Andes	Rhyo-Dac.

† indicates U.S.G.S. duplicate analyses.

TABLE 1. (CONTINUED)

$FeO^* = FeO + 0.9/Fe_2O_3$; $Mg\# = [MgO/(MgO+FeO^*)] \times 100$. Fe_2O_3/FeO standardized at 0.25; B, Br, E, Gs (G also), and LS represent Butter, Barred, Eagle, Great Spruce Head and Little Spruce Head, respectively.

Castine Formation

	Gs1164	Gs1192	Gs1201	Gs1231	Gs1251	Gs1252A	Gs1271	Gs1272	Gs1301C	Gs1352A	Gs1361
SiO_2	74.4	54.2	48.1	69.9	49.7	70.2	70.8	72.2	71.2	74.0	49.7
Al_2O_3	12.8	18.9	15.6	12.1	14.9	14.2	14.6	14.4	14.0	12.1	16.3
FeO*	0.9	7.4	10.2	4.0	9.8	3.2	3.0	2.3	3.2	3.6	9.8
MgO	0.7	4.7	8.5	0.3	5.2	0.5	0.5	0.8	2.0	0.4	5.3
CaO	0.4	0.1	6.1	2.0	8.6	0.5	1.0	0.1	0.1	0.2	4.8
Na_2O	5.1	0.3	4.7	4.1	4.9	8.8	7.3	6.8	3.9	6.2	5.3
K_2O	6.7	9.3	0.2	3.0	1.2	0.5	3.3	0.6	2.2	0.2	1.1
TiO_2	0.1	0.2	1.7	0.2	2.1	0.3	0.2	0.2	0.2	0.2	2.2
LOI	0.8	2.7	2.9	1.3	2.7	–	1.4	1.0	2.1	2.5	3.0
Total	101.9	97.8	98.0	97.0	99.1	98.2	102.6	98.4	98.9	99.4	97.5
FeO*/MgO	1.3	1.6	1.2	13.3	1.9	6.4	6.0	2.9	1.6	9.0	1.9
Mg#	43.8	38.8	45.5	7.0	34.7	13.5	14.2	25.8	38.5	10.0	35.1
Rock type	Rhyol	Hi-K Bas. Andes	Gabbro	Rhyol	Bas.	Rhyol	Rhyol	Rhyol	Rhyol	Rhyol	Hi-Al Bas.

	Gs1362	Gs1384	Gs1421	Gs1425	Gs1213B	Gs2145	GsX	B-1†	B-2†	B-2	B-3†
SiO_2	57.8	73.3	46.4	70.8	62.8	66.6	64.9	52.3	45.1	45.5	47.4
Al_2O_3	15.3	13.5	14.7	13.1	20.1	16.6	15.1	15.4	16.1	15.7	16.6
FeO*	8.6	0.2	11.9	2.7	1.3	2.0	5.4	9.3	10.6	11.1	10.3
MgO	1.3	0.2	7.2	1.4	1.5	0.2	1.7	4.8	6.3	6.4	6.4
CaO	3.0	0.4	8.0	0.9	0.2	0.1	1.4	7.9	9.7	9.6	10.9
Na_2O	6.0	6.3	3.7	4.1	9.7	3.9	6.5	5.5	3.8	3.5	3.1
K_2O	1.0	2.5	1.1	2.7	0.7	8.2	0.5	0.3	0.3	0.3	0.3
TiO_2	1.1	0.7	2.2	0.6	0.8	0.2	0.8	2.1	2.1	2.0	1.8
LOI	3.9	0.5	3.0	1.2	1.2	0.5	1.4	2.5	4.5	2.2	3.1
Total	97.9	97.6	98.2	97.5	98.3	98.3	97.7	101.0	99.0	96.3	101.0
FeO*/MgO	6.6	1.0	1.7	1.9	0.9	10.0	3.2	1.9	1.7	1.7	1.6
Mg#	13.1	50.0	37.7	34.1	53.6	9.1	23.9	34.0	37.3	36.6	38.3
Rock type	Andes	Rhyol	Bas.	Rhyol	Dac.	Dac.	Dac.	Bas.	Hi-Al Bas.	Bas.	Hi-Al Bas.

	B-3	B742	B752	B781A	B811	B834	B883	B922	Br964	Br1035	Br3651A
SiO_2	47.3	50.8	44.7	49.7	68.6	71.9	77.3	73.9	74.1	50.1	49.9
Al_2O_3	16.1	14.5	16.2	15.8	14.5	13.1	11.7	12.1	12.1	16.3	15.7
FeO*	10.2	9.6	14.6	11.6	2.8	3.7	1.1	2.9	3.0	9.5	10.6
MgO	6.1	4.4	4.9	4.8	0.2	0.4	0.2	0.1	0.3	3.2	5.4
CaO	10.8	8.5	9.5	6.6	0.2	0.1	0.1	0.1	0.2	5.7	6.9
Na_2O	2.7	5.4	6.1	5.1	4.5	4.7	2.8	5.4	4.7	5.6	5.1
K_2O	0.2	0.3	0.5	0.4	5.0	3.0	3.1	3.3	2.3	4.6	0.5
TiO_2	1.6	2.0	3.5	2.6	0.2	0.3	0.1	0.2	0.3	2.6	2.0
LOI	2.0	3.1	2.5	2.7	2.6	1.1	0.9	1.2	1.3	3.3	1.8
Total	97.0	98.6	102.6	99.3	98.6	98.4	97.3	99.1	98.2	100.9	98.5
FeO*/MgO	1.7	2.2	3.0	2.4	14.0	9.3	5.5	29.0	10.0	3.0	2.0
Mg#	37.4	31.4	25.1	29.3	6.7	9.8	15.4	3.4	9.1	25.2	33.8
Rock type	Hi-Al Bas.	Bas.	Hi-Al Bas.	Gabbro	Rhyol	Rhyol	Rhyol	Rhyol	Rhyol	Hi-Al Bas.	Bas.

† indicates U.S.G.S. duplicate analyses.

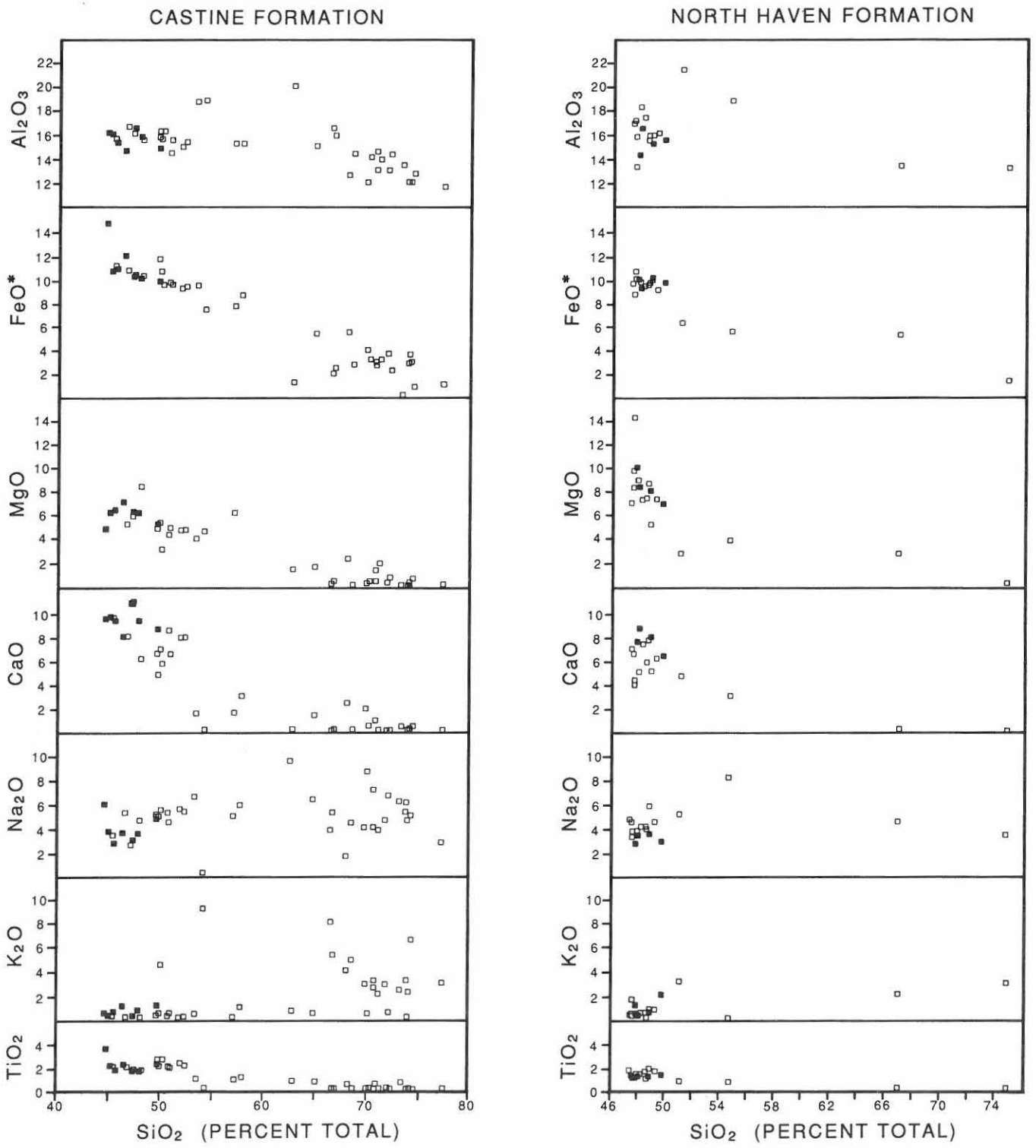


Figure 6. Harker-type variation diagrams for volcanic rocks in the North Haven and Castine Formations. Solid symbols represent least altered analyses.

the same amount in the dacite. CaO values for both of these felsic rocks are comparatively low, perhaps due to Ca depletion.

Mg-numbers (Mg# = [MgO/(MgO + FeO*)] x 100) for basalts of the North Haven Formation (Table 1) vary between 17.6 and 57.1. According to the criteria of Frey et al. (1979), these values indicate that most of these rocks are considerably evolved relative to primitive basalts (Mg-number = 68-75), probably due to fractionation of olivine and pyroxene.

Castine Formation

Chemical analyses of samples from the Castine Formation include basalts and basaltic andesites (~30%) and dacites and rhyolites (~70%). Although few andesites were found in this study, Smith el al. (1907) report andesite on the south shore of the island of North Haven, Shride (1976) described a thick section of andesites from the southern end of the belt in Massachusetts, and Bouley (1978) provides chemical data that indicates the presence of andesite near Castine, Maine. Numerous agglomerates were not chemically analyzed.

Cation abundances indicate that the majority of basalts and basaltic andesites are hypersthene normative and, therefore, alkaline (Irvine and Baragar, 1971). Most of the rhyolites and dacites are corundum normative compared to the rhyolite and dacite averages of Cox et al. (1979). However, alkali enrichment is also suspected in these rocks, and as a result their norms should be viewed with caution.

Major element oxide data are presented in Harker-type variation diagrams in Fig. 6. These diagrams indicate the bimodal character of the formation with a large number of basalts and basaltic andesites (SiO_2 <55%), a paucity of andesites (SiO_2 = 55-62%) and a majority of dacites and rhyolites (SiO_2 >62%). The basalts have a high concentration of Na_2O coupled with a degree of scatter for SiO_2, CaO, and alkalis. These relationships are taken to imply alteration of the original basaltic compositions by alkali enrichment. The concentration of TiO_2 is also high compared to basaltic averages (Chayes, 1975). Because of the relative immobility of TiO_2 during alteration (Pearce and Cann, 1973), the high TiO_2 must reflect magmatic abundances. In the dacites and rhyolites the concentration of Na_2O is also anomalously high and the alkalis have considerable scatter. Approximately 50% of the silicic rocks analyzed have sufficiently high concentrations of alkalis to be labeled alkali dacites or alkali rhyolites. Both potassic and sodic dacites and rhyolites are represented, but the sodic types predominate. These high alkali concentrations may be due to alkali enrichment, which may account for the low concentrations of CaO in these silicic rocks relative to the Cenozoic averages of Chayes (1975).

The Mg-numbers for the majority of the basalts are in the intermediate range (25.1 - 53.6) as compared to primitive basalts (Frey et al., 1979). These values indicate at least moderate fractionation of the parent magma and/or possible interaction with crustal materials. One sample from Little Spruce Head Island has an anomalously low Mg-number of 11.7.

Geochemical Comparison of the North Haven and Castine Formation Volcanic Suites

A major difference between the bimodal volcanic suites is that the North Haven Formation is dominated by mafic rocks with only minor occurrences of felsic rock, whereas the Castine Formation is dominated by felsic rocks with subordinate mafic varieties. Furthermore, the Castine Formation contains large amounts of agglomerate and tuff, and the North Haven Formation contains only limited amounts of these rocks.

Because felsic volcanic rocks are very sparsely represented in the North Haven Formation, only the mafic volcanic rocks are discussed in the following chemical comparisons. Referring to Table 1, SiO_2 shows a much broader range in the mafic volcanic rocks of the Castine Formation (41.7 - 54.2 wt%) compared to that in the North Haven Formation (47.5 - 54.7 wt%). Average TiO_2 concentrations are higher in the Castine Formation (1.89±0.60 wt%) than in the North Haven Formation (1.37±0.31 wt%). Al_2O_3 is less abundant in the Castine Formation than in the North Haven Formation, and a significant portion of the volcanic rocks of the North Haven Formation are high alumina basalts. MgO concentrations are slightly greater in the North Haven Formation, and FeO* levels for both formations are nearly the same. Accordingly, the FeO*/MgO values for the Castine Formation are somewhat greater than those for the North Haven Formation; on the other hand, the Mg-numbers are larger for the mafic volcanic rocks of the North Haven Formation. Although the concentration of CaO is greater in the mafic volcanic rocks of the Castine Formation than in those of the North Haven Formation, and the average concentrations of alkalis is uniformly high in both formations, these undoubtedly have been affected by alteration and do not reflect original magmatic values.

The marked differences in (1) the relative abundances of the major rock types in each suite, (2) the relative abundance of felsic tephra in each suite, and (3) the average TiO_2 concentration in the basalts in each suite suggest that the two volcanic suites represent different magmatic conditions.

Magmatic Lineage and Tectonic Affinity

In this section various discrimination diagrams are presented and discussed in an attempt to fix the magmatic lineage and tectonic setting of these rocks.

An alkalis-FeO*-MgO triangular plot (Fig. 7) is commonly used to establish the differentiation path of comagmatic igneous rocks. In the case of volcanic rocks, the parent magma is thought to be a basalt having a field as indicated near the FeO*-MgO sideline. Two differentiation paths have been established which describe the manner in which the chemistry of the parent magma is changed by the process of crystallization: the tholeiitic trend and the calc-alkalic trend. The tholeiitic trend, path A in Figure 7, moves away from the parent magma toward the FeO* apex, and at some large FeO*/MgO value turns toward the alkali apex.

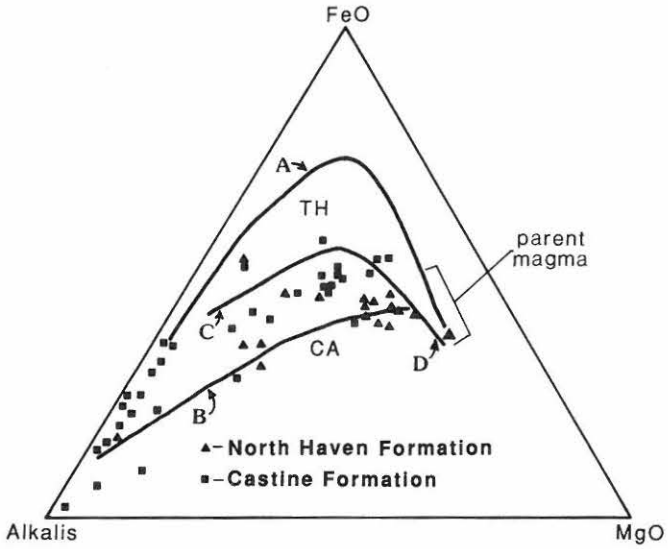


Figure 7. AFM diagram for volcanic rocks in the study area. Triangles and squares represent the North Haven and Castine Formations respectively. Line A represents a tholeiitic differentiation trend and line B represents a calc-alkaline trend. Discriminiation line C-D is after Irvine and Baragar (1971).

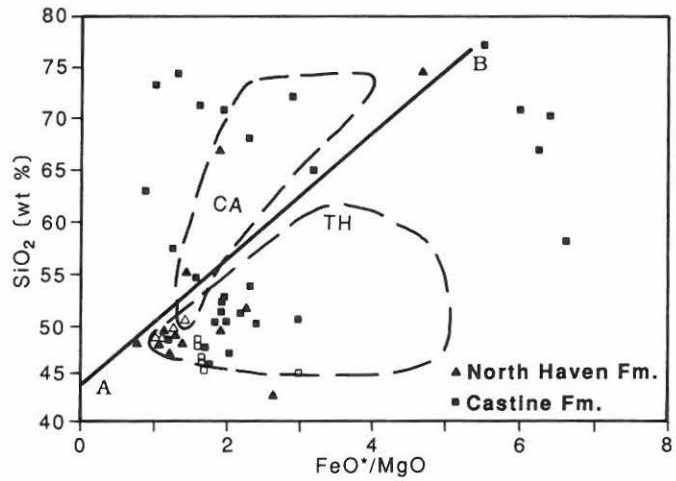


Figure 8. FeO*/MgO versus weight percent SiO_2 variation diagram for the volcanic rocks in the study area. Symbols as in Fig. 6. After Miyashiro (1974).

This trend is thought to be typical of the crystallization of rift-type volcanic rocks or some immature arcs. The calc-alkalic trend moves more directly from the parent compositions toward the alkali apex, as illustrated by path B in Figure 7. This trend is thought to be more common for the crystallization of island arc and Andean-type lavas. The heavy line C-D in Figure 7 separates the tholeiitic and the calc-alkaline field.

The compositions of samples from the North Haven and Castine Formations are plotted in Figure 7. Most of the samples lie in the calc-alkaline field, but because the rocks of both formations are thought to have been chemically altered, particularly with respect to the alkalis, the validity of this diagram may be questioned. However, the concentrations of the alkalis cannot be higher than 2 or 3%. The adjustments to the positions of the compositional points in Figure 7 must be made in a direction away from the alkali apex, approximately parallel to the line separating the tholeiitic and the calc-alkaline fields over the left half of the diagram. These arguments would not greatly change the distribution of compositional points in relation to line C-D. Those compositional points nearer the FeO*-MgO sideline would be displaced toward the area of the parent magma. Thus, we conclude that on this diagram the compositional points representing both the North Haven and the Castine Formations suggest a calc-alkaline trend.

Miyashiro (1974) has suggested the use of the SiO_2-FeO*/MgO plot to determine the differentiation paths of volcanic rocks. He argues that these parameters are less affected by alteration than would be those used in the alkalis-FeO*-MgO diagram. This conclusion is supported by the work of Vallance (1974) on the Deccan basalts, in which he showed that FeO*/MgO and SiO_2 were only slightly affected by spilitization. Miyashiro (1974) plotted several suites of volcanic rocks on this diagram and demonstrated that calc-alkaline suites plotted along a SiO_2 enrichment trend, whereas tholeiitic suites plotted along an FeO enrichment trend. These two trends are overlaid on Figure 8. The solid line (A-B) purportedly separates calc-alkaline from tholeiitic compositions, but it should be noted that this diagram does not satisfactorily differentiate the two suites below 50 wt% of SiO_2, and it is only at intermediate and higher concentrations of SiO_2 that the effects of differentiation are evident. The compositions of samples from the North Haven and the Castine Formations plot mainly within the field showing tholeiitic trends. A large number of basalts, however, plot in a part of the diagram where discrimination between the two suites is poor. In addition, the absence of any rocks of intermediate composition makes it difficult to establish a differentiation trend.

Both Figures 7 and 8 assume that the chemical characteristics of the rocks plotted on them are related by a differentiation process, and to the extent that this condition is not met, these diagrams may lack their stated significance. The proportions of basaltic to felsic rocks in the North Haven Formation does allow them to be related through differentiation, but even here the absence of intermediate rocks casts some doubt that they represent a true differentiation series. In the case of the Castine Formation, the preponderance of felsic rocks over basalts and the lack of intermediate rocks almost certainly precludes their being related by a differentiation process. This rock series must have a large component of crustal material that skews the compositions toward the felsic end of the compositional spectrum. For these reasons, we do not regard the trends on these diagrams to represent compelling arguments about the compositional lineage of these rocks.

The mineralogy of the pyroxenes is more definitive. Kuno (1968) regarded his hypersthenic and pigeonitic series as repre-

TABLE 2. TRACE AND RARE-EARTH ELEMENT DATA FROM SELECTED BASALTIC SAMPLES IN EAST PENOBSCOT BAY (U.S.G.S. UNPUBLISHED DATA). CONCENTRATIONS IN PARTS PER MILLION (ppm).

	E623	Gs-2	G-2	B-1	B-2	B-3
Ba	87	45	80	37	43	26
Nb	5	8	1	6	5	7
Rb	7	5	<5	<5	5	5
Sr	377	276	225	255	258	355
Zr	113	176	159	175	179	150
Y	29	32	30	32	34	31
Ti	6,589	13,777	9,584	12,579	12,579	10,782
Co	41.6	38.0	42.8	37.7	42.2	42.7
Cr	237	189	38.3	190	195	157
Cs	0.36	<0.4	0.47	<0.4	0.27	<0.4
Hf	2.47	3.53	3.22	3.57	3.80	2.93
Rb	11	<20	13	<20	<20	<20
Sb	1.91	0.6	<0.9	0.52	<0.9	0.70
Ta	0.22	0.54	0.69	0.55	0.50	0.33
Th	0.34	0.56	0.48	0.60	0.61	0.33
U	<0.4	<0.5	0.25	<0.5	<0.4	<0.4
Zn	77	81	81	79	96	90
Sc	34.4	39.7	34.8	39.6	35.4	42.7
La	4.36	8.4	9.51	8.9	8.9	6.6
Ce	13.3	21.8	23.1	20.9	24.7	18.3
Nd	15	16	17	22	15	18
Sm	3.05	3.97	3.60	3.87	4.7	3.82
Eu	1.24	1.41	1.29	1.42	1.75	1.55
Gd	5.4	6.4	4.6	<2	5.4	<6
Tb	0.77	0.88	0.68	0.83	1.01	0.87
Tm	0.38	0.5	0.4	0.5	0.5	0.30
Yb	3.12	3.04	2.18	2.88	3.27	3.10
Lu	0.447	0.429	0.319	0.447	0.495	0.446

Ba – Y analyses by X-Ray Spectroscopy
Ti analysis by Atomic Absorption Spectrophotometry
Co – Lu analyses by Instrumental Neutron Activation

senting volcanic suites of the calc-alkaline and tholeiitic series, respectively. According to the conclusions of Kuno (1968), the presence of monoclinic pyroxene and the absence of orthorhombic pyroxene in the basalts from the North Haven and Castine Formations puts them in the pigeonitic series and classifies them as tholeiitic. As noted by Miyashiro (1974), there are cases where this mineralogic definition disagrees with the chemical definitions.

The concentration of TiO_2, because of the immobility of this oxide during alteration, has been used as a discriminant of tectonic setting of tholeiites (Pearce and Cann, 1973). In addition, Pearce (1974) has reported mean concentrations of TiO_2 for various suites of mafic rocks interpreted to have certain tectonic settings as follows: ocean-floor tholeiites = 1.42 wt%, calc-alkaline mafic volcanic rocks = 1.07 wt%, and continental rift basalts = 2.47 wt%. Feiss (1982) has suggested that TiO_2 concentrations of 1.0 - 1.3 wt%, or less, in mafic rocks are typical of subduction zones. Taken at face value, the mean TiO_2 concentration of the samples taken from the North Haven basalts (1.37 wt%) compares most favorably to that of Pearce's ocean-floor tholeiite, and the mean of samples from the Castine Formation (1.89 wt%) compares most favorably to that of Pearce's mean value for continental rift basalts.

Trace and rare-earth element abundances and abundance ratios have been used by numerous researchers to define the tectonic setting of volcanic suites (e.g. Frey et al., 1968; Schilling, 1969; Pearce and Cann, 1973; Wood et al., 1979b; Frey, 1980; Brooks and Coles, 1980). To the extent that magmatic processes are controlled by tectonic conditions, the trace and rare-earth element abundances of volcanic suites may be used to infer their tectonic setting. In addition, the immobility of trace and rare-earth elements make them especially useful as guides to magmatic processes that operated in ancient volcanic suites.

Trace and rare-earth elements for six basaltic samples were analyzed by the U.S. Geological Survey laboratory in Reston, Virginia. The analytical data are presented in Table 2. One analysis is from the North Haven Formation and five are from the Castine Formation. The data presented here are too few to establish any convincing interpretations regarding magmatic or tectonic conditions. However, as more trace and rare-earth element data become available, the accumulated data base may better establish magmatic trends and tectonic settings. In order

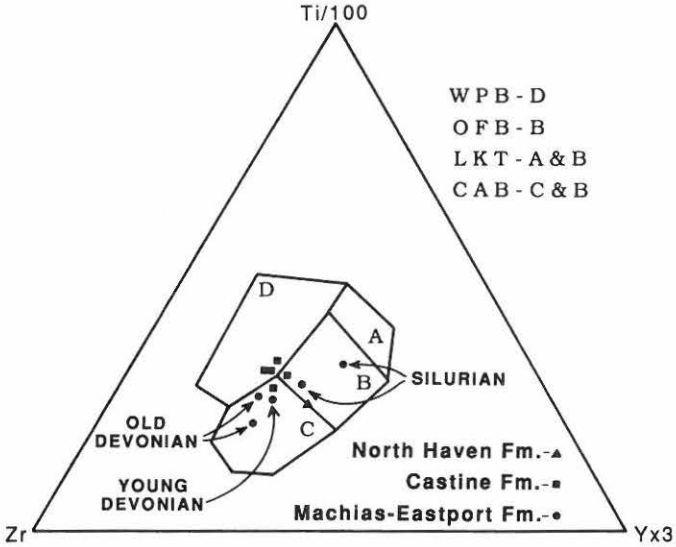


Figure 9. Discrimation diagram using Ti, Zr, and Y. Within-plate basalts (WPB), low-potassium tholeiites (LKT), ocean-floor basalts (OFB) and calc-alkaline basalts (CAB) plot in fields as indicated above. After Pearce and Cann, (1973). Triangle, squares, and circles represent North Haven and Castine Formations, and rocks from the Machias-Eastport area (Gates and Moench, 1981), respectively.

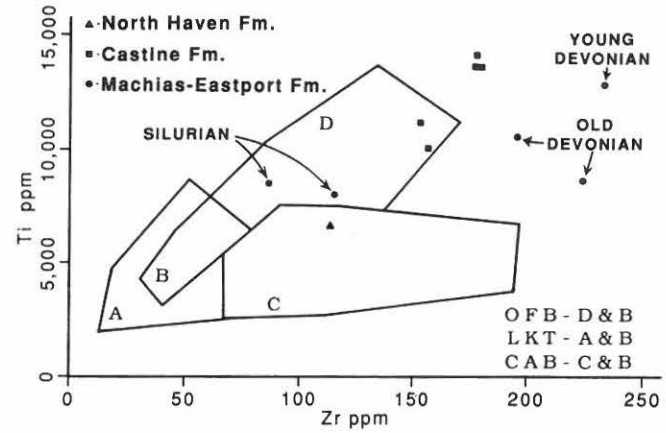


Figure 10. Discrimination diagram using Ti and Zr. Symbols and labels are as in Figure 9. After Pearce and Cann (1973).

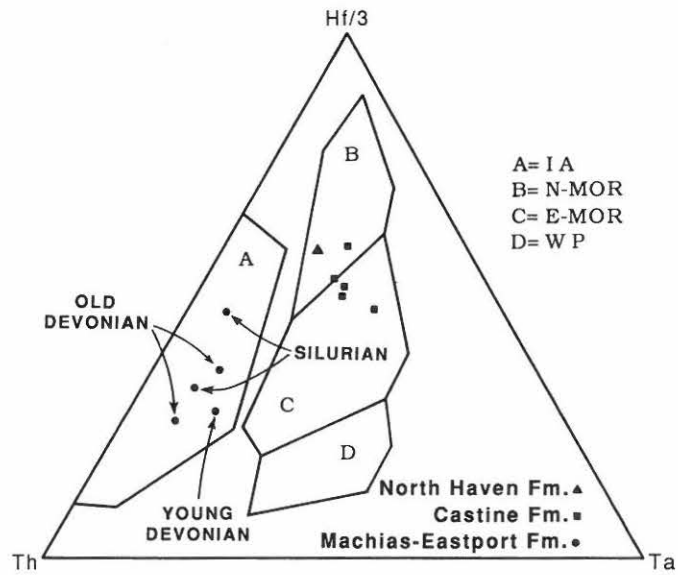


Figure 11. Discrimination diagram for basalts using Hf, Th, and Ta. Island arc (IA), normal MOR (N-MOR), enriched MOR (E-MOR), and within-plate (WP). Symbols as in Fig. 9. After Wood et al. (1979a).

to augment our data base, the diagrams that follow include the trace and rare-earth element analyses presented by Gates and Moench (1981).

Pearce and Cann (1973) outline a procedure for using discrimination diagrams involving the elements Ti, Zr, and Y to discriminate amongst various tectonic settings. First, they utilize the Ti/100-Zr-3Y diagram to determine whether within-plate basalts are represented. Figure 9 shows the positions of the analyses presented here and those reported by Gates and Moench (1981). The analysis from the North Haven Formation plots near the boundary between fields B and C. Most of the analyses from the Castine Formation lie in field D, but close to its junction with fields B and C, and one analysis each lies in fields B and C. The five samples reported by Gates and Moench (1981) fall into fields B and C. Field B was designed to contain ocean-floor basalt, low-potassium tholeiite, and calc-alkaline basalt. Field C was drawn on the basis of calc-alkaline basalt, and field D is purported to contain within-plate basalt.

Pearce and Cann (1973) then use the Ti-Zr diagram only if within-plate basalts are not represented. Although within-plate basalts are represented in Figure 9, their positions on the diagram are very near the fields for ocean-floor and calc-alkaline basalts. For this reason we present data from the study area along with data from Gates and Moench (1981) for the Machias-Eastport area on a Ti-Zr diagram (Fig. 10). The one analysis from the North Haven Formation plots within the field of calc-alkaline basalt, two analyses from the Castine Formation plot within the field for ocean-floor basalt, and the remaining analyses from the Castine Formation plot outside any of the three fields represented. Three of the analyses presented by Gates and Moench (1981) also plot outside of the fields delineated by Pearce and Cann (1973). This diagram does not successfully fix the tectonic setting of the basalts from the coastal volcanic belt.

Wood et al. (1979a) utilized a diagram based on the abundances of Hf, Th, and Ta to differentiate between rocks generated in different tectonic settings (Fig. 11). They distinguished normal oceanic ridge basalt, enriched oceanic ridge basalt, within-plate basalt, and basalt generated at destructive plate margins. The one analysis from the North Haven Formation plots in the field for normal oceanic ridge basalt, whereas the analyses from

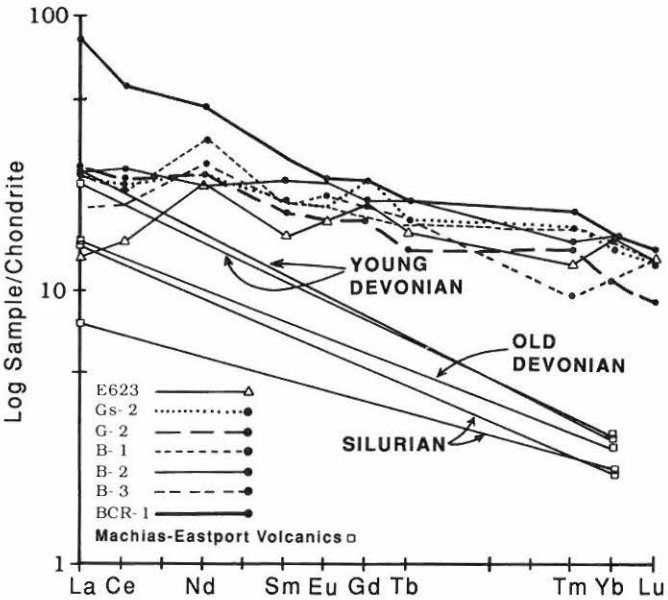


Figure 12. Chondrite-normalized rare-earth element diagram. Values normalized to data of Haskin et al. (1968). Data for La and Yb from Machias-Eastport area (Gates and Moench, 1981) are connected with straight lines. Symbols as in Fig. 9.

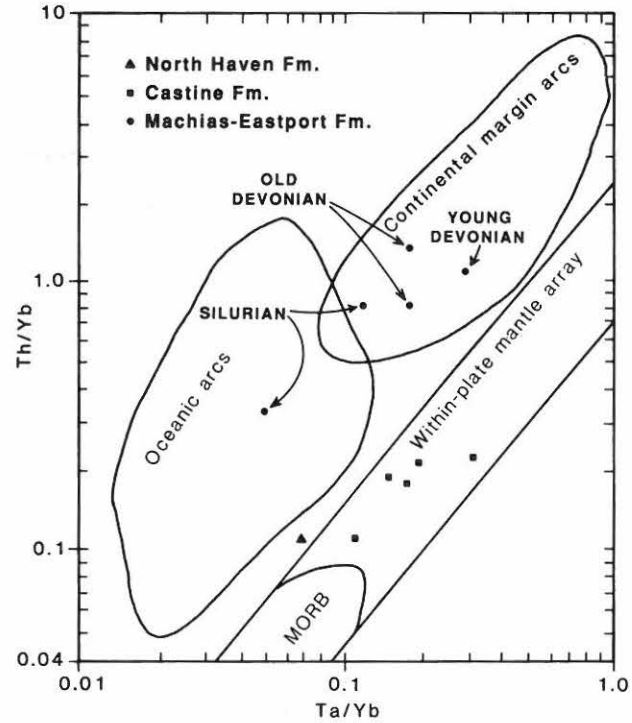


Figure 13. Th/Yb-Ta/Yb diagram showing distribution basalts from the North Haven Formation, Castine Formation, and the Machias-Eastport area (Gates and Moench, 1981). Symbols as in Fig. 9. Adapted from Condie (1986).

the Castine Formation plot in the fields of normal oceanic ridge and enriched oceanic ridge basalts. The data of Gates and Moench (1981) plot in the field for island arcs.

Rare-earth element data for the six basalts from Penobscot Bay are plotted in Figure 12 on a chondrite-normalized diagram. The single analysis from the North Haven Formation displays a very irregular, slightly depleted light rare-earth element pattern with La/Yb equal to 0.84. Individual rare-earth element concentrations range between 12.6 and 25 times chondrite values. In general, the pattern of slightly depleted light rare-earth elements and the observed chondrite-normalized values are similar to those for ocean-floor basalts reported by Schilling (1969) and Frey (1980).

The five rare-earth element analyses from the Castine Formation display mild light rare-earth element enrichment patterns relative to chondrites and have chondrite-normalized La/Yb values ranging between 1.29 and 2.64 (Fig. 12). Individual concentrations of rare-earth elements range between 9.9 and 36.5 times chondrite values. These mildly enriched rare-earth element patterns are similar to those for both continental tholeiitic basalt (Herrmann et al., 1974; Frey, 1980; and Dupuy and Dostal, 1984) and calc-alkaline basalt and andesite (Peccerillo and Taylor, 1976; and Garcia, 1978). The lack of any Eu anomaly in these basalts indicates that fractional crystallization of plagioclase was not an important factor in their differentiation history.

Gates and Moench (1981) supplied data on the abundance of La and Yb for five basalts from the Machias-Eastport area. These chondrite-normalized values are shown as individual points on Figure 12. The Machias-Eastport rocks show greater La enrichment relative to Yb compared to the basalts of this study. La concentrations for the younger Devonian and one of the older Devonian basalts are within the range of La concentrations in the Castine Formation. The basalts from this study have higher Yb concentrations than any of the basalts studied in the Machias-Eastport area. Gates and Moench (1981) observed that the high values for La and Yb and the relatively high La/Yb values (3.5 to 8) in the basalts from Machias-Eastport are similar to those for calc-alkaline basalts, but they are unlike those characteristic for ocean-floor, island-arc, and back-arc tholeiites.

In the Th/Yb-Ta/Yb diagram (Condie, 1986) presented as Figure 13, the analysis from the North Haven Formation plots outside all fields. It should be noted, however, that its position lies fairly close to the fields for oceanic arc and within-plate basalts. The analyses from the Castine Formation all plot in the field for within-plate basalts. Four basalt analyses from the Machias-Eastport area (Gates and Moench, 1981) plot in the field for continental arcs and one analysis plots in the field for oceanic arcs.

On the MORB-normalized incompatible element diagram shown in Figure 14, a pronounced negative Rb anomaly, negative Nb anomalies, and overall light incompatible element depletion are evident for the basalt analyses from the North Haven Formation. The five basalt analyses from the Castine Formation show approximately similar trends for the lighter elements, but,

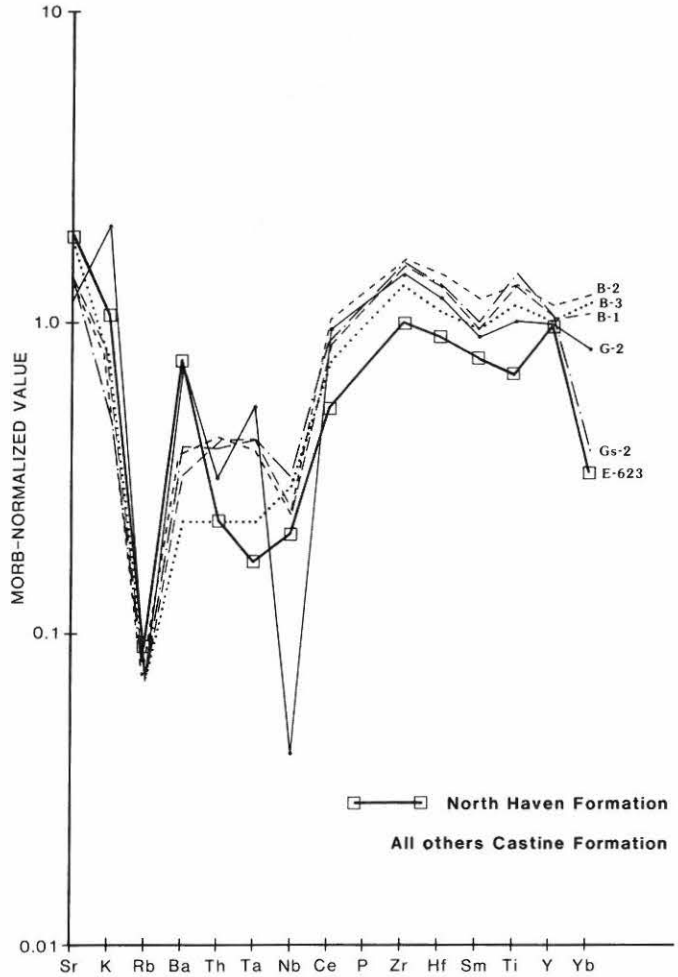


Figure 14. MORB-normalized incompatible element distributions for basalts from the study area. Symbols as in Fig. 9. Data normalized using averaged data for MORB basalts in Wood et al. (1979b).

generally, Nb is more depleted than either Th or Ta. The heavier elements (Ce to Yb) are in general equal to or slightly enriched compared to the MORB incompatible element abundances. In all of the six analyses presented, enrichment of Sr and K relative to the other light incompatible element may be related to secondary chemical alteration discussed earlier.

A strong negative Ta-Nb anomaly is thought to be characteristic of calc-alkaline basalts at immature arcs (Atherton et al., 1985; Condie, 1986). This feature is evident in the incompatible element distribution for the analysis from the North Haven Formation. While Ta-Nb depletion is evident in the basalts from the Castine Formation, no strong negative anomaly is apparent. This contrasts the strong negative Ta-Nb anomaly observed by Hon and Thirlwall (1985) in the Silurian and Devonian Newbury volcanic rocks of Massachusetts. They cite this feature as one piece of evidence indicating that the volcanism in this area occurred in an island arc environment. It should be noted that we have not compared their actual data to those data presented in Figure 14.

GEOLOGICAL CONSTRAINTS

North Haven Formation

The disproportionate abundance of mafic over felsic volcanic rocks in the North Haven Formation is consistent with oceanic basalt or basalts of a young island arc. A large number of the basalts are pillowed; therefore, they must have been deposited subaqueously. The proportion of felsic material is consistent with an origin by partial melting of basalt (Gibson, 1969) or by differentiation of a parent basaltic magma (Yoder, 1973). Yoder has calculated that a basaltic magma could yield by differentiation up to 20% of its original volume as felsic rocks. Production of such relatively minor amounts of felsic material in a continental setting appears unlikely since the anatexis of crustal material would result in a greater abundance of silicic rocks than are present (Battey, 1966; McBirney, 1969).

Castine Formation

The voluminous felsic volcanic rocks in the Castine Formation are inconsistent with an origin by differentiation from basalts, and their abundance must mean that they have at least in part been derived by partial fusion of continental crustal material. This contention is supported by the initial Sr^{87}/Sr^{86} value of 0.7079±0.0015 reported by Brookins et al. (1973) for the Castine Formation. Further support for this conclusion comes from the observation of dikes of Castine volcanic rocks cutting the Islesboro sequence which is interpreted to form part of the cover sequence of the Avalon basement. Many of the basalts are pillowed, suggesting subaqueous deposition, but felsic flows are characterized by lapilli tops, perlitic fractures, and columnar joints, features that suggest a terrestrial environment. A large portion of the volcanic rocks in the Castine Formation are agglomerates and tuffs which are characteristic of arc-type volcanism (Gill, 1983). Although no andesites are present in the area of this study, they have been reported within the Silurian-Lower Devonian volcanic rocks of the coastal belt from Castine and the island of North Haven in Maine and from the Newbury volcanic rocks in Massachusetts. Our present knowledge of the regional proportions of basalt, andesite, dacite, and rhyolite in the coastal belt is not good.

SUMMARY AND CONCLUSIONS

The North Haven Formation lies unconformably beneath a sedimentary unit that in turn is overlain by the Silurian-Lower Devonian volcanic rocks. The age of the North Haven Formation is not well fixed; possible ages range from Precambrian to Ordovician. Both units are metamorphosed to pumpellyite-actinolite subfacies of the low greenschist facies. The North Haven

Formation is polydeformed, whereas the Castine Formation is not.

The North Haven Formation consists mainly of basalts. The pillowed nature of some of these basalts suggests a subaqueous origin. Felsites make up less than 10% of the formation, and the proportion of basalt to felsite suggests that the felsites may be differentiation products of basalts in an oceanic environment (i.e., away from continental crust). Because rocks of intermediate compositions are absent and the felsites are sparse, they cannot be used to establish a differentiation trend. However, SiO_2-FeO*/MgO diagrams for the basalts and the presence of only a monoclinic pyroxene suggest a tholeiitic composition.

The TiO_2 contents of the basalts in the North Haven Formation are in the range of typical ocean-floor basalts. However, most of the other major element and trace element data that we have do not distinguish unequivocally among the possible tectono-magmatic settings for the formation. Hf/3-Th-Ta data suggest ocean-floor basalts and the rare-earth element data have similarities to those for ocean-floor basalts. A negative Ta-Nb anomaly in these rocks suggests that they may be part of an immature oceanic arc.

The Castine Formation contains a preponderance of felsic volcanic rocks, much felsic tephra and water-laid volcanic conglomerates, and a smaller amount of basalt, some of which is pillowed. Features related to the cooling of these volcanic rocks suggest mixed terrestrial and subaqueous deposition. The abundance of felsites mitigates against their origin by simple differentiation of basalt but suggests that much of the felsite must have originated by partial melting of continental crustal material. This origin for the felsites indicates that the Castine Formation must have been deposited on continental crust and rules out oceanic origins. Two origins are possible: continental rift and Andean-type arc. The chemistry is far from definitive in discriminating between these two cases. Because it is unlikely that the felsites are differentiation products of basalt, they cannot be used in alkalis-FeO*-MgO or SiO_2-FeO*/MgO diagrams to establish lineage trends. In those diagrams the basalts alone do not establish a trend, and they must be regarded as essentially tholeiitic. The TiO_2 concentrations of the basalts are intermediate between those typical of ocean floor basalts and basalts found in continental rifts. The rare-earth element data are not definitive in determining a tectonic setting, showing similarities to both continental rift tholeiites and calc-alkaline basalts and andesites. The trace element discrimination diagrams that are most successful in classifying the basalts suggest that they were generated in a within-plate environment. Due to the large amount of felsic rocks present in the area and an inferred continental basement for the region, we interpret this to be an ensialic rift. The large amount of agglomerate in the Castine Formation and the existence of andesites in the outlying areas may contradict this conclusion and suggest that the Castine Formation and its associated volcanic rocks in the coastal volcanic belt in general may have formed in an Andean-type plate margin.

To further advance the understanding of the tectono-magmatic character of the coastal volcanic belt in Massachusetts, Maine, and New Brunswick, additional research should focus on expanding the geochemical data base for the volcanic rocks within this belt. The major element oxides, trace elements, and rare-earth elements for the basalts and andesites along the entire volcanic belt will be key to identifying their tectonic setting. Chemical analyses of the Silurian-Devonian andesites and associated basalts that occur on North Haven Island in Penobscot Bay, Maine, and near Newbury, Massachusetts, would appear to be the most logical next phase of research.

ACKNOWLEDGMENTS

A significant portion of the research presented in this article was part of Pinette's M.S. thesis. During this period, the University of Maine Graduate Student Board and the Maine Department of Conservation (Maine Geological Survey) provided some valuable financial assistance to support the field program. The University of Maine Departments of Oceanography and Civil Engineering each provided a powerboat for transportation during the two field seasons. The University of Maine Department of Geological Sciences provided laboratory support for major element analyses (through Dr. Stephen A. Norton) and X-ray diffraction studies (through Dr. Joseph V. Chernosky). The U.S. Geological Survey provided trace and rare-earth element analyses. Mr. B. A. Bouley kindly provided microprobe data for basalts from Butter Island. Special thanks are due to the inhabitants of Eagle Island, Mr. John Porter, and Mr. Thomas D. Cabot for providing access to Eagle, Great Spruce Head, and Butter Islands, respectively. Mr. and Mrs. Robert L. Quinn of Eagle Island deserve special appreciation for their overall support and friendship throughout the field program.

This article benefited from critical review and comments by James W. Skehan and Olcott Gates. Discussions with David B. Stewart provided important information regarding the stratigraphy of the study region and the metamorphic mineral assemblages present in the two volcanic suites.

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Maine Geological Survey
Studies in Maine Geology: Volume 3
1989

Stream-Sediment Geochemistry of the Attean Quartz Monzonite and Nearby Rocks, Somerset and Franklin Counties, Maine

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ABSTRACT

A stream-sediment geochemical survey of terrain underlain by the Upper Ordovician Attean quartz monzonite in the Jackman area of west-central Maine was undertaken in order to identify potentially favorable areas for mineral deposits in addition to the known occurrences of base-metal veins and copper-molybdenum porphyries. The known occurrences of mineralization in the study area are at Catheart Mountain (Cu-Mo porphyry), west Sally Mountain (Cu-Mo porphyry), east Sally Mountain (disseminated chalcopyrite in metasedimentary rocks that unconformably overlie granite; minor galena in quartz veins), Pyrite Creek (quartz veins with minor galena, sphalerite, and chalcopyrite), and Bean Brook Mountain (minor sphalerite, galena, and chalcopyrite in felsite dikes and quartz veins in arkosic metasedimentary rocks that unconformably overlie granite). The studies show that the anomalies associated with the known mineralized rock extend beyond the immediate vicinity of the showings, and the intensity of the anomalies is greater than can be explained by the amount of known mineralized rock. Areas near Catheart Mountain and Sally Mountain appear to be favorable for the existence of additional copper-molybdenum porphyry deposits. The geochemistry of stream sediments and the geology at the unconformity between Ordovician granitic rocks and overlying metasedimentary rocks warrant further studies into the possible existence of sandstone Pb deposits. An areal relationship between faults and stream-sediment geochemical anomalies suggests that undiscovered structurally controlled vein deposits may exist.

INTRODUCTION

Several occurrences of sulfides associated with Ordovician granitic rocks were discovered in the early 1960's near Jackman, Maine, as a result of stream-sediment geochemical surveys. The Attean area was included in an earlier stream-sediment geochemical survey of west-central Maine (Post and Hite, 1964; Chaffee et al., 1972). The present study was undertaken in order to areally define the extent of anomalies associated with those mineralized areas and to locate other possible anomalies. Samples were collected from 1978 to 1982. In 1979 the study was integrated with investigations of the Sherbrooke and Lewiston 1° by 2° quadrangles (Fig. 1) as part of the Conterminous United States Mineral Assessment Program (CUSMAP) of the U.S. Geological Survey (Curtin, 1985).

DESCRIPTION OF STUDY AREA

The Attean study area comprises about 430 square miles in west-central Somerset and northern Franklin Counties (Fig. 1). Access to most of the study area is by private logging roads that branch from U.S. Route 201 north and south of Jackman; the southwestern part of the area is reached by private logging roads from Maine Route 27 north of Stratton. The Attean area, in general, is mountainous, with peaks as high as 3,718 ft (Coburn Mountain) and an average relief of 500-1,000 ft.

Excellent descriptions of the geomorphology and surficial geology of the Attean quadrangle are given by Albee and Boudette (1972), and much of that information applies to the entire Attean study area. Pleistocene glaciation resulted in a blanket of till over the study area, ranging in thickness from a

few inches to as much as 90 ft. Albee and Boudette (1972) estimate that 80 percent of the Attean quadrangle is covered by till. Glacial outwash deposits, alluvial deposits, and bog deposits cover much of the remaining 20 percent of the land surface. Additional information on the surficial geology of the area can be found in maps by Caldwell and Hanson (1976a,b,c,d) and Caldwell et al. (1976).

Bedrock Geology

The bedrock geologic map (Fig. 2) was modified from a preliminary geologic map of the Sherbrooke and Lewiston 1° by 2° quadrangles (Moench, 1984). In addition, geologic data were obtained for the Skinner area from Goldsmith (1985), for the Pierce Pond quadrangle from Boone (1985), for portions of the Spencer Lake quadrangle from Burroughs (1979), and for the Catheart Mountain pluton from Schmidt (1974).

The oldest rock unit in the study area is the Middle(?) Proterozoic Chain Lakes massif (Yc on Fig. 2), which consists of massive to poorly stratified diamictite. Most of the Attean area is underlain by the Upper Ordovician Attean quartz monzonite (Oa on Fig. 2) of the Highlandcroft Plutonic Suite (Lyons et al., 1986). The largest mass of Attean quartz monzonite (300 mi^2) is the Attean pluton, which is actually hornblende-biotite granite (Moench, 1984; nomenclature of Streckeisen, 1976). South of Little Big Wood Pond, rocks of the Attean pluton are described as mottled pink and green, medium- to coarse-grained porphyritic rock characterized by large phenocrysts of potassium feldspar (Albee and Boudette, 1972). North and west of Little Big Wood Pond, rocks of the Attean pluton are highly altered and locally have a cataclastic schistosity; fresh surfaces are green and large potassium feldspar phenocrysts are only locally prominent (Albee and Boudette, 1972). A smaller mass of Attean quartz monzonite, the Skinner pluton (25 mi^2), is hornblende-biotite granite to granodiorite (Moench, 1984; Goldsmith, 1985). Rocks of the Skinner pluton are similar to the larger part of the Attean pluton which crops out south of Little Big Wood Pond, but are less porphyritic and locally contain less potassium feldspar (R. H. Moench, unpub. data, 1986). The Skinner pluton is probably a satellite body of the Attean pluton and is barely unroofed, as interpreted from the map outline (Goldsmith, 1985). The Skinner pluton and the west side of the Attean pluton intrude the Middle(?) Proterozoic Chain Lakes massif.

Aplitic granite and quartz porphyry dikes are apparently common in the Attean quartz monzonite (Albee and Boudette, 1972; Schmidt, 1978; Goldsmith, 1985). Similar rock types occur as small plutons within the Attean quartz monzonite at Catheart Mountain (Schmidt, 1974) and Sally Mountain (Albee and Boudette, 1972) (Oc on Fig. 2). The granite of the Catheart Mountain pluton has a uniformly medium-fine texture that contrasts with the coarser, porphyritic granite of the Attean pluton. Granites of the Attean pluton and Catheart Mountain pluton are chemically and modally similar (Schmidt, 1974, Table 1; Nowlan, 1976b, Table 3). Isotopic data (Lyons et al., 1986) and the field relationships between rocks of the Attean pluton and the bodies of aplitic granite, medium-fine granite, and quartz porphyry at Catheart Mountain and Sally Mountain suggest that the smaller bodies have intruded the Attean pluton (Albee and Boudette, 1972; R. H. Moench, unpub. data, 1986).

The northeast, east, and southeast margins of the Attean pluton are unconformably overlain by Silurian and Devonian metasiltstone, metasandstone, and metaconglomerate (Pzls on Fig. 2), with minor Ordovician volcanic rocks (Pzlv on Fig. 2), Silurian calcareous rocks (DSc on Fig. 2), and Devonian limestone (DSc on Fig. 2). The Attean pluton is in fault contact with Silurian calcareous metasedimentary rocks (DSc on Fig. 2) at the southern end of the pluton and Devonian siliciclastic metasedimentary rocks (Pzls on Fig. 2) at the northern end. A Lower or Middle Devonian pluton, the Hog Island granodiorite (Dh on Fig. 2), intrudes the north margin of the Attean pluton south of Jackman.

Rocks of the area were metamorphosed to chlorite grade at the time of the Acadian orogeny (Albee and Boudette, 1972). The study area is crossed by the northeast-oriented Boundary Mountains anticlinorium (Albee and Boudette, 1972).

Mineralized Areas

Within the study area, mineralized rock occurs at Catheart Mountain (Cu-Mo), Sally Mountain (Cu-Pb and Cu-Mo), Pyrite Creek (Pb-Cu-Zn), and Bean Brook Mountain (Zn-Pb-Cu) (Fig. 2). Exploration based on studies of stream-sediment geochemistry led directly to the discovery of these five prospects (Canney and Post, 1964; Post and Hite, 1964; Delaney, 1968; Young, 1968).

The Catheart Mountain deposit at the western end of Catheart Mountain is described as a Cu-Mo porphyry deposit associated with breccia zones in granite of the Catheart Mountain pluton (Young, 1968; Schmidt, 1974, 1978; Atkinson, 1977). Alteration zones similar to those in other porphyry deposits are present (Schmidt, 1974). Chalcopyrite and molybdenite are the principal ore minerals. The deposit contains an estimated 20-25 million metric tons of mineralized rock with an average grade of about 0.25 percent Cu and 0.04 percent Mo (F. C. Canney, pers. commun., 1987). The land surface is generally well drained.

Mineralized rock is present at two sites on Sally Mountain, a 5-mile-long, east-west-trending ridge (Fig. 2). Industry-sponsored exploration led to the discovery of sparse chalcopyrite and pyrite at east Sally Mountain at the unconformity between the Attean quartz monzonite and the overlying Silurian metaconglomerate, but few details are available. The mineralization probably occurred after the formation of the unconformity (F. C. Canney, pers. commun., 1987). In addition, sparse galena in a quartz vein was discovered in the metasedimentary rocks within about 100 ft of the unconformity (F. C. Canney, pers. commun., 1987). At west Sally Mountain, chalcopyrite and molybdenite occur disseminated and in quartz veins associated with shear zones in quartz porphyry of the Sally

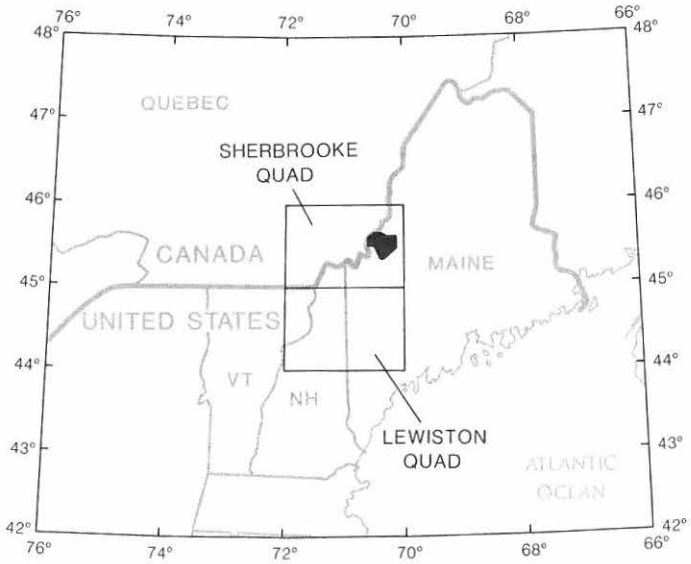

Figure 1. Location of the study area.

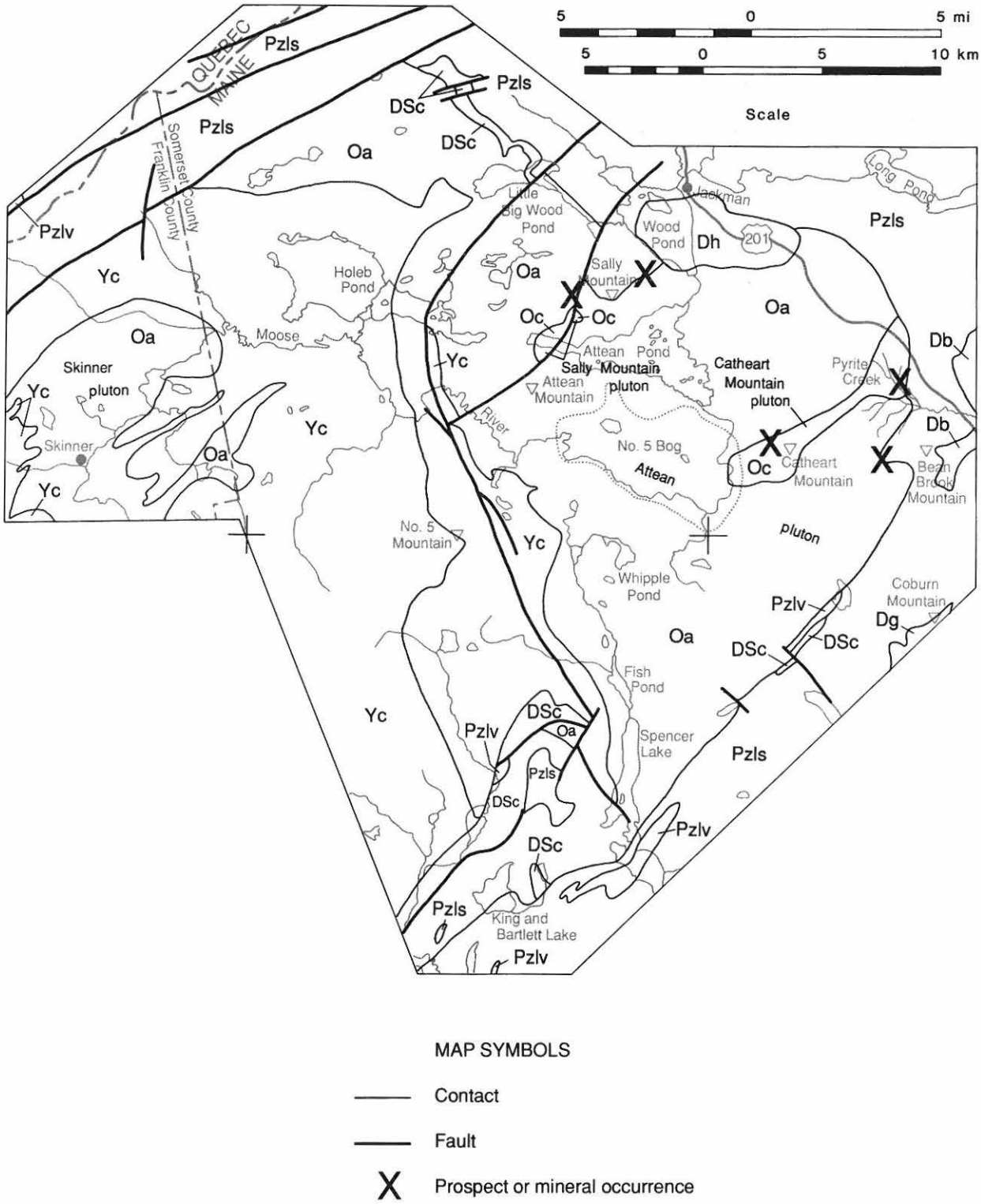

Figure 2. Generalized bedrock geologic map of terrain underlain by the Attean quartz monzonite and nearby geologic units, Somerset and Franklin Counties, Maine. Geology modified from Moench (1984), Boone (1985), Goldsmith (1985), Burroughs (1979), and Schmidt (1974).

INTRUSIVE ROCKS

Dh Hog Island granodiorite (Middle or Lower Devonian)--Hornblende- biotite granodiorite

Db Diabase sill at Bean Brook Mountain (Devonian)

Dg Intrusive garnet rhyolite at Coburn Mountain (Devonian)

Oc Aplitic granite and quartz porphyry at Catheart Mountain and Sally Mountain (Upper Ordovician)

Oa Attean quartz monzonite (Upper Ordovician)--Hornblende-biotite granite (Attean pluton) and hornblende-biotite granite to granodiorite (Skinner pluton)

STRATIFIED METAMORPHIC ROCKS

DSc Calcareous shallow marine-shoreline metasedimentary rocks (Devonian and Silurian)--Includes McKenney Ponds Limestone Member of the Tarratine Formation (Lower Devonian, limestone with subordinate arkose, conglomerate, slate and sandstone); Beck Pond Limestone (Lower Devonian, coarse quartzose limestone, stromatoporoid biostromes, and granite-boulder conglomerate); Hardwood Mountain Formation (Upper Silurian, mixed calcareous mudstone, siltstone, and slate with minor limestone conglomerate); calcareous rocks at Foxs Camp (Silurian, massive to slightly fissile, fine-grained limestone and calcareous slate and shale).

Pzls Siliciclastic metasedimentary rocks (Devonian, Silurian, and Ordovician)--Includes Seboomook Formation (Lower Devonian, deep marine, cyclically bedded slate and graded metasiltstone and wacke); Tarratine Formation (Lower Devonian, marine, interbedded sandstone, slate, and siltstone); Hobbstown Formation (Lower Devonian to Upper Silurian, shallow marine-shoreline, coarse-grained arkose and conglomerate); conglomeratic sandstone at Foxs Camp (Devonian and Silurian, shallow marine-shoreline, medium- to coarse-grained calcareous quartz-feldspar sandstone and granule-conglomerate sandstone); conglomerate at Foxs Camp (Silurian, shallow marine-shoreline, boulder and cobble conglomerate, conglomeratic sandstone, and feldspathic quartzite; metasedimentary member of the Frontenac Formation (Silurian(?) and Ordovician(?), marine, interbedded fine-grained feldspathic sandstone and slate or pelitic phyllite; possibly derived from felsic volcanic rocks).

Pzlv Metavolcanic rocks (Lower Paleozoic)--Includes Camera Hill Greenstone Member of Seboomook Formation (Lower Devonian, feldspar-porphyritic flows of intermediate(?) composition, locally vesicular); basaltic greenstone within the Frontenac Formation (Silurian(?) and Ordovician(?), pillow basalt and related basaltic tuff, agglomerate, breccia, and resedimented mafic volcanics; at northwest margin of map); Lobster Mountain Volcanics (pre-Silurian, porphyritic felsite and bedded felsic tuff; three miles west of Coburn Mountain and two miles north of King and Bartlett Lake).

Yc Diamictite (Middle(?) Proterozoic)--Massive to poorly stratified diamictite of Chain Lakes massif. Matrix is massive, aluminous, quartz-feldspar-mica granofels; clasts include vein quartz, mafic and felsic plutonic rocks, volcanic rocks, and previously deformed and metamorphosed sedimentary rocks. The rocks are dioritic along the intrusive contact from No. 5 Mountain to the west end of Holeb Pond.

CORRELATION OF MAP UNITS

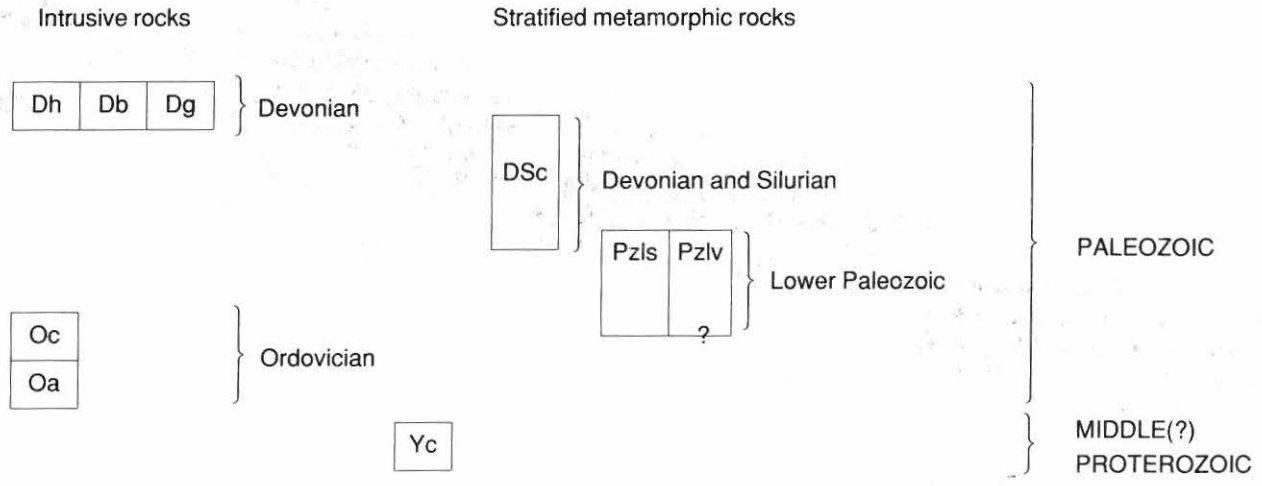

Figure 2 (continued). Explanation of map units.

Mountain pluton. This occurrence was interpreted as porphyry-type mineralization (Young, 1968); sparsely mineralized bedrock was observed from near the ridge top to near the southern base of the mountain, a distance of 0.5 mi (F. C. Canney, pers. commun., 1987). Only minor prospecting has taken place at either site. The slopes of Sally Mountain are generally well drained. Boggy areas 2-5 acres in size are situated on either side of the ridge top near the western occurrence of mineralized rock.

The mineralized rock at Pyrite Creek is in granite of the Catheart Mountain pluton within several hundred feet of the unconformity with overlying Devonian metasedimentary rocks. Mineralization deposited very minor galena, pyrite, chalcopyrite, and sphalerite in quartz veins and in silicified granite near the quartz veins (Canney and Post, 1964; Nowlan 1976b). Three holes were drilled to a depth of 100 ft or less, but only sparsely mineralized rock was encountered. Mineralized rock at Pyrite Creek is present 500 ft upslope and the same distance downstream from a 40-acre forested bog. Profuse coatings and abundant discrete concretions of Mn-Fe oxides containing as much as one percent Pb and one percent Zn have been deposited in sediment of the active stream and flood plain for as much as 1,000 ft downstream from the bog (Nowlan, 1976b).

Mineralization at Bean Brook Mountain occurred mainly as minor sphalerite and galena and traces of chalcopyrite. The sulfides are present in Devonian conglomerate and arkose, in quartz veins that cut the conglomerate and arkose, and in quartz veins cutting felsite dikes that intrude the conglomerate and arkose; the occurrence of sulfides is near the unconformity with the Attean quartz monzonite (Delaney, 1968). The coarse fraction of the sedimentary rocks appears to be derived from the Attean quartz monzonite (Boucot and Heath, 1969). Only minor prospecting has taken place at Bean Brook Mountain. The land surface is well drained.

Mn-Fe Oxides

The influence of deposits of secondary hydrous Mn-Fe oxides on stream-sediment geochemistry in Maine and elsewhere has long been recognized (see, for example, Post and Hite, 1964). In Maine, these Mn-Fe oxides occur as stains and coatings on clastic particles and bedrock, as cement in the interstices of alluvial material and filling fractures in bedrock, and as discrete concretions. Concretionary Mn-Fe oxides are defined as cement, discrete concretions, and coatings that are thick enough to be readily scraped from their host by hand (Nowlan, 1976a). Stains or incipient coatings of Mn-Fe oxides were observed at 80 percent of the sample-collection sites in the Attean area; concretionary deposits were observed at about 12 percent of the sites. Heavy stains and concretionary deposits characteristically form at the interface between oxygenated stream waters and Mn-Fe-charged sediment-pore waters (Nowlan et al., 1983). Therefore, the existence of high concentrations of Mn and Fe in stream sediments may be indicative of certain drainage conditions, of mineralization, or of both.

The concretionary deposits have concentrations of Mn plus Fe ranging from a few percent to as much as 50 percent. Because the Mn oxides and Fe oxides usually are intermixed, it is convenient to speak of them as Mn-Fe oxides. However, Mn oxides are generally much more efficient scavengers of trace elements than are the associated Fe oxides (Nowlan, 1976a). The concentration of Mn in an individual sample will have a bearing on whether that sample is classified as background or anomalous, especially with respect to Zn. Studies strongly suggest that in the Attean area and other parts of Maine, Zn is readily scavenged by Mn-Fe oxides even in non-mineralized areas, but Cu, Pb, and Mo are readily scavenged only in the vicinity of mineralized rock (Nowlan, 1976a). However, Nowlan et al. (in press, b) suggest that Zn concentrations of 200 parts per million (ppm) or greater are probably significant even when the Mn concentration is high. Therefore, scavenging of elements by Mn-Fe oxides does not appear to present serious problems for regional interpretation in the Attean area; regional anomalies of Cu, Pb, Zn, Mo, or Ag appear to be true anomalies whether or not anomalous Mn is present. From both regional and individual-sample viewpoints, scavenging by Mn-Fe oxides should be looked at as a distinctly beneficial phenomenon. An element must be available to be scavenged and, intuitively, mineralization means availability.

SAMPLE COLLECTION, PREPARATION, AND ANALYSIS

Samples were collected at 497 sites and the average site density was about one for each 0.7 mi^2. Samples were collected primarily from first-order streams. In addition, samples were collected about every mile along larger streams. A well-developed drainage system exists in the Attean area and, despite the abundance of bogs, adequate amounts of silt-sized material were available at most sites. Organic muck and material with obvious concentrations of concretionary Mn-Fe oxides were avoided as much as possible. Samples were placed in kraft paper envelopes, dried, and then sieved through 60-mesh (0.25 mm) stainless-steel sieves. The fine material was split and one split was analyzed for Zn by atomic absorption spectroscopy after digestion in hot nitric acid using the method described by Ward et al. (1969). The other split was pulverized to minus-100 mesh (0.15 mm) and analyzed for 31 elements by emission spectrography according to the method described by Grimes and Marranzino (1968).

Zinc values were reported in 5-ppm increments for concentrations less than 100 ppm and in 10-ppm increments for concentrations of 100 ppm or greater. The precision of the Zn method, based on the analysis of six reference samples, is from one to five percent relative standard deviation (O'Leary and Meier, 1986). Results from the emission spectrographic method were reported as one of the numbers 10, 15, 20, 30, 50, and 70 multiplied by some power of ten. The precision of the spectrographic method is approximately plus or minus one reporting interval at the 83 percent confidence level and plus or

TABLE 1. STATISTICAL SUMMARY OF ANALYSES OF STREAM-SEDIMENT SAMPLES COLLECTED FROM TERRAIN UNDERLAIN MOSTLY BY THE ATTEAN QUARTZ MONZONITE, SOMERSET AND FRANKLIN COUNTIES, MAINE

ELEMENT	NUMBER OF SAMPLES	LIMITS OF DETERMINATION, ppm		MINIMUM, ppm	MAXIMUM, ppm	GEOMETRIC MEAN	GEOMETRIC DEVIATION	NUMBER OF QUALIFIED VALUES			REPLACEMENT VALUES, ppm		
		LOWER	UPPER					N	L	G	N	L	G
Cu	497	5	20,000	<5	1,000	12	2.0	0	24	0	–	3	–
Pb	497	10	20,000	N	300	26	1.7	2	2	0	5	7	–
Zn	494	5	–[a]	5	750	71	1.9	0	0	0	–	–	–
Mo	497	5	2,000	N	100	–	–	416	22	0	2	3	–
Ag	497	0.5	5,000	N	7	–	–	455	21	0	0.2	0.3	–
Mn	497	10	5,000	100	G	1,000	2.4	0	0	19	–	–	7,000

N – not detected at lower limit of determination.
L – detected but less than lower limit of determination.
G – greater than upper limit of determination.
[a]Upper limit for Zn by atomic absorption is open ended because sample weight is decreased when necessary.

minus two reporting intervals at the 96 percent confidence level (Motooka and Grimes, 1976). The results of analyses are tabulated by Domenico and Nowlan (1984).

DATA TREATMENT

The U.S. Geological Survey STATPAC system (VanTrump and Miesch, 1977) was used for data management, statistical procedures, and production of geochemical symbol plots. Table 1 is a statistical summary of data for the elements presented in this report. In order for all samples to be included in certain mathematical manipulations such as correlation analysis or factor analysis, values outside the determination range of the analytical method were replaced by arbitrary values. These replacement values (Table 1) were set at two spectrographic steps below the lower limit of determination when the element was not detected, one spectrographic step below the lower limit when the element was detected in an amount less than the lower limit, and one spectrographic step above the upper limit when the element was present in amounts greater than the upper limit. Table 1 lists replacement values for Cu, Pb, Mo, Ag, and Mn.

Factor analysis is a mathematical technique designed to re-express a group of variables in terms of fewer and more representative associations or factors. Factor analysis in geology is discussed by Davis (1973) and in exploration geochemistry by Howarth and Sinding-Larsen (1983). In the present study, R-mode factor analysis was used in which a matrix of correlation coefficients having values of one in the diagonal was used as input to the factor analysis. Only those variables with less than 20 percent of the values qualified by "not detected" or "greater than" were included in the factor analysis, as suggested by Miesch (1976). Elements included in the factor analysis were Fe, Mg, Ca, Ti, Mn, B, Be, Co, Cu, La, Ni, Pb, Sc, Sr, V, Y, Zr, and Zn. Varimax rotations were performed on logarithmically transformed data. Three- to eight-factor models were generated and the 6-factor solution was the most interpretable. Loadings from the 6-factor solution are listed in Table 2. The association of elements in each of the factors in Table 2 suggests that a loading of about 0.40 and greater is significant and that absolute values of 0.30 and greater are probably significant.

Table 3 compares percentile values between the 497 samples of this report and the 6,931 samples from the Sherbrooke and Lewiston 1° by 2° quadrangles (Nowlan et al., in press, a, b). Table 4 separates the analytical values into background and anomalous categories. The anomalous categories for Cu, Pb, Zn, Mo, Mn, and factor 6 are based on the geochemical patterns in the vicinity of mineralized areas. Any detectable amount of Ag is considered anomalous because only 42 samples contain enough Ag to be detectable.

GEOCHEMICAL PATTERNS AND THEIR SIGNIFICANCE

Figures 3-9 show the areal distribution of Cu, Pb, Zn, Mo, Ag, Mn, and factor-6 scores, based on analyses of stream-sediment samples. Anomalous areas, identified by groupings of anomalous samples, are labeled by capital letters on Figures 3-9. Boundaries of the anomalous areas are drawn so as to roughly include the drainage basins of sampling sites plotted within the anomalous area.

Geochemical patterns of some elements in the Sherbrooke and Lewiston 1° by 2° quadrangles were strongly influenced by Pleistocene continental glaciation. In particular, a linear trend of high concentrations of Cr in stream-sediment samples crosses the quadrangles from northwest to southeast. This trend is almost certainly the result of the glacial transport of material derived from mafic and ultramafic bedrock in Quebec (Nowlan et al., 1987). Therefore, patterns of other elements in the Attean area may be the result of glacial transport of material from remote areas.

Although glaciation has undoubtedly modified geochemical patterns in the Attean area, several lines of evidence support the

TABLE 2. LOADINGS FOR A 6-FACTOR R-MODE VARIMAX FACTOR ANALYSIS OF ANALYTICAL DATA FOR 494 SAMPLES OF MINUS-60 MESH STREAM SEDIMENT COLLECTED FROM TERRAIN UNDERLAIN MOSTLY BY THE ATTEAN QUARTZ MONZONITE, SOMERSET AND FRANKLIN COUNTIES, MAINE

Factor	1		2		3		4		5		6	
	element	loading	element	loading	element	loading	element	loading	element	loading	element	loading
Element and loading	Mg	85	Mn	89	La	77	Ca	87	Be	90	Cu	69
	V	84	Co	80	Y	77	Sr	82	Zr	47	Pb	56
	Ti	79	Fe	49	Sc	52	Sc	23	Sr	23	Zn	47
	Fe	73	Zn	47	Cu	40	Be	13	Y	23	B	30
	Ni	71	Pb	45	Zn	28	Zr	13	Zn	23	Mg	22
	Zr	66	Ni	33	Ti	28	V	9	Pb	19	Ni	22
	B	61	Sc	28	Ni	23	Ti	9	B	17	La	17
	Sc	48	V	27	V	15	Pb	5	Ni	3	Fe	17
	Co	37	Cu	19	Ca	14	La	3	Co	1	Mn	16
	Y	31	Y	13	Zr	12	Mg	1	V	0	Ca	13
	Cu	17	Be	7	Mn	5	Y	0	Fe	−1	Be	8
	Pb	16	Ti	6	Co	5	Cu	−3	Ca	−1	Co	8
	Sr	5	Sr	−1	Mg	4	Mn	−5	Mn	−1	V	2
	La	4	Mg	−1	Be	2	Fe	−5	La	−5	Sc	0
	Mn	4	La	−6	Fe	0	Co	−9	Cu	−10	Y	−4
	Ca	−0	Ca	−13	Sr	−10	Ni	−30	Ti	−11	Ti	−8
	Zn	−8	Zr	−14	B	−30	B	−31	Sc	−18	Sr	−18
	Be	−14	B	−26	Pb	−33	Zn	−36	Mg	−18	Zr	−19
Presumed geochemical associations	Detrital mafic minerals		Manganese scavenging		Rare-earth minerals		Calcareous rocks(?)		Late differentiates		Sulfide mineralization	

Significant loadings are enclosed in solid boxes. Probable significant loadings are enclosed in dashed boxes. Loadings have been multiplied by 100.

TABLE 3. PERCENTILE VALUES FOR STREAM SEDIMENT SAMPLES FROM TERRAIN UNDERLAIN MOSTLY BY THE ATTEAN QUARTZ MONZONITE (A), SOMERSET AND FRANKLIN COUNTIES, MAINE, AND FROM THE SHERBROOKE AND LEWISTON 1° BY 2° QUADRANGLES (SL), MAINE, NEW HAMPSHIRE, AND VERMONT

Variable	Number of samples		Percentile							
			50		75		90		95	
	A	SL	A	SL	A	SL	A	SL	A	SL
Cu	497	6,925	10	11	15	17	20	26	50	35
Pb	497	6,925	30	19	30	26	50	35	70	44
Zn	494	6,838	75	50	100	73	150	110	180	130
Mo	497	6,931	N(5)	–	N(5)	2.9	5	4.1	7	6.5
Ag	497	5,753	N(0.5)	–	N(0.5)	–	N(0.5)	–	<0.5	–*
Mn	497	6,930	1,000	960	1,500	1,500	3,000	2,500	5,000	3,400
Factor 6	494	–	−0.036	–	0.508	–	1.165	–	1.603	–

N – not detected at value shown in parentheses.
< – detected but less than value shown.
– – not determined.
Values are in ppm except for factor 6 values which are scores.
*99th percentile is 1 ppm.

TABLE 4. RANGES OF VALUES ASSIGNED TO BACKGROUND AND ANOMALOUS CATEGORIES FOR STREAM-SEDIMENT SAMPLES COLLECTED FROM TERRAIN UNDERLAIN MOSTLY BY THE ATTEAN QUARTZ MONZONITE, SOMERSET AND FRANKLIN COUNTIES, MAINE

Variable	Category		Mineralized areas used as basis for establishing anomalous category
	Background	Anomalous	
Cu	<5-30	50-1,000	Catheart Mountain and west Sally Mountain
Pb	N(10)-30	50-300	Pyrite Creek and Bean Brook Mountain
Zn	5-160	170-750	Pyrite Creek and Bean Brook Mountain
Mo	N(5)-5	7-100	Catheart Mountain
Ag	N(0.5)	<0.5-7	None. Any detectable amount is anomalous
Mn	100-3,000	5,000->5,000	Pyrite Creek
Factor 6	−3.30-1.00	1.01-5.10	Catheart Mountain, west Sally Mountain, Pyrite Creek, and Bean Brook Mountain

N – not detected at value shown in parentheses.
< – detected but in amounts less than value shown.
Values are in ppm except for factor 6 which are scores. All reported analytical values are in ranges listed.

concept that the patterns on Figures 3-9 are locally derived. (1) Known mineral deposits and occurrences of Cu, Pb, Zn, Mo, and Ag are areally or structurally associated with the Attean quartz monzonite. (2) Patterns of high values of Cu, Pb, Zn, and Mo are areally associated with the Attean quartz monzonite when viewed in the larger setting of the Sherbrooke and Lewiston quadrangles (Nowlan et al., in press, a, b). (3) Stream-sediment geochemical anomalies of Pb, Zn, and Mo are areally associated with other members of the Highlandcroft Plutonic Suite, in addition to the Attean quartz monzonite, elsewhere in the Sherbrooke and Lewiston quadrangles (Nowlan et al., in press, a, b). Permissive evidence is that most of the geochemical patterns on Figures 3-9 that parallel the general southeastward movement of Pleistocene glaciers also parallel structural or lithologic features.

The distribution of Cu is shown on Figure 3. The occurrences of porphyry Cu mineralization (Catheart Mountain and west Sally Mountain) have associated Cu anomalies (areas A and B). Samples collected from stream beds on the east flank of Catheart Mountain do not have anomalous concentrations of Cu. Except at Catheart Mountain, samples containing anomalous concentrations of Cu are generally from sites within a few hundred feet of faults, or else the site is a mile or two downstream from where a fault crosses the drainage basin.

Figure 4 shows the distribution of Pb. All of the samples containing 150 ppm Pb or greater are from the Pyrite Creek area (area A); scavenging by Mn-Fe oxides may have enhanced these values. A number of samples from the Bean Brook Mountain area (area B) have concentrations of 50-100 ppm. The samples in anomalous area C are mostly from streams that drain the unconformity between Upper Ordovician Attean quartz monzonite and younger metasedimentary rocks. Anomalous area D includes samples from streams draining the western mineralized area on Sally Mountain; the large size of area D is remarkable, considering the small extent of known mineralization. Anomalous area E roughly coincides with the Skinner pluton.

Figure 5 shows the distribution of Zn. Area A includes Pb-Zn mineralization at Pyrite Creek. A few samples from Catheart Mountain (area B) and west Sally Mountain (area C) contain anomalous concentrations of Zn. The linear outline of area D parallels the east-northeast orientation of the faults north of area D and roughly coincides with an anomalous area based on Mn (Fig. 8). Samples collected in area E are mostly from streams draining the unconformity between granite and younger metasedimentary rocks.

The distribution of Mo is shown on Figure 6. Area A includes streams draining all sides of Catheart Mountain, in contrast to Cu which is not anomalous in streams draining the east flank (Fig. 3). Analyses of samples from west Sally Mountain (area B) reflect known Mo mineralization. Molybdenum mineralization is unknown in areas C and D but their proximity to area A greatly enlarges the geochemical anomaly areally related to Catheart Mountain

Figure 7 shows the distribution of Ag. Areas A and B include the Pyrite Creek, Bean Brook Mountain, and Catheart Mountain mineralized areas. The two samples at Catheart Mountain (area B) that contain the highest concentrations of Ag also contain the highest concentrations of Cu (1,000 ppm) and Mo (100 ppm). The sample with the highest concentration of Ag (7 ppm) is from area D.

Figure 8 shows the distribution of Mn. Concentrations of Mn in stream sediments above the median concentration of 1,000 ppm (Table 3) indicate that formation of secondary Mn-Fe oxides in the stream sediment has probably occurred. Anomalous Mn is conspicuously absent from the Catheart Mountain, Sally Mountain, and Bean Brook Mountain mineralized areas, perhaps because they are generally well drained and conditions are therefore not conducive to the formation of anomalous amounts.

The distribution of scores for factor 6 is shown on Figure 9. The consistent and uniform patterns of anomalous scores for samples from the mineralized areas support the results of factor

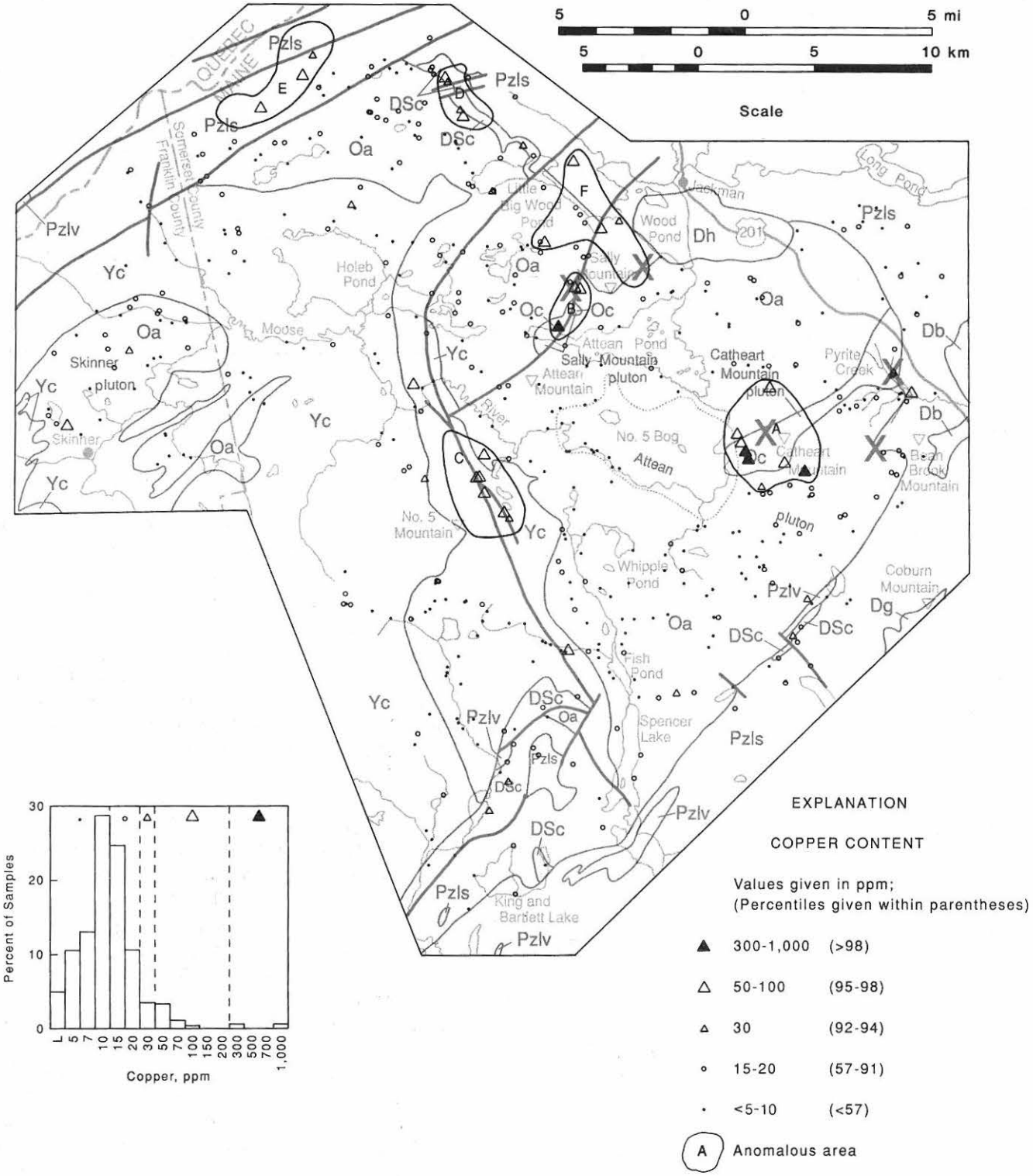

Figure 3. Copper distribution in stream sediments, Attean area, Somerset and Franklin Counties, Maine.

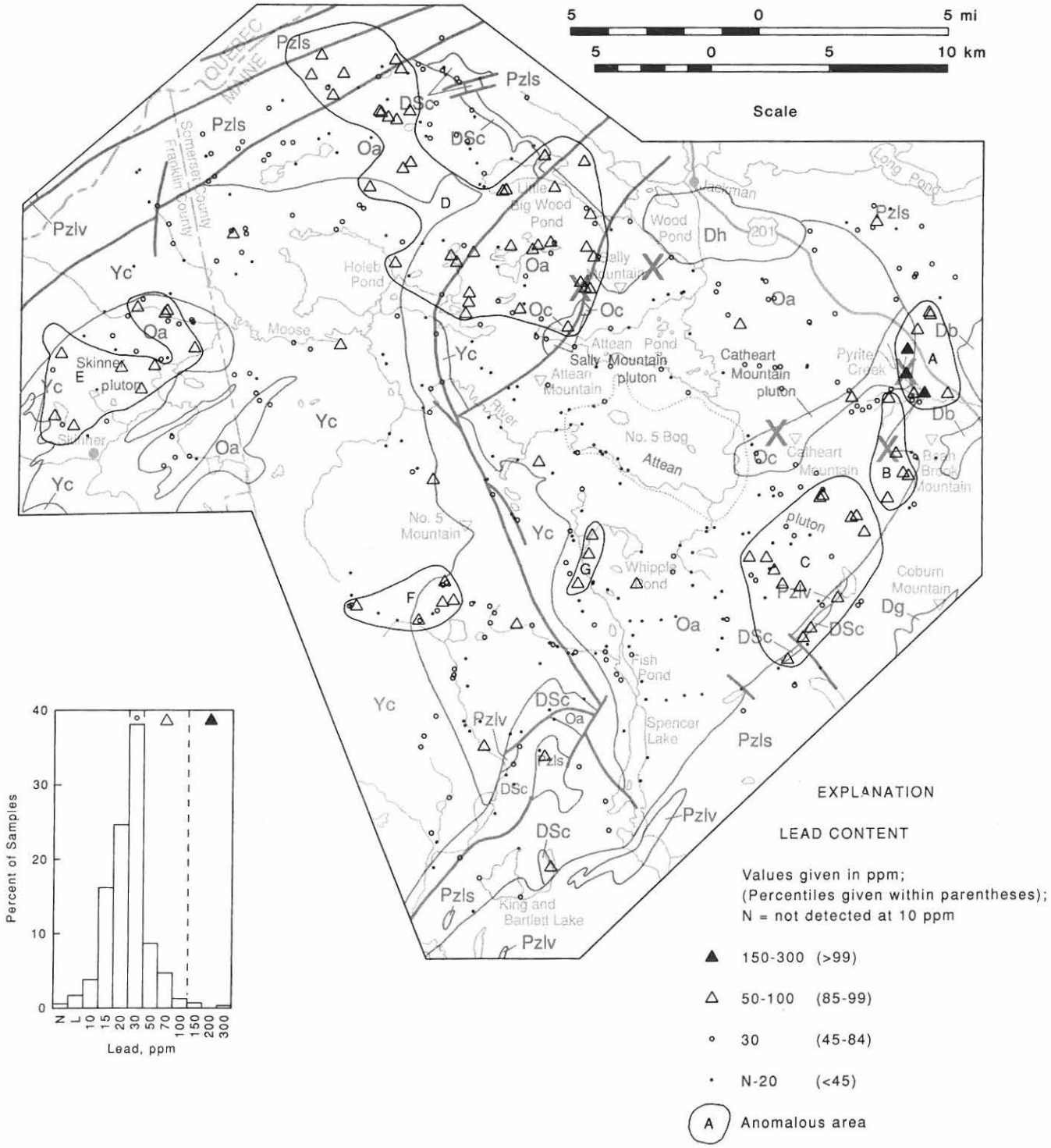

Figure 4. Lead distribution in stream sediments, Attean area, Somerset and Franklin Counties, Maine.

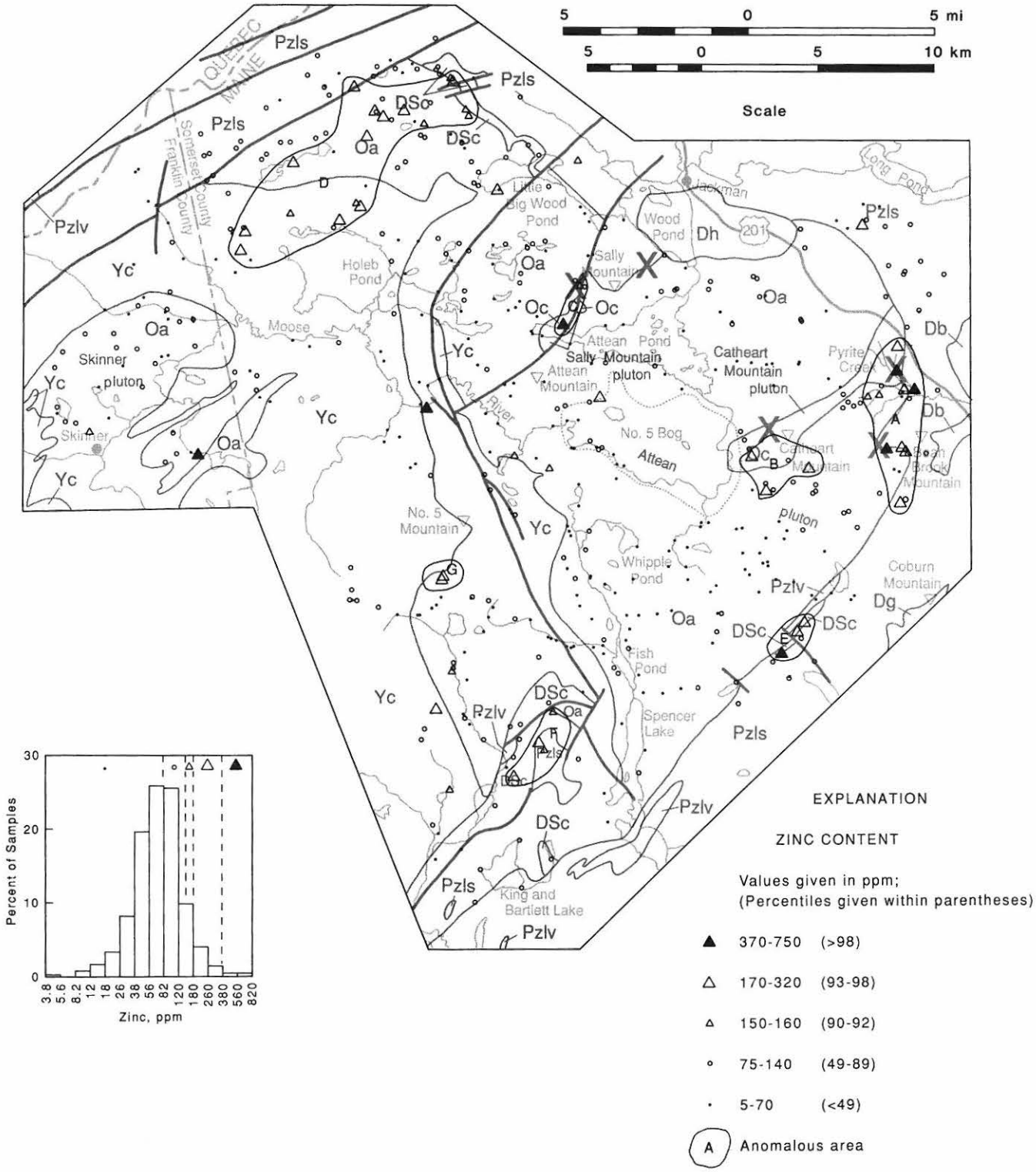

Figure 5. Zinc distribution in stream sediments, Attean area, Somerset and Franklin Counties, Maine.

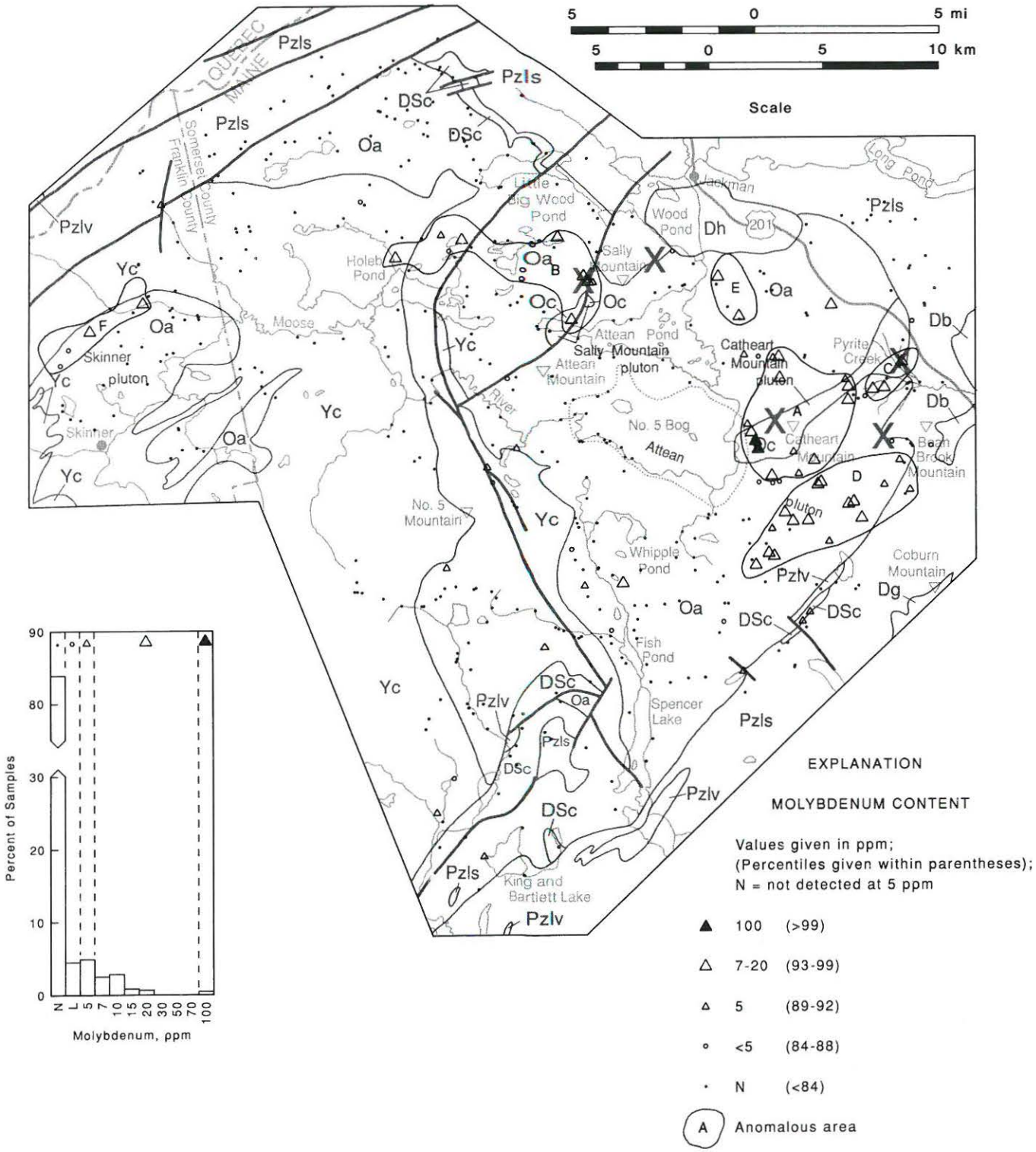

Figure 6. Molybdenum distribution in stream sediments, Attean area, Somerset and Franklin Counties, Maine.

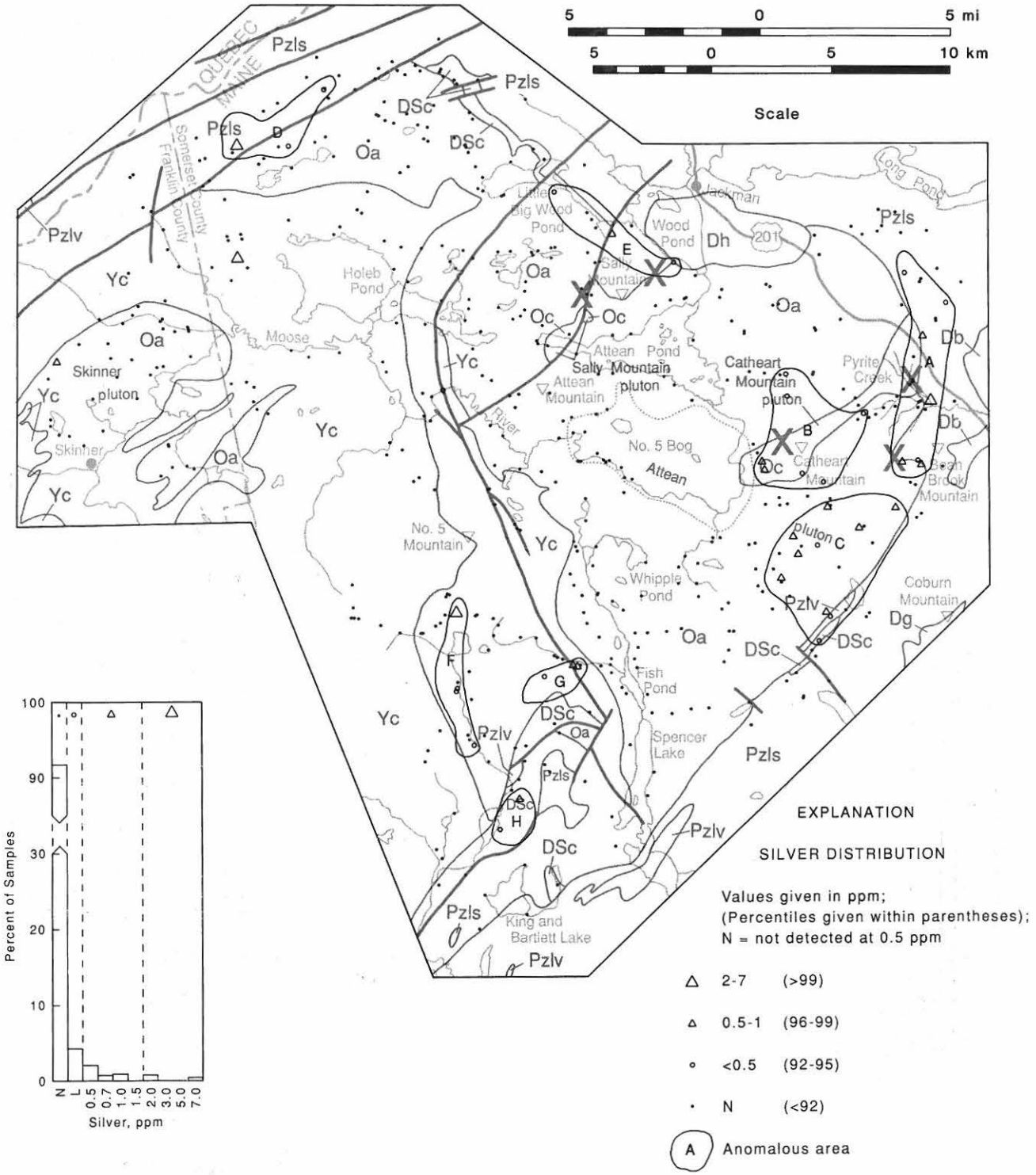

Figure 7. Silver distribution in stream sediments, Attean area, Somerset and Franklin Counties, Maine.

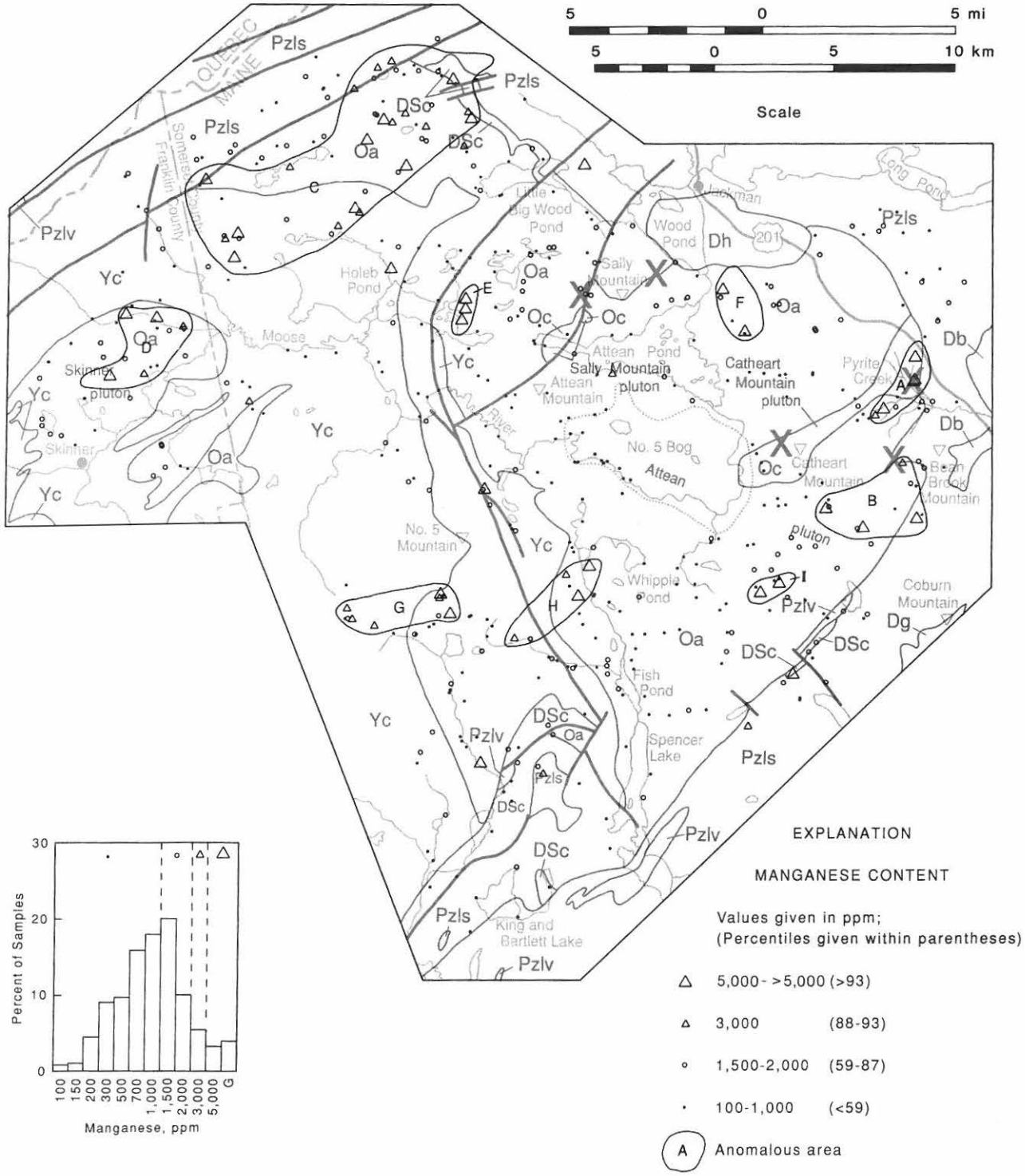

Figure 8. Manganese distribution in stream sediments, Attean area, Somerset and Franklin Counties, Maine.

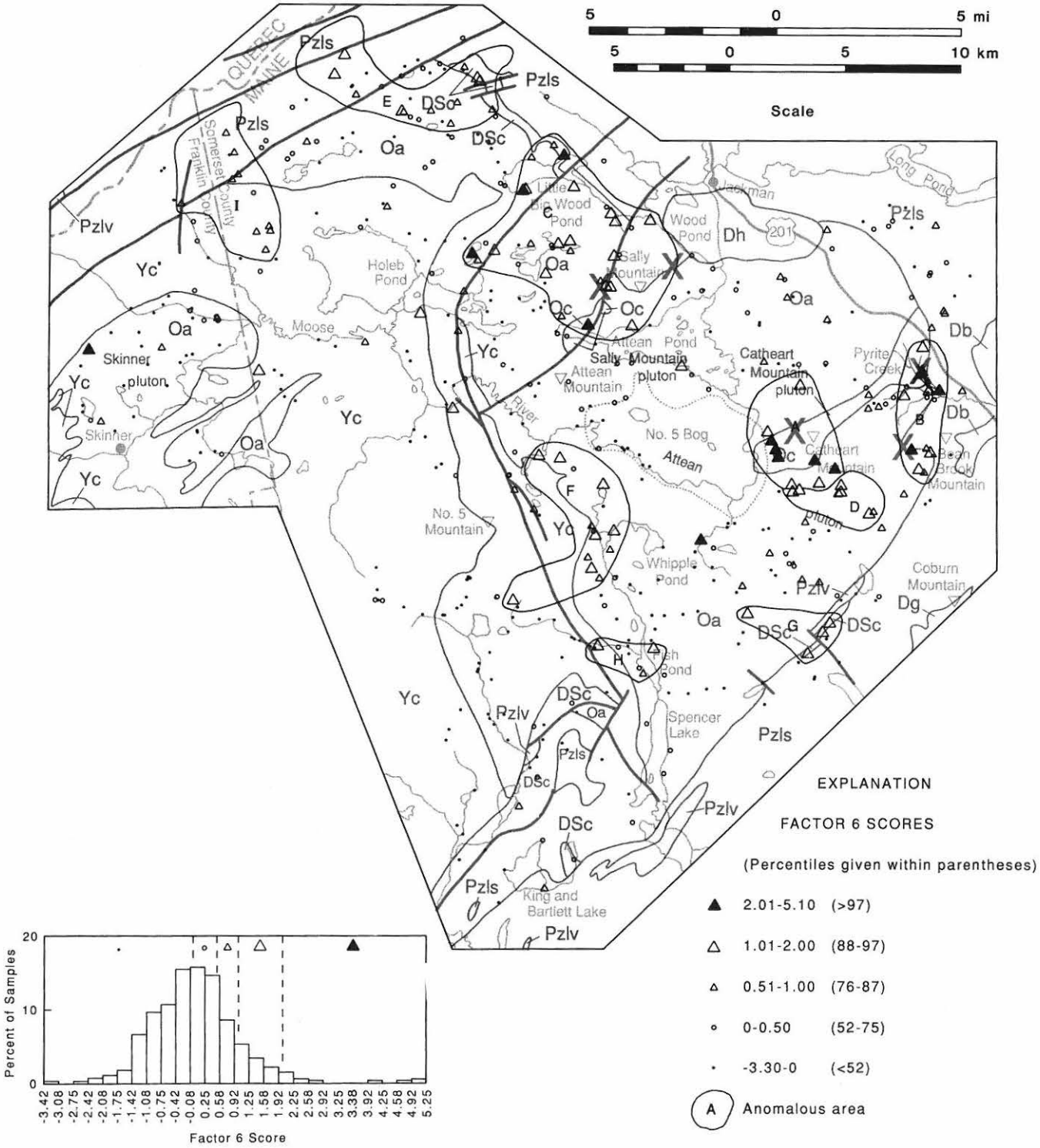

Figure 9. Distributions of scores for factor 6 of R-mode factor analysis of analytical data from stream sediments, Attean area, Somerset and Franklin Counties, Maine.

analysis. The mineralized areas at Sally Mountain are part of a much more extensive geochemical anomaly shown on Figure 9 as area C. Area D adjoins area A and area B is close enough that the entire Catheart Mountain-Pyrite Creek-Bean Brook Mountain area may constitute one large geochemical anomaly; if threshold scores between areas D and G are included, then area G becomes part of that large geochemical anomaly. Anomalous area I is a distinctive cluster of threshold scores.

Figure 10 is a compilation of the results shown in Figures 3-9; it shows relative favorabilities for the existence of additional, undiscovered mineral deposits in the study area. The rankings take into account the presence of known mineral deposits, the coincidence or overlapping of anomalous areas determined from plots of single elements, and the results of factor analysis. Of the areas outlined, region 1 has the highest favorability and region 9 the lowest. The regions are large target areas where geology and geochemistry suggest conditions are favorable for the existence of certain types of mineral deposits. Anomalous patterns or samples may or may not reflect individual mineral deposits.

Region 1 includes the Catheart Mountain Cu-Mo porphyry deposit. Samples from this region have anomalous concentrations of Mo, Cu, Ag, Zn, and Pb, and scores for factor 6 are anomalous. The data suggest that porphyry mineralization may have extended further north, south, and east than is presently known, as proposed by Schmidt (1974; 1978).

Region 2 is spatially related to the unconformity between the Attean quartz monzonite and overlying metasedimentary rocks that are predominantly Devonian siliciclastic rocks. The vein-type, base-metal mineral deposits at Pyrite Creek and Bean Brook Mountain are in region 2. Samples from this region have anomalous concentrations of Pb, Ag, Mo, Zn, Mn, and Cu. Scores from factor 6 show three anomalies in region 2. This association of elements is compatible with the possible existence of two types of mineral deposits that are, at present, unrecognized in the Attean area: sediment-hosted stratiform deposits and sandstone Pb deposits.

According to Gustafson and Williams (1981), sediment-hosted stratiform deposits of Cu, Pb, and Zn tend to be either Cu deposits or Pb-Zn deposits. Copper is relatively unimportant in samples from region 2 while Zn, and especially Pb, are relatively important.

The importance of Pb in region 2 is compatible with the possible existence of sandstone Pb deposits; the geology also is favorable. According to Bjørlykke and Sangster (1981), such deposits characteristically form in basal quartzitic sandstones deposited on sialic basement rocks. Sometimes the granitic basement rocks are relatively rich in Pb, compared to average granite. Bjørlykke and Sangster (1981) suggest a genetic model that envisions red-bed Cu, sandstone Pb, and carbonate-hosted Zn-Pb as related but distinct deposit types. The three types formed at different stages of crustal evolution, according to the model. Red-bed Cu deposits were deposited first, in a continental environment when the area was more tectonically active. Sandstone Pb deposits formed later in the continental-shallow marine environment during a tectonically stable period when fairly intense and prolonged chemical weathering and peneplanation took place. Carbonate-hosted Zn-Pb deposits were deposited later or at the same time as the sandstone deposits in a shallow-marine environment during transgression.

The Attean area appears to meet the geologic and geochemical criteria for the existence of sandstone Pb deposits closely enough that further study is warranted. Sediments from streams draining the area underlain by Attean quartz monzonite tend to have higher concentrations of Pb than samples from many other granitic terranes in the Sherbrooke and Lewiston 1° by 2° quadrangles (Nowlan et al., in press, b). Boucot and Heath (1969) present evidence that suggests that the Attean area was part of an island in much of Late Silurian and Early Devonian time. The evidence includes the widely differing ages of rocks that rest unconformably on the Attean pluton, the lithofacies relationships within the various units that rest on the Attean pluton, scattered occurrences of conglomerate that are interpreted as sea cliffs, and the presence of fossil shell communities consistent with a shoreline or near-shoreline environment. Parts of the proposed island had very low relief and parts had higher relief as determined from the coarseness and composition of metasedimentary rocks unconformably overlying the Attean pluton.

Basal sandstones and conglomerates that might serve as hosts to Pb deposits are the conglomerate at Foxs Camp (Silurian), the Hobbstown Formation (Upper Silurian to Lower Devonian), and sandstone beds of the Lower Devonian Seboomook and Tarratine Formations, which are all included in unit Pzls on Figure 2. Stratigraphically equivalent or higher calcareous shallow marine rocks that fit the model of Bjørlykke and Sangster (1981) are represented by the Beck Pond Limestone (Lower Devonian) and the Lower Devonian McKenney Ponds Limestone Member of the Tarratine Formation which are included in unit DSc on Figure 2. Red beds are not present in the Attean area but unconformably overlie the pre-Silurian Lobster Mountain Volcanics (included in unit Pzlv on Fig. 2) 30 miles northeast of the Attean area; Boucot and Heath (1969) interpret the red beds (Upper Silurian Big Claw Redbed Member of the Lobster Lake Formation) as non-marine sediments that were deposited in an oxidizing environment.

The discussion of sediment-hosted deposits does not rule out the existence of vein-type base-metal mineral deposits in region 2 in addition to the known occurrences at Pyrite Creek and Bean Brook Mountain. Another way of looking at regions 1 and 2 is to picture the classic porphyry model where a central core of Cu and/or Mo is surrounded by deposits of base metals (Jerome, 1966). However, mineralization at Bean Brook Mountain was much later and could not have been part of the primary porphyry model.

Region 3 includes the two mineral occurrences at Sally Mountain. Samples from the region have anomalous concentrations of Pb, Mo, Cu, Zn, Ag, and Mn. Much of region 3 is anomalous, based on scores from factor 6. The geochemistry of

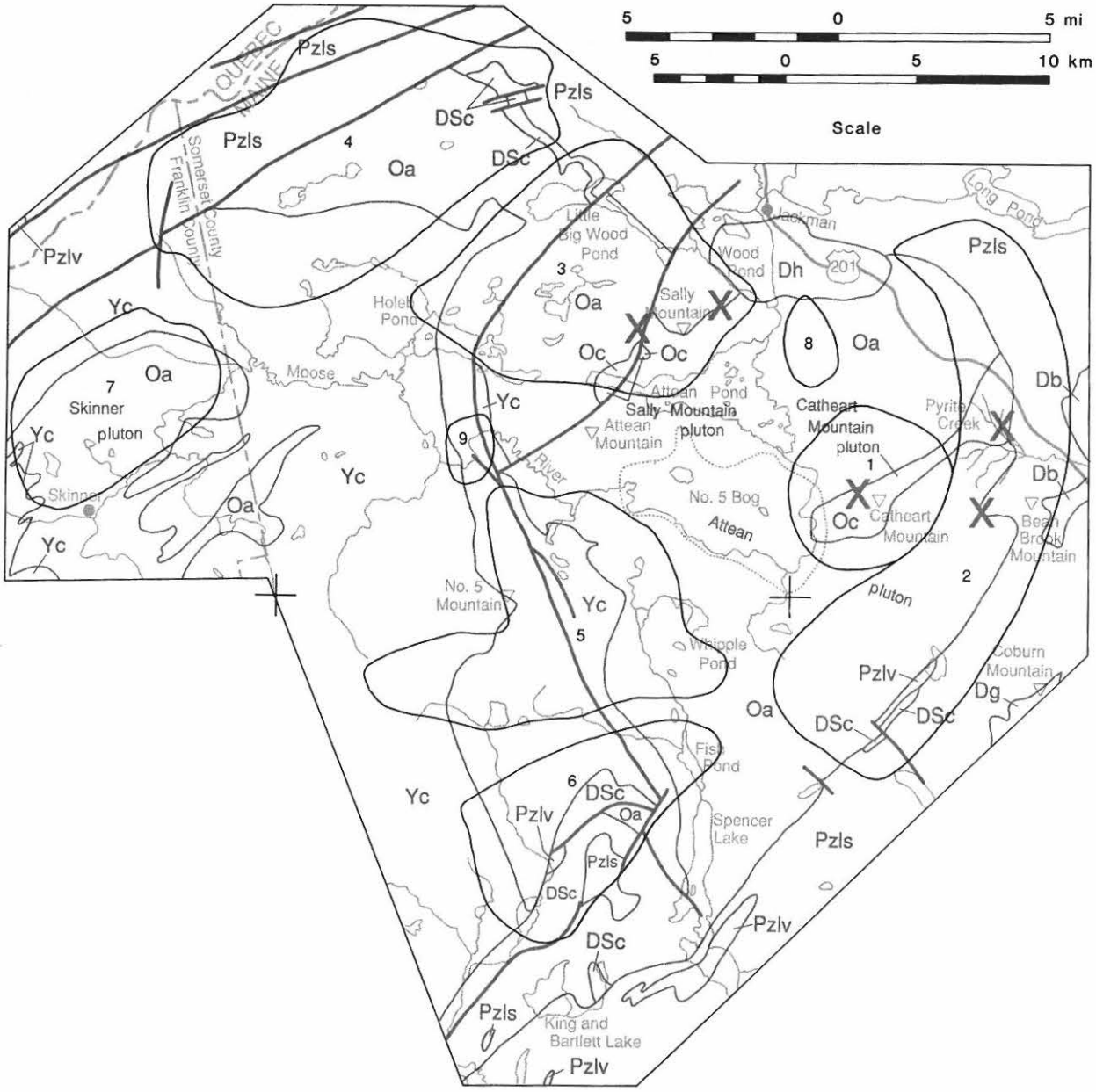

Figure 10. Regions favorable for the existence of additional, undiscovered mineral deposits, Attean area, Somerset and Franklin Counties, Maine. Region 1 is the most favorable and region 9 is the least favorable.

samples from region 3 suggests the possibility of several types of mineralization. Areas of anomalous concentrations of Cu and Mo suggest that porphyry-type mineralization may have taken place in addition to the known mineralization at west Sally Mountain; however, the Cu and Mo anomalies do not coincide as at Sally Mountain. The Cu anomalies are spatially related to faults, suggesting structural control of mineralization. As in region 2, the presence of the unconformity, where granite of the Upper Ordovician Attean quartz monzonite is overlain by Silurian and Devonian metasedimentary rocks, suggests the possibility of sediment-hosted deposits.

In region 4, Zn, Mn, Pb, Cu, and Ag concentrations are anomalous. Two areas within region 4 were deemed to be anomalous on the basis of scores from factor 6. Molybdenum was not detected in any sample collected from region 4, suggesting that porphyry-type mineralization probably did not take place. The elongated patterns of samples with anomalous concentrations of Zn and Mn that parallel the east-northeast-trending fault between Upper Ordovician Attean quartz monzonite and Silurian and Devonian siliciclastic metasedimentary rocks suggest the possibility of structurally controlled Cu, Pb, or Zn base-metal deposits. Region 4 includes terrain underlain by both

siliciclastic and calcareous metasedimentary rocks, so sediment-hosted Cu, Pb, or Zn base-metal deposits are a possibility.

The Frontenac Formation (Ordovician(?) and Silurian(?)) (included in unit Pzls on Fig. 2) occupies the north and northwest margins of the study area. The Frontenac Formation is host to volcanogenic massive-sulfide deposits 15-30 miles southwest of the study area at Ledge Ridge in Maine and at the Clinton River in Quebec (Stephens et al., 1984). The small area shown as unit Pzlv at the northwest edge of the study area (Fig. 2) is underlain by mafic volcanic rocks of the Frontenac Formation, but the possibility of felsic volcanic rocks and associated massive sulfide deposits within the Frontenac Formation near the study area should not be ruled out. If volcanogenic massive-sulfide deposits did form, they could have been remobilized along the faults that cross region 4 or are nearby.

Anomalous areas in region 5 are spatially related to faults and to the intrusive contact between the Upper Ordovician Attean quartz monzonite and the Middle(?) Proterozoic Chain Lakes massif. Concentrations of Pb, Mn, Cu, Ag, and Zn are anomalous. Structurally controlled base-metal veins would appear to be a possible type of deposit in that region.

Region 6 covers an area where a variety of rock types occur in a relatively small area, including Middle(?) Proterozoic, Silurian, and Devonian siliciclastic metasedimentary rocks, pre-Silurian volcanic rocks, Ordovician granite, and Silurian and Devonian calcareous metasedimentary rocks. A number of faults intersect in region 6. Concentrations of Ag, Zn, Pb, Cu, and Mn are anomalous. Structurally controlled veins and skarn base- and precious-metal deposits are possibilities.

Region 7 is spatially associated with the Skinner pluton. Concentrations of Pb, Mn, Mo, and Cu are anomalous. Base-metal veins are the most plausible type of mineral deposit to be expected. A few samples contain anomalous amounts of Mo or Cu so there is some possibility of porphyry deposits.

In region 8, only Mo, Mn, and Pb are present in anomalous concentrations. The anomalous Mo suggests porphyry-type mineralization. Another possibility is that Mo mineralization took place near the intrusive contact between the Upper Ordovician Attean quartz monzonite and the Lower or Middle Devonian Hog Island granodiorite (unit Dh), a feature associated with other granitic intrusive bodies in New England (Schmidt, 1978).

In region 9, Zn, Cu, and Ag are present in anomalous concentrations. Proximity of region 9 to faults suggests that structurally controlled veins could be present.

CONCLUSIONS

Except at Catheart Mountain, the observed amount of mineralized rock is not sufficient to explain the high concentrations of heavy metals in stream-sediment samples from streams draining the mineralized areas. The results of this study suggest that, on the basis of geology and stream-sediment geochemistry, heretofore unrecognized mineral deposits may exist in the Attean area and mineralization associated with known occurrences may have been more extensive than has previously been recognized. The geology and stream-sediment geochemistry are compatible with the possible presence of additional deposits of Cu-Mo porphyry and Cu, Pb, Zn base-metal veins. Sandstone-type Pb deposits, sediment-hosted stratiform Cu, Pb, or Zn base-metal deposits, skarn deposits of Cu, Pb, Zn, Ag, and Au, and volcanogenic massive-sulfide deposits could be present although none have yet been recognized in the area. The prime target areas are faults, the unconformity between Upper Ordovician Attean quartz monzonite and overlying younger metasedimentary rocks, and terrain around known mineral occurrences.

ACKNOWLEDGMENTS

Access to their lands and permission to publish the results were kindly granted by Coburn Lands Trust, D'Auteil Lumber Co., Great Northern Paper Co., Megantic Manufacturing Co., Scott Paper Co., Mr. Gary Merrill, GCO Minerals Co., and Timberland, Inc. Beren Harrington, J. T. Kline, and Andrew Sprecher assisted with fieldwork. Analysts were James A. Domenico, Carl Forn, Stephen Sutley, E. Carson, D. Rohlf, G. C. VanGaalen, R. Vaughn, and D. L. Kelley. Reviews by F. C. Canney and T. D. Light resulted in a greatly improved paper.

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