Studies in Maine Geology

Volume 1: Structure and Stratigraphy

Edited by Robert D. Tucker and Robert G. Marvinney

Walter A. Anderson, State Geologist
Maine Geological Survey
DEPARTMENT OF CONSERVATION

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Robert D. Tucker
Director, Cartography and Information Services
Maine Geological Survey

Robert G. Marvinney
Physical Geologist
Maine Geological Survey

Walter A. Anderson, State Geologist

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Cover photograph:

Tightly folded coticule bands in biotite granofels of the Cape Elizabeth Formation, Georgetown, Maine. Photo by Arthur M. Hussey II.

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

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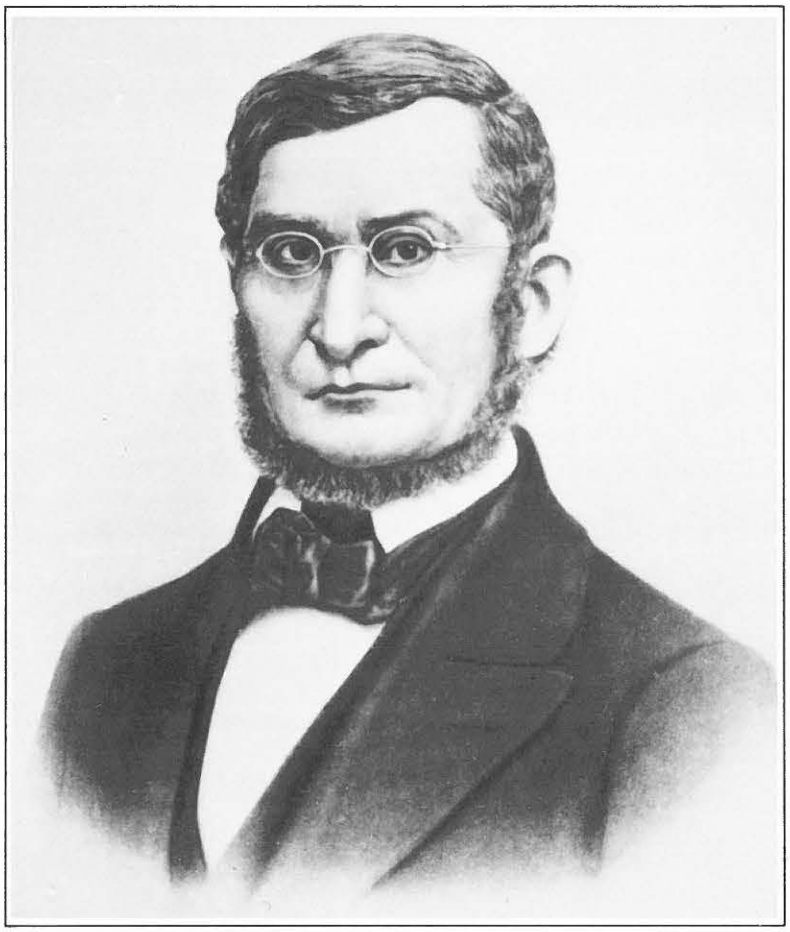

Charles Thomas Jackson (1805-1880)

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
Maine Geological Survey
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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Charles Thomas Jackson and the First Geological Survey of Maine, 1836-1838

Mark Hineline
Department of Geosciences
University of Southern Maine
Gorham, Maine 04038

ABSTRACT

Most historians of science agree on the need to avoid retrospective judgments when evaluating the work of historical figures in science. Thus, when assessing and interpreting the work of Maine's first state geologist, Charles T. Jackson, one must be careful to do so in light of what was known in the period in which he worked (1836-1839), and not upon "what we now know." As an example, Jackson's work in Maine was contemporary with a revision of the stratigraphic column by the British geologists Adam Sedgwick and Roderick Murchison. As a result of their work, purely lithologic descriptions of older stratified rock as "transition" would be replaced by the Cambrian, Silurian, and Devonian systems, determined in large part by characteristic fossils. Application of these systems was problematic in the structurally and stratigraphically complex geology of Maine; moreover, the naming of formations and the development of a stratigraphic column was a source of contention in American geology throughout this period.

Similarly, Jackson's diluvial interpretation of the glacial drift that occurs in the state went far toward explaining the sources of soils in Maine. The explanatory power of the diluvial theory permitted Jackson to describe phenomena, such as glacial striations, which might have been otherwise overlooked.

Jackson's interpretation of the geology of Maine has been the subject of criticism by historians of science and geologists alike. A fresh look reveals a field researcher, committed both to his task and to his methodology, grappling with the complex geology of Maine.

INTRODUCTION

Between 1837 and 1839 Charles Thomas Jackson (1805-1880) published three reports of a state-financed geological survey of Maine (Fig. 1). These reports make excellent reading and today remain interesting in their own right, but contradictory judgments about their scientific worth pose a problem for the historian of science. Contemporary assessments of Jackson's surveys in Benjamin Silliman's *American Journal of Science* were laudatory (Silliman, 1839). Jackson was a member of the scientific elite in the United States. This, combined with favorable acceptance of his field work and interpretations, suggests his being a competent, respected geologist. Élie de Beaumont, with whom Jackson studied in the early 1830's, makes kind mention of him in the three-volume *Notice sur les Systèmes des Montagnes* (Woodworth, 1897). The obituaries and memorials published shortly after Jackson's unfortunate "insanity" and death may be read alternatively as apologia or as tributes.

Later judgments have been less enthusiastic. Merrill (1904), in his ambitious study of the state surveys, concluded that "These reports, examined in the light of to-day, contain very little which would be considered of geological importance." Aldrich, in a more recent reassessment (1981), reaffirmed the essence of that judgment, but tempered it by suggesting that Jackson "should be credited with providing the science with basic data about states without which other American geologists would have found their work lacking."

The disparity between the contemporary assessment of Jack son's published survey of Maine and later assessments is intrigu ing. Two possible explanations for the disparity will be explored in this paper. The first of these is that Silliman, in praising Jackson in *American Journal of Science*, for example, made an error in judging Jackson's science, and that this error was corrected by later, less biased and more sophisticated observers. Set

Figure 1. Charles Thomas Jackson.

against this interpretation is the second possibility: that the context in which geology was practiced changed significantly between Jackson's time and the period of reassessment, such that the virtues of Jackson's work became obscured while the vices stood out in sharp relief.

These options may be overdrawn. Overstatement of historical judgments appears, however, to be a systematic flaw in the history of geology which, according to Greene (1985), may be characterized by three approaches to writing and interpreting history: the attack, the celebration, and the review. Correction of these obvious biases in the history of geology has lagged behind the remarkable revision carried out in the histories of other sciences, notably the history of physics, over the past twenty years. Recent investigators, notably Greene (1982) and Rudwick (1985) have begun to cement a more rigorous tradition in the history of geology, but as Turner (1986) has pointed out for the case of Rudwick's "non-retrospective" history of the "Devonian controversy," this leads to a transfer and diminution, not an elimination of bias. If Turner is correct in pointing out that "all historical analysis requires that the historian exercise judgment," what, then, are to be the criteria for judgment? A provocative essay by Donovan (1981) suggests two apparently mutually exclusive approaches based upon differences in interest between geologists and historians. Geologists, on the one hand, have sought a history that

> provides object lessons in scientific method and reinforces the belief that modern geology is a fundamentally empirical science. Such a history exults in conceptual conflict. . . . And, if the founding of modern geology is to be seen as the establishment of a discipline, then this approach defines the discipline in terms of its conceptual structures and its subject matter.

In this tradition, debate between Neptunists and Vulcanists is disposed in favor of the latter through recourse to some physical evidence – in this case, the basalts of France. Kuhn (1970) has explained such an approach to the history of science as pedagogically functional but inadequate for several reasons, one of which is the naive realism that bespeaks of "a fundamentally empirical science." Kuhn's criticism of "whiggish" history of science came at a time when social and intellectual historians turned their backs upon naive realism in favor of an overarching theory of social constructivism. The result has been an "externalist" history of science which may leave all empirical evidence for facts, laws, and theories out of focus in the overall depth of field, preferring to envision debates, such as those of the Neptunists and the Vulcanists, the Catastrophists and the Uniformitarianists, as exercises, as Donovan puts it, of "social control." The result of this second tradition, as several critics have suggested, may be a history of science without science (Donovan, 1981; Greene, 1985).

As Donovan has pointed out, neither of these traditions is wholly adequate to characterize the history of a science; both are amplification/reduction devices which pay too high a price for their overall resolving power. Very recently, Greene (1982) and Rudwick (1985) have approached the origins of modern geology in ways that, while differing one from the other in such aspects as narrative technique, nevertheless manage to preserve science while doing justice to the cultural milieu in which science must take place. In his conclusion to *The Great Devonian Controversy*, Rudwick suggests that

> it is possible to see the cumulative empirical evidence in the Devonian debate, *neither* as having determined the result of the research in any unambiguous way, as naive realists might claim, nor as having been virtually irrelevant to the result of the social contest on the agonistic field, as constructivists might maintain. It can be seen instead as having had a *differentiating* effect on the course and outcome of the debate, constraining the social construction into being a limited, but reliable and indefinitely improvable, representation of reality. (Rudwick, 1985, p. 455-456)

Geological interpretation, in other words, may be *underdetermined* by empirical evidence, but it is not *undetermined* by it. Put another way, the rocks are not self-interpreting, but they do *constrain* interpretation in a way that the subject matter of other sciences sometimes does not. With this position as a guiding bias, it should be possible to satisfy the needs of both the geologist and of the professional historian when reconsidering a figure such as C. T. Jackson. The careful historian may attempt to discern the difference, as enhanced by the obvious advantage of hindsight, between errors of applying a chosen methodology and "errors" of methodological choice. In this paper, the scientific context in which Jackson's geological education and survey was carried out provides a background against which to assess Jackson's reports. With the context made partially clear, several of the problems faced by Jackson will be examined in greater detail. Before concluding and reassessing the value of Jackson's work in Maine, several aspects of his

later life are considered as possible "circumstantial" but supporting evidence for the devaluation of Jackson's science by Merrill and others. Although a closer examination of the social and political contexts of the survey might well prove valuable, they are beyond the scope of this paper and remain areas for fruitful study.

THE CONTEXT

After 1800 and through the present day, all periods in geology are both significant and transitional. To isolate any one period and attempt to describe it using these qualifiers would be to say very little. Nevertheless, the years 1830 to 1840 have a special significance for the geology of Maine, for it is in this decade that the stratigraphy of lower Paleozoic rocks coalesced in the work of Adam Sedgwick and Roderick Murchison. Also, Louis Agassiz developed and published his glacial theory. In later years, these developments were applied successfully to the interpretation of Maine geology, but they are virtually absent in the work of C. T. Jackson. The modern reader of Jackson's reports may, accordingly, regard Jackson as having hailed from the rearguard of science and as an unfortunate choice for Maine's first state geologist. The historical record shows, however, that this was not the case. Quite to the contrary, Jackson was at the very core of the social and intellectual center of American geology from 1835 to 1845 (Rudwick, 1985, p. 420-421). An analysis of Jackson's competence and influence during the decade of the Maine, Rhode Island, and New Hampshire surveys shows him to have been among the elite of American geologists. Jackson was among Governor Marcy's first choices for head geologist of the New York survey, but apparently was passed over because Marcy felt he could not afford Jackson's services (Reingold, 1979). Jackson published in the *American Journal of Science* and scientific journals in the United States and in Europe; he helped found the Association of American Geologists and Naturalists (later the American Association for the Advancement of Science), and served as chairman of that organization for the year 1845-1846 (Woodworth, 1897, p. 86). It is only later that Jackson's status was reduced to a lesser position among American geologists.

C. T. Jackson was born in Plymouth, Massachusetts, on July 21, 1805, the son of Charles Jackson, a merchant, and Lucy Cotton Jackson. Demographically, such beginnings suited Jackson to a career in science. Better than one third of American scientists active from 1800-1863 were New Englanders; a third came from families with commercial backgrounds; and many were trained in medicine, although to have earned a doctorate and to have gained post-doctorate training, as Jackson did, was unusual (Elliott, 1982). There seems to be no evidence that Jackson had an early interest in geology. Instead, his tutors, James Jackson and Walter Channing, prepared him for the study of medicine (Woodworth, 1897, p. 70). With that preparation, Jackson continued his training at Harvard College under John White Webster.

Webster, Erving Professor of Chemistry and Mineralogy during Jackson's tenure as a medical student, was a popular lecturer at Harvard and enjoyed a long teaching career there before being hanged for a sensational murder. He kept abreast of English and Continental journals and spent enormous sums of money on chemical apparatus and mineral samples (Cohen, 1950). Webster also had a keen interest in geology. In 1826 he published "a somewhat detailed account of the geology of Boston and vicinity" and worked on the Roxbury conglomerate, but deferred interpretation of that formation because he found it "inexplicable with the geological information then available" (Merrill, 1904). From Webster, Jackson learned techniques of chemical analysis which he later employed in his state surveys.

Although Jackson seemingly could not help but be influenced by such an individual as John White Webster, Woodworth suggests that Jackson's first interest in mineralogy "was aroused, while staying in Lancaster, Mass., by finding the crystals of macle or chiastolite which there abound in the glacial drift." During the summer of 1827, Jackson traveled to Nova Scotia with Francis Alger where they collected minerals; he returned there for a second field trip with Alger following his graduation from Harvard in 1829. These collecting trips formed the basis of Jackson's first published paper (coauthored by Alger), "A description of the mineralogy and geology of a portion of Nova Scotia" in Silliman's *American Journal of Science* (Jackson and Alger, 1828 & 1829). He also visited New Jersey and New York with Gerard Troost, later the state geologist of Tennessee, and William Maclure, whose Wernerian classification for American rocks Jackson would still be using, with modifications, nearly twenty years later (Woodworth, 1897, p. 71).

Jackson graduated from Harvard in 1829 and traveled to France in the fall of that year, "evidently with the intention of fitting himself for a high place in the profession for which his tutors had prepared him" (Woodworth, 1897, p. 70). In Paris he studied at the University of Paris and attended lectures at the École de Médicine, the Collège de France, and the École des Mines. There he met Jean-Baptiste Élie de Beaumont, with whom he "formed a friendship which lasted many years" (Woodworth, 1897, p. 71). Élie de Beaumont would later write of Jackson, somewhat noncommitally, that he was "bien connu par ses travaux" (Woodworth, 1897).

As it frequently was throughout his life, the timing of Jackson's study in France was unfortunate. Through this period, Élie de Beaumont was primarily concerned with mapping the upper Paleozoic in France (Rudwick, 1985, p. 91); only later would he concern himself with the structure of the Alps and what was known then as "dynamical" geology, a subject that, had he lectured on it, might have been more useful to Jackson in the northern Appalachians (Greene, 1982). Nevertheless, Jackson traveled extensively throughout France and Italy, and saw a good many rocks. Also, for once, his timing improved: he witnessed the 1831 eruption of Vesuvius. But his time in France would have significant consequences later, for it distanced him from the growing English stratigraphic tradition. This was, in part, a matter of choice and of nationalistic prejudice. England was

still an enemy nation for Jackson, and it is difficult to determine how much his patriotism affected his scientific judgment.

Jackson's work in Maine occurred at the same time as the consolidation of the Devonian system by Roderick Murchison and Adam Sedgwick in England. This consolidation was the background for a major scientific debate over which characteristics of a stratum should be used to date it. In his study of that debate, Rudwick outlines some of the points of contention. For many geologists, including the English field geologist Henry de la Beche and the father of the Oxford school of geology, William Buckland, fossils were but one of several indicators of age – just as important to consider were lithology and position in the sequence. Figure 2 shows the sequence used and advocated by Buckland and de la Beche. Murchison and Sedgwick, conversely, advocated a sequence determined primarily by paleontological evidence, following the lead of the English engineer, William Smith.

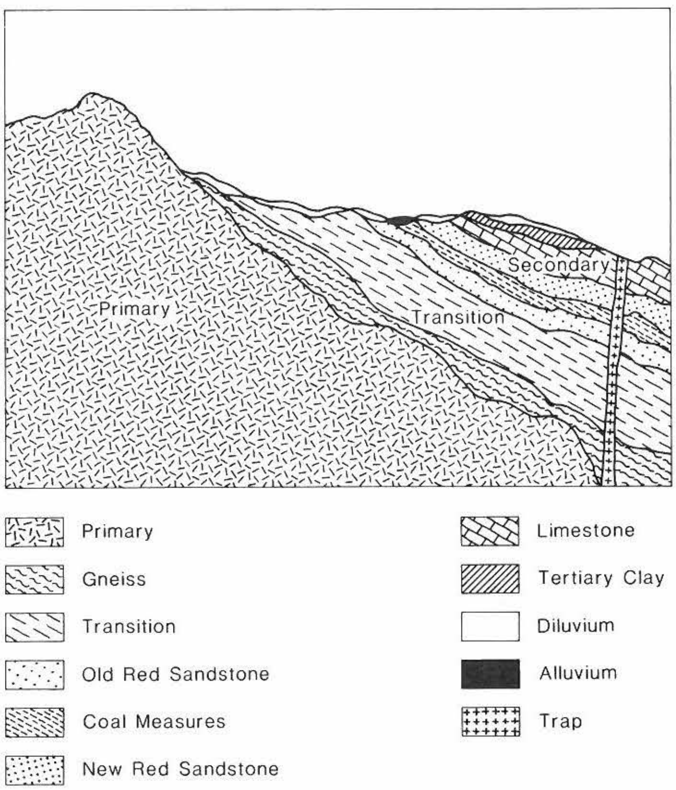

Figure 2. Ideal section as advocated by C. T. Jackson, showing the positions of the Old Red and New Red Sandstones.

The Devonian controversy began as the result of an error in structural correlation by de la Beche, who placed rocks containing fossil land plants at the bottom of a sequence in Devonshire overlain by Murchison's Silurian system. Murchison found this quite impossible; there were no land plants in the Silurian. The battle over the Devonshire fossils raged over a full decade, and was finally resolved by inserting the Devonshire fossils between the Silurian and the Coal Measures, or Carboniferous. Moreover, Louis Agassiz showed that the Old Red Sandstone (see Fig. 2) correlated with the Devonian system. Ultimately, these conventions were applied to the bedrock geology of Maine, but not by Jackson. He continued to reject the English conventions even after 1840, and his rejection of "Cambrian and Silurian as names for our rocks" (Jackson, 1840) did not enhance his stature in the eyes of later historians of geology.

Upon his return to Boston, Jackson set up a chemical laboratory attached to his house at 21 Green Street, married Susan Bridge of Charlestown in 1834, and began to establish a reputation as a physician and chemist. Between 1829 and 1837 he published his paper with Francis Alger on the geology of Nova Scotia; an account of the chiastolite that had inspired his interest in mineralogy; a paper on the conglomerates and dikes of Roxbury; and several chemical analyses of coal, water, and copper. In the commercial directory of Boston, Jackson listed himself as a physician, "but finding his services in demand as a chemist and mineralogist, he gradually and not against his inclination, entered upon a career in these pursuits" (Woodworth, 1897, p. 71). By 1835, with his reputation for field work established by his publication on Nova Scotia, Jackson was hired by private concerns in Maine to examine several possible commercial prospects near Thomaston, Foxcroft, and Williamsburg. When the legislatures of Maine and Massachusetts set aside funds for a survey of the state and its public lands, Jackson was a sensible if not inevitable choice for the position of state geologist.

Although a full discussion of the economic, political, and scientific significance of the early state surveys is beyond the scope of this paper, a glance at the constraints placed upon the early geologists and their surveys will help to round out the context. Hendrickson (1961) has argued that the first state surveys were authorized primarily for economic reasons. Aldrich (1979), however, has shown that the geologists were also interested in contributing to the growth of geologic knowledge, independent of economic concerns. The debates in England and on the continent about stratigraphic conventions were echoed by the state geologists of the 1830's as they searched for uniform meanings for terms such as "formation," "group," and "series" (Aldrich, 1979). And Schneer (1981) has cautioned against searching for systems in the work of the state geologists of this period. The "microgeologic techniques" of Abraham Gottlob Werner, the German geologist whose classification of lithology provided a foundation for the study of rocks, provided necessary tools for field geology in the vast, unmapped, and undescribed wilderness of the North American continent. It was to these microgeologic techniques that the state geologists turned, for the most part; theory and system could come later. Rudwick (1985) rounds out the context when he captures a sense that geology in the 1830's was a "new, exciting, and fashionable science." From the scientific societies of London, Philadelphia, and Boston to the town of Blue Hill, where "no less than forty" townspeople joined Jackson in his ascent of Blue Hill to measure its altitude (Jackson, 1838, p. 38), new ideas about the earth and its history combined with what Rudwick calls "the

romance of fieldwork" to make geology much more than a merely economic activity.

THE PROBLEMS OF THE GEOLOGY OF MAINE

C. T. Jackson spent three full field seasons in Maine, from 1836 to 1838, and published three annual reports based upon that field work. In each of the reports, Jackson introduced the year's work with comments about the nature and value of geology, followed closely by an account in "travelogue style" of apparently "undigested field notes" (Aldrich, 1981, p. 6) in which he described the lithology of each place visited earlier in the year. Here, an attempt is made to recapture the essence of the field work and reports by examining some of the problems Jackson attempted to solve.

In late June of 1836, C. T. Jackson was formally contacted by an emissary from the state of Maine to prepare a geologic survey of the state. Because public lands in Maine continued to be held by Massachusetts, that state contributed to the funding of the survey. The joint commission was received by Jackson early in July. Given the short field season in Maine, he was off to a late start. Only 31 years old, Jackson must have felt daunted by the prospect of surveying an area as large as Maine. He wrote that he "hesitated at first, doubtful whether I should be able to accomplish so Herculean a task and do justice to the subject" (Jackson, 1837, p. 9). Several considerations helped to settle the matter of planning. First, he had been asked by Robert P. Dunlap, the President of the Board of Internal Improvements, to begin his work at the mouth of the St. Croix River, where previous field work led Jackson to suspect a deposit of coal associated with the red sandstone. Second, the survey would take advantage of exposures along the coast. Beyond this starting point, the survey would proceed along a division of Maine into "squares" bounded by the St. Croix and St. John, Penobscot, Kennebec, and Androscoggin Rivers. This division had the advantage of organizing the state into roughly equal areas, and it also took advantage of the rivers for transportation and as likely prospects for outcrops (Jackson, 1837, p. 11). In the first field season, Jackson was assisted by Dr. T. Purrington of Brunswick for the state of Maine, James T. Hodge for Massachusetts, and F. Graeter (draftsman). In the second season, the draftsman was eliminated as an economy, while Mr. W. C. Larrabee replaced Dr. Purrington for Maine. In the third season Jackson was assisted by Dr. S. L. Stephenson – whose report on the headwaters of the Androscoggin River is appended to the third report – and Ariel Wall of Holloway. Figure 3 follows the progress of Jackson's field work through the 1836, 1837, and 1838 field seasons. Altogether, the survey cost the state $12,000 (Merrill, 1920).

Toward the end of his first report (Jackson, 1837, p. 86), Jackson made this statement about his philosophy of science:

> I feel I am attempting to compress the geological history of a great country into too narrow limits. . . . I only hold the pen; Nature dictates the facts, and I have presumed to put in, here and there, a word of interpretation, which I hope may not come amiss.

This statement has led Aldrich (1981) to conclude that Jackson's ". . . primary mission was to describe and record, not to theorize. . . ." Apart from philosophical considerations about whether such a theoretical/descriptive distinction is possible, one might wonder whether Aldrich's characterization of Jackson's "primary mission" is correct. Did Jackson "only hold the pen" and attempt to apply Francis Bacon's scientific method of theory-free induction? To answer this, it is necessary to examine the geological problems faced by Jackson in Maine, and to comb through them for a sense both of the solutions and of the *kinds* of solutions he proposed for them.

Topographic geology

The first of these problems concerned the topography of Maine. Following the lead of Hitchcock in Massachusetts, Jackson made no clear distinction between measuring and describing topography, on the one hand, and noting lithology on the other (Aldrich, 1981, p. 6); herein, each will be dealt with successively.

In 1836, little topographic control had been established for maps of the state, at any scale. Jackson depended on nautical charts, town maps, and the map of the state published by Greenleaf in 1830. From his first field season, Jackson carried a barometer in order to establish altitudes for the mountains he ascended, and would take the bearings of other points of high relief from each summit; but apart from these attentions and occasional corrections to existing maps (Jackson, 1837, p. 10 and 63), he did no systematic mapping. The modern reader may express surprise that Jackson would attempt any kind of survey in the absence of proper maps, or that he seemed to consider mapping outside the demands of the survey. Two observations may serve to explain this deficiency in the plan for the survey. The first is the obvious size and cost of a mapping project, which may have been economically unjustifiable because Jackson's geological survey was funded by the state on a season-to-season basis with no promise of renewal. Accordingly, Jackson sought to provide the greatest amount of information at the least expense of time and money. Thus, the survey years were a reconnaissance in the strictest sense of the term. Moreover, Jackson found his funds reduced in the second year of the survey and argued that the minimal topographic information he wished to provide was compromised by the cut. The result is an unsystematic account of the topography of the state. These measurements are dispersed throughout the reports as lists of bearings and altitudes complemented by verbal descriptions of regions, and the occasional woodcut in which relief is characteristically exaggerated.

The reconnaissance nature of the survey also had an effect on Jackson's discussions of stratigraphy and his descriptions of lithology, but an explanation of Jackson's lithologies and stratig-

Figure 3. Approximate routes followed by Jackson during the three field seasons of his survey.

raphy should not be reduced to economic grounds. The dominant concern of geologists in the 1830's for the stratigraphic sequence has been discussed above; Jackson's handling of this problem, accordingly, will tell something about his competence as a field geologist.

Stratigraphy

Jackson began his 1836 field season at the mouth of the St. Croix River, a significant area because Jackson suspected that it would represent the western extent of the "New Red Sandstone" and underlying Coal Measures of New Brunswick and Nova Scotia, which Jackson had previously seen at first hand. The importance of this aspect of the geological survey was underscored by the boundary dispute with Great Britain over the north and east boundaries of Maine. If the Passamaquoddy Bay area or regions to the north along a line through Calais and Houlton contained coal deposits, or even if they could be correlated with the Coal Measures, the fact would be significant in any settlement of the dispute. "Here," notes Aldrich (1981, p. 7), "was 'mission-oriented' geology indeed."

In the area surrounding Passamaquoddy Bay, Jackson recorded the occurrence of red sandstone extending along the coast from Perry to Robbinston (Fig. 4). Having found charred fossil plants near Pulpit Rock, which he described as "marine," and noting the composition of the sandstone and contiguous lithologies, Jackson confidently correlated it with the New Red Sandstone, stating that "it is . . . an undoubted fact, that the sandstone in question is identical with the red sandstone of Nova Scotia which contains gypsum, salt springs and coal." He believed the sandstone to be an apparent "continuation of that, which exists in New-Brunswick, and in which the bituminous coal of Grand Lake is probably contained" (Jackson, 1837, p. 17). According to the stratigraphic conventions followed by Jackson, the New Red Sandstone was a top bed marker of coal-bearing strata; thus, "No geological observations would imply, that the red sandstone in question should not contain coal, for if it should be found equivalent to the new red sandstone formation of Europe, it will belong to the upper coal series" (Jackson, 1837, p. 18).

Jackson's correlation of the Passamaquoddy sandstones provides a useful case study for examining his commitment to a stratigraphic methodology. Before exploring the minutiae of the case, it is important to point out that Jackson often reached for the gross correlation: a case in point is his promise, for a "future excursion" to "trace the known coal-bearing strata of New Brunswick, up the St. John, from the Grand Lake coal mines to the Aroostic; and thence, if the strata are found to be continuous, following their course until they intersect the public lands" (Jackson, 1837, p. 69).

Jackson's tendency to reach for a gross correlation does not entirely explain why he placed the sandstones in Perry with the New Red Sandstone. Perhaps he knew nothing of the Old Red Sandstone. Alternatively, he may have discounted or ignored the possibility of its occurrence in America. The former is unlikely; although there is no evidence that he was familiar at this time with Henry de la Beche's textbook (1831), and it is improbable that he could have seen Buckland's "Bridgewater Treatise" (1836) at this early date, he certainly had read both by the time of the third report, which contains a reaffirmation of the correlation. Both Buckland and de la Beche showed the Old Red Sandstone below the Coal Measures, and offered Jackson an alternate interpretation. But he seems never to have entertained the alternative. The latter explanation is more likely. The Old Red Sandstone, an important stratum in England and parts of Europe, does not occur among the rocks he saw in Europe, or among those in Nova Scotia where he cut his geological teeth. What should lead him to suspect that the rocks of Passamaquoddy Bay should differ from those of Nova Scotia or New Brunswick?

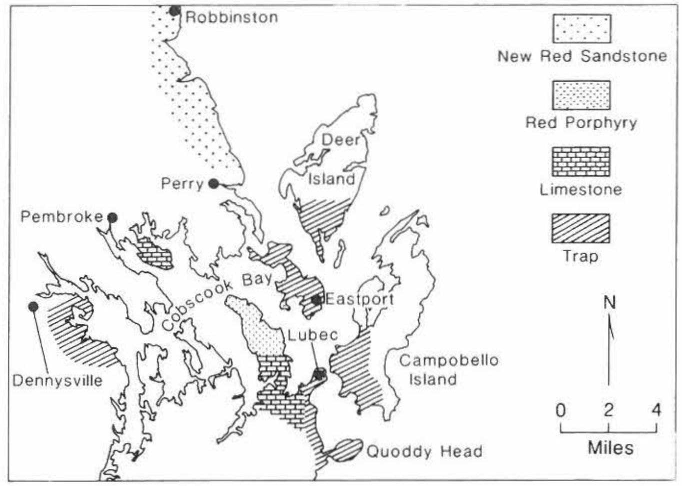

Figure 4. The "New Red Sandstone" of Perry.

The answer is to be found in the two criteria for correlation he mentions: fossils and superposition of strata. Inasmuch as the 1836 field season took place at the height of the Devonian debate, it is inconceivable that fossil evidence could have settled the matter, even if Jackson had known the details of the debate and had taken a position in it, which he did not, prior to 1840. The "charred fossil plants" at Perry he identified as *Fuci*; he also found the tests of "Natica (socialis?)" in limestone at Machias. These, Jackson designates as "fossils of the secondary series" (Jackson, 1839, Catalogue, p. xvii). Again, however, the Devonian system was as yet inchoate and could not have been used in 1836 to decide the age of the Perry strata. One is left with superposition and Jackson's promise to trace the New Red Sandstone through New Brunswick to the rocks west of the St. John River. This, however, may have been impossible given the international boundary dispute; a proper traverse from Grand Lake through to the St. John River by a state-commissioned geologist, however innocuous, would have been construed by the British as impolitic at best and as an act of war at worst. And, of course, such a traverse would not have been innocuous, given the "mission-orientation" of the research.

Thus, the "New Red Sandstone" correlation was the best pos-

sible interpretation for the Perry strata at the time and given the circumstances. Jackson was cautious, moreover, to point out that "it not unfrequently happens that some members of the coal series are wanting, which may be the case here. It is however worthy of exploration; and by boring through this rock in a few places, the question may be settled at little expense, to those who may enter on the task" (Jackson, 1837, p. 18).

Subsequent work has established that Jackson made an error in his correlation, and that the sandstone of the Perry Formation is the Old Red Sandstone. But it is important to understand that Jackson's error was a *correlation* error, not necessarily an error of stratigraphic methodology, given the shifts of methodology taking place at the time. Indeed, the error may appear methodological only in retrospect.

To understand this, Jackson's stratigraphy must be examined in the context of the 1830's. It is clear from the texts of his reports, from the definitions in the glossary printed with the first report, and from the following comment in the introduction to Jackson's report on the geology of Rhode Island (Jackson, 1840), that Jackson's formations followed the "Wernerian" style:

> A numerical division, will doubtless be found preferable to any of those fanciful names, which have lately been proposed for certain groups of strata of the Transition series, and it is evident that the names Cambrian and Silurian, proposed for certain groups in England, will never be regarded in this country as appropriate terms for our rocks; and I observe that they have not been adopted by De la Beche, in his late Report upon the Geology of Cornwall, Devon, and West Somerset. (Jackson, 1840, p. 11)

This notorious comment, often quoted (Woodworth, 1897, p. 74; Merrill, 1904, p. 347-348), may be interpreted along several lines. Was it a reaction to the parochialism of using regional names to refer to a global sequence? If so, Jackson was in good company, concurring with the Rogers brothers of the Pennsylvania and New Jersey surveys (Aldrich, 1979, p. 136). Or was it a statement of commitment to Wernerian categories? This is the position taken by Merrill, who wrote that Jackson "was conservative almost to the point of obstinacy, as is shown by his steady adherence to the older forms of classification, though finding it necessary to depart somewhat from the ideas of Werner" (Merrill, 1904, p. 290). Woodworth, referring to the preface of the Rhode Island report, suggests that Jackson "was not an advocate of biological methods in geology," and that "his predilection for chemistry and mineralogy manifestly made geology for him a mineralogical rather than a stratigraphical science, and the peculiarly crystalline character of the rocks of New England fostered this view of geology" (Woodworth, 1897, p. 73). While Woodworth's points are well taken, they ignore the possibility that Jackson's "predilection" represented an *evolution* in commitment rather than mere obstinacy, as Merrill would have it. In his first annual report, Jackson discussed "fossils, which are very important, in determining the relative age of the rocks, in which they are found" (Jackson, 1837, p. 27). Wherever he found them in Maine, Jackson commented on the fossils. But the fossils Jackson found in Maine occurred predominantly in the "tertiary" clay formations and in erratic boulders. While the former were useful in establishing the age of the clays, the latter were useless unless the source of the erratics could be traced. Given the status of stratigraphy in the period and the relative absence of fossils in the rocks of Maine, a numerical division of sequence, established through lithology, superposition, and structures, was the more useful – and often the only – correlation tool. Even in 1987 fewer than one fourth of the formations and their members shown on the Bedrock Geologic Map of Maine have paleontological controls; of those, one fifth are not correlated to a single system (Osberg et al., 1985). If Jackson's predilections made geology a mineralogical rather than a biostratigraphic science, a good number of geologists have followed him in that tradition. If this is the case, it may be reasonable to conclude that Jackson made an attempt to forestall a set of conventions which had only tenuous application to vast expanses of rock. Jackson's notorious comment about the Silurian and Cambrian systems is followed in context by his cautious plea that "a new nomenclature would be wholly irrelevant while Geology is in its present imperfect state, and it is highly desirable for us to maintain the old landmarks, until new ones can be established by general agreement" (Jackson, 1840, p. 12).

Bedrock geology

In each of his annual reports, Jackson promised to provide a geological map of Maine as part of a final report. Although the legislatures of Maine and Massachusetts did not require one, Jackson seems to have considered a map a prerequisite of a satisfactory survey. The following remark from the second report contributes to the sense that Jackson hoped to prepare some systematic summary of the data he collected in the state:

> How is a geological survey to be conducted? This question may be answered as follows: The district in question is first to be examined, so as to ascertain the order of strata, and the relative age of each stratum, while, at the same time, the intersecting rocks are to be observed. The method pursued is first to form a plan of operations, so that all the observations may be recorded, in an orderly manner, that no confusion may arise in the completion of the work. (Jackson, 1838, p. ix)

A non-ironic reading of this paragraph requires a large measure of charity; the confusion of the published reports may be explained away by arguing that the cessation of funding precluded "completion of the work." But confusion is the dominant feature of the reports. As noted above, Jackson's stratigraphic methods gave priority to the superposition and composition of the rocks. With his assistants, he

> described all the rocks exactly as we saw them, and the annual reports must be regarded as the mere field notes that may serve for a more thoroughly rational system, illuminated by a comparison of the results with each other. . . . (Jackson, 1839, p. ix)

It is difficult, however, to read any "rational system" into the reports because Jackson provides little sense of scale in his "field notes." Where he describes intercalated limestone and argillaceous slate, the reader usually gets no indication of the relative thicknesses of the beds, and few clues to the dominant lithology. This confusion is systematic, and Jackson's comments about how a survey "is to be conducted" may have been disingenuous. As Aldrich (1979) has found in her study of the state geological surveys, "successful surveys courted the voters partly by discoveries in economic geology. . . ." By his own admission, Jackson's primary interest was to find economic benefit from the geology he described (Jackson, 1839, p. 1-2). Jackson's admitted preference for discussions of fine-scale lithologic features and occurrences of minerals – bog iron deposits, attention to minor beds of limestone, a mention of the use of chlorite by Indians for making pipes – obscures the picture of regional geology. The state legislature, in bringing the survey to a close in 1838, may well have considered the former sufficient, the latter unneeded.

Jackson did, however, provide lithologies aplenty in his reports; from these, a blurred mental map of the bedrock geology of Maine emerges. But if Jackson's reports are "mere field notes," is it possible to construct a true geological map from them? Such a suggestion is tempting, and several clues to the appearance of such a map – Jackson's lithologic descriptions; his tendency toward gross correlation combined with his intuition that the "general direction of strata in Maine is N.E. to S.W." (Jackson, 1837, p. 11-12) and measurements of beds confirming the intuition; the handful of sections that appear in the reports; Jackson's comment that the promised map would be shaded; etc. – make that temptation irresistible. Accordingly, Figure 5 shows a geological map based upon Jackson's field notes. It is important, however, to understand that this is *not* in any sense the map that Jackson himself would have prepared, for a number of reasons.* First, despite his written intentions of preparing a shaded or tinted map, the maps of Rhode Island and New Hampshire (see Aldrich, 1981, p. 7 and 9) show a numerical symbolization of bedrock on a town-by-town basis. Second, few contacts are indicated in the reports; even if they were, they would have been of minimal value in the absence of a properly contoured base map. Also, Jackson had no clear understanding of the significance either of large or small-scale folds, although how this may have affected his decisions about mapping is a matter for pure conjecture. Finally, modern knowledge, based upon the 1985 Bedrock Geologic Map of Maine (Osberg et al., 1985), could hardly be banished from the cartographer's bias. Accordingly, the accompanying map is best read as a diagram or summary of Jackson's lithologies of rock units organized around a presumption of Jackson's presentation of stratigraphy.

*As David Gooding (1986) has recently pointed out, however, the reconstruction of experiments conducted by historical figures in science is fraught with complications. Thus the reconstruction of an experiment – or, in the present study, a geological map – that was never attempted by a former scientist probably should not be attempted by the historian.

Woodworth, in his memorial to Jackson, provided the following insight:

> Dr. Jackson did not always push his theories of geological phenomena to the fullness of conclusion and statement which would enable us at the present day fully to understand them. He had too many irons in the fire to do as he would with all of them. (Woodworth, 1897, p. 83)

Nowhere is this better seen than in the unkept promise of a geological map of Maine.

Diluvialism and Geomorphology

As it must have been for Merrill, working at the turn of the twentieth century, the most striking aspect for the modern reader of Jackson's reports is the recurrence of references to the "mighty rush of waters" (Jackson, 1837, p. 65) that carved "diluvial scratches" in every part of Maine. According to Merrill,

> Jackson's views on the glacial deposits were naturally crude. The "horsebacks" (ridges of glacial gravel) were regarded as diluvial material transported by a mighty current of water. (Merrill, 1904, p. 347)

Apart from a reference to Jackson's "criterion for distinguishing ice-borne from water-transported detritus," Woodworth the apologist never discusses Jackson the diluvialist (Woodworth, 1897, p. 83). With a greater stock of hindsight, Aldrich states that Jackson

> adhered to the theory of a catastrophic deluge, patterned closely on the Biblical flood, to account for these phenomena. (Aldrich, 1981, p. 6)

Although it is true that Jackson was a diluvialist, it is not at all clear that the deluge as a mechanism for transporting the diluvial materials was "patterned closely on the Biblical flood" or that a reading of "catastrophic" in the sense in which Whewell, the nineteenth-century English geologist and philosopher, used the term precisely describes Jackson's views on the matter. There is at least some evidence that Jackson postulated more than one period of global flooding, as when he wrote of "the *last* grand deluge that overwhelmed the globe" (Jackson, 1837, p. 74-75, emphasis added). Such a suggestion of cyclical flooding is not true to the Biblical account; it is rather an echo of one of several theories held by Buckland (Rupke, 1983). Moreover, the words "cataclysm" and "catastrophy" must be approached with care when they appear in the reports of geologists prior to 1840, the year in which Whewell framed the catastrophist-uniformitarianist debate, making these terms to some extent taboos in geology. When these roadblocks to historical inquiry have been cleared away, what emerges is remarkable growth in Jackson's commitment to the diluvial theory in the years 1836-1839, as well as in his ad hoc explanations for anomalies to the theory.

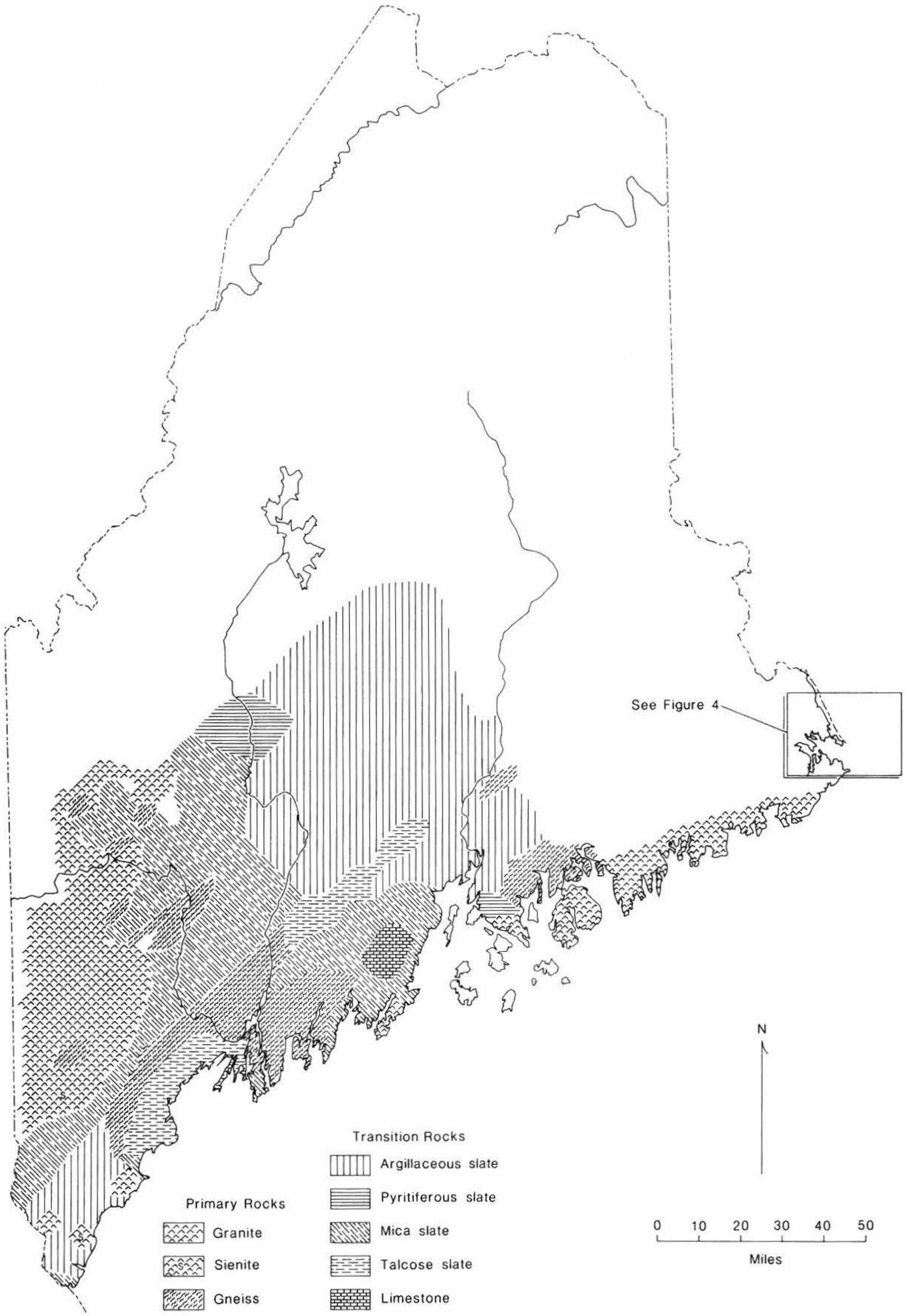

Figure 5. A reconstruction of Jackson's field notes on lithologies in Maine, in map form. Contacts are not indicated. As noted in the text, this figure is a summary and should not be considered the map Jackson would have drawn.

As Aldrich (1981) notes, Jackson's "preoccupation with recording facts led him to report the compass bearing of virtually every scratch on the rocks." The preoccupation begins early in the 1836 field season and continues unabated through the third year of the survey. This apparently inductive activity was accompanied by a strengthening belief that most of the unconsolidated materials found in Maine, as well as the scratches, were caused by a current of water. No observation of an isolated scratch stood alone; each was used to reinforce Jackson's position that

. . . there is a striking coincidence between the direction of these marks and the diluvial grooves which I have noticed before. Those in Portland run from N. 15° W. to S. 15° or 20° E. and here [in Charleston], in a distant portion of the state, nearly the same direction is observed. We have, however, many more equally good illustrations of this subject. (Jackson, 1838, p. 28)

Ideally, Jackson might have mapped the scratches, but the verbal descriptions of the diluvial phenomena are nearly as graphic as a map. More important, the recording of diluvial scratches allowed Jackson to determine the source directions of erratics and of the parent materials for soils:

It will be readily conceived, that if solid rocks were moved from their native beds, and carried forward several miles, that the finer particles of soil should have been transported to a still greater distance, so we find that the whole mass of loose materials on the surface has been removed southwardly, and the soil resting upon the surface of the rocks, in place, is rarely, if ever, such as results from the decomposition of those rocks, but was evidently derived from those ledges which occur to the Northward. (Jackson, 1838, p. 149)

Thus, despite the wealth of lime-rich rocks in Thomaston, Jackson finds the soil calcium-poor,

. . . derived from the decomposition and disintegration of granite, gneiss, and mica rocks which lie to the northward of that town.

This fact accounts for the almost entire absence of carbonate of lime in the soil, and indicates at once to the farmer, that liming is there extensively required. (Jackson, 1839, p. 62)

It would be convenient to interpret Jackson's diluvialism as *heuristic*, that is, as a unifying concept which furthered his investigations but which was held tentatively or understood to have no basis in nature. It would be a mistake, however, to treat Jackson's commitment to diluvialism as a mere heuristic; for better or for worse, the commitment was extensive. Controverting the Hutton-Playfair account of valley excavation, in which streams are understood to have carved the valleys through which they flow, Jackson claimed that the geomorphology of the Kennebec valley area in and around Augusta could be explained through the deluge hypothesis: ". . . anyone who looks upon the general direction of these valleys, will feel satisfied, that they were excavated by a current of water" (Jackson, 1837, p. 84). Moreover, Jackson developed ad hoc explanations for diluvial data where these countered the overall trend. In Phillips, Jackson encountered "several remarkable phenomena":

First, the occurrence of diluvial markings, which do not coincide with the direction formerly noted, as the general bearing. Secondly, the occurrence of extremely heavy masses of iron ore of foreign origin, and granite rocks also erratic, poised upon the summit of an insulated hill. The questions that naturally arise are, first – how came these scratches on the surface of the ledge? And secondly – why, if they owe their origin to causes I have formerly assigned, do they vary in their course? (Jackson, 1839, p. 28)

The anomaly, though puzzling, did not lead Jackson to question the diluvial explanation. Instead, he adjusted to the data, using the surrounding mountainous topography to provide a distorted sluiceway through which the deluge behaved with corresponding turbulence:

. . . this apparent anomaly in the direction of the diluvial scratches, is a most striking and wonderful confirmation of the theory which we have enunciated; because the shape of the country, as is evident to any observer, would have caused the precise deflection observed in this case; for Mt. Abraham arrested the current on the north and turned it into Sandy River valley on the west, from which deflection it struck against the Mt. Saddleback range, continued to Mt. Blue, and by Saddleback was reflected, precisely according to the well known laws of physics, towards French's Mountain; and thus the marks coincide with the direction of the two forces. It moreover proves incontestably that the current did not set in from the S.E., for the course would have been at right angles with the present markings. (Jackson, 1839, p. 29)

Here, a map would have been helpful, for Jackson's description is insufficiently graphic to provide a reading of the phenomenon. The problem, alas, is not reconstructible in modern terms. Yet it is an ad hoc explanation to which Jackson became committed in explaining additional anomalous striae (Jackson, 1839, p. 32, 43).

As noted by Aldrich (1981), Jackson observed "the power of ice in moving boulders during the spring thaw on New England rivers, but he did not use the mechanism, in the form of glaciers, to explain the rock gouges, transport, scouring, or moraines which decorate the state's landscape." Additionally, Jackson seems to have rejected Lyell's proposal for transport by icebergs as well as the glacial theory (Merrill, 1904, p. 348), the latter on the grounds that the striations should show a radial distribution in mountainous areas, which they generally do not.

Such an objection seems to be out of proportion to obvious objections to the diluvial theory. Buckland, beginning with his recantation of diluvialism in the Bridgewater Treatise, continued to search for an explanation for the diluvial phenomena and accepted Agassiz's explanation well in advance of Lyell. As Rupke has argued,

Intellectually the change from a diluvial to a glacial mechanism of boulder emplacement was very small indeed; in the place of [Sir James] Hall's tidal wave came a huge mass of frozen water. (Rupke, 1983, p. 106)

Jackson, however, remained adamantly opposed to the glacial theory, apparently throughout his life. Nevertheless, he made a contribution to understanding the surficial geology of Maine. In part, this is because the difference between the glacial theory and the discredited diluvial theory has been established through inferences from terrains that are currently glaciated. No *direct* empirical evidence exists to decide whether a flood or a sheet of ice existed here at a particular point in time. The theories are, however, clearly incompatible vis-à-vis some aspects of flow mechanisms and drift transport. But in terms of heuristic power, they are virtually equivalent systems when applied to the workaday problems of the state geologist in the 1830's. Greene shows that this distinction is not by any means trivial. Quoting an account by Darwin of a field trip in Wales with Adam Sedgwick, wherein abundant glacial phenomena were completely overlooked by both geologists, Greene states that

> Darwin did observe glacial phenomena in the colloquial sense, but he did not see them as elements joined together by a theory and therefore did not remark upon them. What he lacked to understand them on that first visit was not powers of observation but some concept that would have extracted an organized body of fact from a jumble of stones. (Greene, 1982, p. 59-60)

Jackson lacked no such system. Moreover, in the collection of data, the term "diluvial scratch" may be transposed with "glacial striae" with virtually no loss of meaning, as Aldrich points out (1981, p. 6). And Jackson's contribution to an understanding of the origin of Maine's soils changes not a whit. Characteristically, Jackson did not prepare a map from his diluvial data. In Figure 6 such a map has been constructed from the reports. The difference between the reconstructed map and the present surficial map (Thompson and Borns, 1985) – apart from terminology – is quantitative, not qualitative.

Beyond the question of whether Jackson's interpretation of the striae, gravels and erratics was "right" or "wrong," there emerges another theme: the clear sense that the diluvial geology in Jackson's reports is, in fact, interpretation. Although the diluvial interpretation stands opposed to Jackson's apparently inductive methodology, giving lie to his claim that "nature dictates . . .," Jackson's descriptions of diluvial phenomena go far beyond the purely observational.

In addition to recording and interpreting the diluvium, Jackson also commented upon the appearance of clays throughout the state's coastal plain. At Lubec, in 1836, he found recent "marine shells . . .in regular layers imbedded in the clay" in an excavation for a tidal power canal. Jackson related this phenomenon to the erosion of greenstone trap at some distance from the ocean and asks "Has the level of the sea become depressed or have the rocks been elevated?" Lacking evidence for a regression of sea level, Jackson concluded that the land had emerged "within the recent Zoological period" (Jackson, 1837, p. 19). As was the case for the diluvial interpretation, Jackson's recognition of emergence provided a coherent explanation for the clay deposits in the state. In the second report, he placed the limit of sea level transgression at 100 feet, but increased that limit to 150 feet in the third season.

To call Jackson a "catastrophist" would not be correct. But he clearly saw the geomorphology of Maine as of two types: that of some prior time and that of the contemporary topography. Jackson was no devotee of Lyell, as Aldrich notes (1981, p. 6). A final example of his investigations of unconsolidated materials, however, may serve to provide a perspective on his theoretical commitments. While conducting a reconnaissance through Limerick, Jackson was drawn to a peat bog that had recently been drained, and in which one Ebenezer Adams claimed to have found coal "amid the remains of rotten logs and beaver sticks" (Jackson, 1838, p. 80-81). Jackson did a chemical analysis of the finding, and pronounced it "a true bituminous coal." If it occurred to Jackson that Adams was simply having some fun at the expense of a government geologist, he did not say so. Instead, Jackson wrote that

> The discovery of the recent formation of bituminous coal cuts the gordian knot which geologists and chemists are endeavoring to unravel, and shows that the process is still going on. (Jackson, 1838, p. 81)

This was not the sort of comment a "catastrophist" might be expected to make. As Greene has argued, the catastrophist/uniformitarianist debate was the invention of Charles Lyell – a rhetorical device used by the trained barrister to make a case (Greene, 1982, p. 25-26). Jackson did not systemize his findings; instead, he employed *and expanded upon* a series of heuristics to interpret the landscape of Maine. So long as none of the interpretations contradicted any other, such an approach was both necessary and sufficient.

Economic and agricultural geology

Jackson's overwhelming concern in the survey of Maine was to find economic value in the rocks. Accordingly, much space in the reports was given over to discussions of the granite and limestone quarries in the state, and to determining the economic values of bog irons, peat bogs, and minor veins and deposits of ores throughout the state. Aldrich (1981) has discussed Jackson's contribution to the economic and agricultural geology of Maine, and little more needs to be said. It is noteworthy, however, that Jackson recognized the natural beauty of Maine and recommended encouragement of a tourist industry in places like Moosehead, Blue Hill, and Denmark.

Although he suspected that coal might be found in the eastern part of the state, Jackson was unequivocal about coal speculations elsewhere in Maine. At Small Point, Jackson inspected some coal that had washed up on the shore. This, by chemical analysis, was identical with Orrel coal from England, and Jackson pointed out that

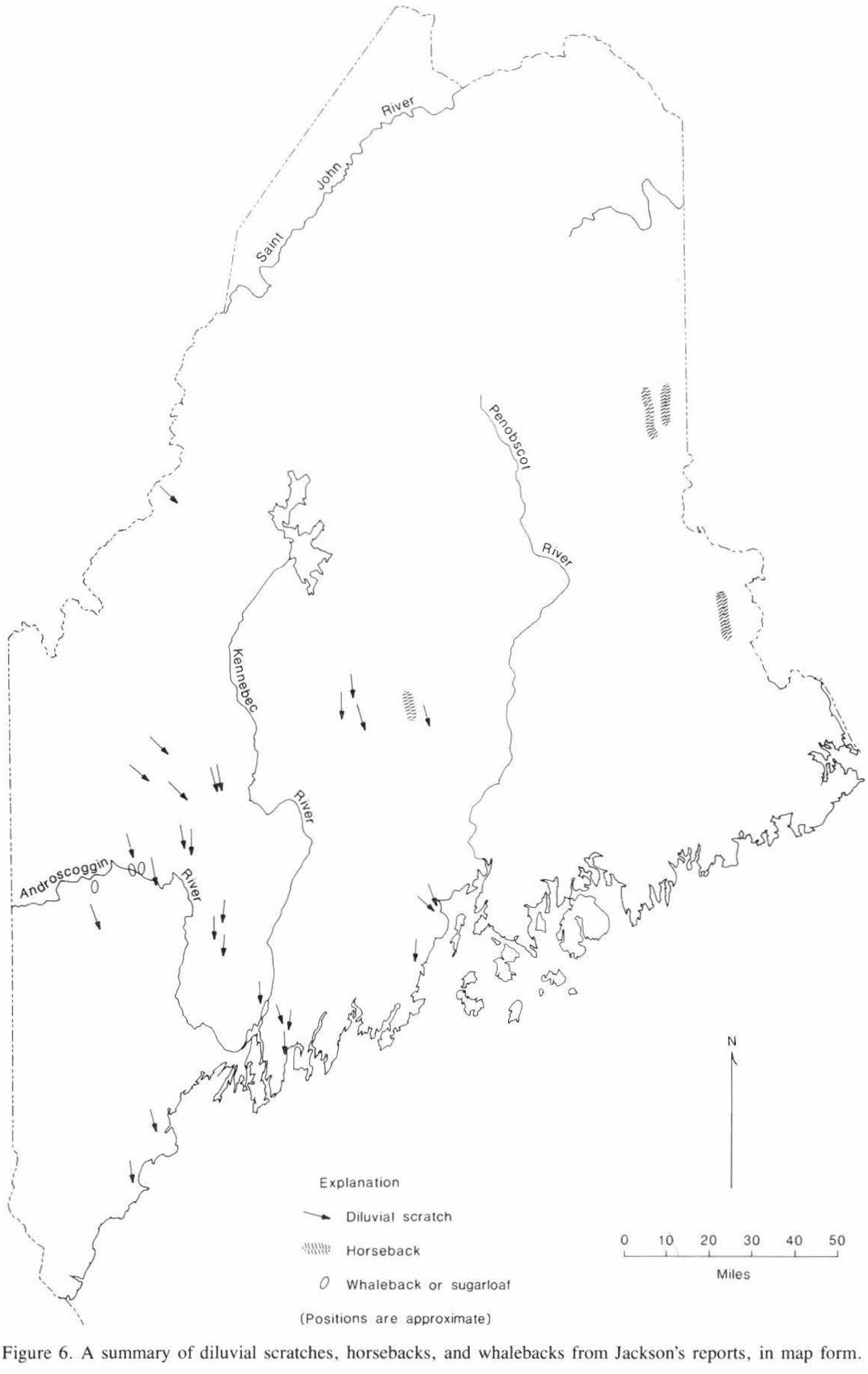

Figure 6. A summary of diluvial scratches, horsebacks, and whalebacks from Jackson's reports, in map form.

> . . . the rocks along this coast were . . . gneiss, a primary rock in which coal is never found, and the beach consists of silicious sand . . . evidently derived from the disintegration of similar rocks. (Jackson, 1837, p. 82)

Often, he would hear of a suspected coal-bearing rock but, on examining the rock, would find tourmaline, black oxide of manganese, or graphite. Jackson debunked these speculations, usually by pointing out simply that primary, or crystalline rock, could contain no coal.

> We are never to look for that combustible lower down in the series than the newer transition, nor above the secondary. Hence the absurdity of searching in granite and mica slate rocks, for beds of coal, and the mistakes arising from the occurrence of lignite in the tertiary clay – both common and fatal errors to those who engage in such absurd enterprises. (Jackson, 1839, p. xiii)

Rocks with a high sulfur content near Castine were thought by many to be a coal indicator:

> It was . . . originally imagined by the English, during the late war, that a coal mine existed in this spot, for as coal frequently contains sulfur, they thought it probable that a rock containing sulfur must necessarily contain coal. Several other persons have since been deceived in a similar manner, and within a few years borings were made for coal. The auger penetrated to the depth of 100 feet, and brought up nothing but pyritiferous slate, as might have been anticipated. . . . Now had this locality been a coal formation, as it certainly is not, there would have been no need of boring, for the strata stand upon their edges, or at an angle of 70° with the horizon, and no person, at all acquainted with the structure of the earth, would ever think of such an operation, for it would not give any information of the kind required. A geological observer can penetrate a thousand feet deep, when such is the position of the rocks, without digging into them at all. It is an open book that is laid before him, and he has only to observe attentively. (Jackson, 1838, p. 47)

This last comment returns us to Jackson's philosophy of science. In his introduction to the second report, Jackson wrote:

> Geology is a science composed almost entirely of facts, and the theories serving to explain them, are but the *rationale* of those facts. Such, at least, is the modern aspect of the science, and the more rigid are we in our deductions, the more imperishable will be the results. Hypotheses may be exploded, theories are subject to continual modifications, according to the light that may be shed upon their subject, but FACTS are in their nature immortal. (Jackson, 1838, p. viii)

Jackson's fact/theory distinction, while not as rigidly held as that of Henry de la Beche (Rudwick, 1985, p. 452-453), was sufficient to blind later readers to the extensiveness of Jackson's theoretical interpretation. Several examples – the diluvial anomaly and the bitumenization of peat – have been cited as examples of this interpretive bent on Jackson's part. For late nineteenth century geologists and for modern readers, these interpretations were and are considered wrong, of course, but they are interpretations nonetheless. Just as important is the reciprocal problem in the fact/theory distinction: the presence of folded strata in Maine is a fact, but one that Jackson did not recognize because he held no comprehensive theory of dynamic processes to organize the data-collecting process here. Perhaps, it may be argued, he ought to have devised such a theory.

EPILOGUE

Jackson's reports were published soon after each field season and totaled, including two reports on the public lands, nearly 850 pages. The citizens of Maine undoubtedly considered this much information on rocks sufficient for reasonable purposes, and ceased funding in spite of Jackson's lobbying efforts in print and in person, an effort that was also taken up in Benjamin Silliman's journal. One major review and two short reviews of the reports on Maine published in the *American Journal of Science*, and almost certainly written by Silliman (1839) himself, call Jackson "able and perspicuous" and "one so thoroughly qualified by study and observation. . . ." The lobbying efforts were, however, to no avail and the first geological survey of Maine ended following publication of the third annual report. No further state-financed work would be carried out in the state until the Hitchcock survey of the 1860's.

Jackson, his reputation bolstered by his work and publications in Maine, went almost immediately to work in Rhode Island and published a single (and final) report for that state in 1840. From there, Jackson moved on to New Hampshire; this survey required four years (Aldrich, 1981, p. 8) but, having learned a lesson in Maine, Jackson withheld his findings for the publication of a "final report" in 1844. Geologic maps accompany both the Rhode Island and New Hampshire reports. Moreover, Jackson made an attempt to interpret the overall geology of New England in his New Hampshire volume (Aldrich, 1981, p. 8-9).

Throughout this period in his life, Jackson belonged to and helped to found several professional societies. Through them and through letters, he sought to increase public education in geology, and seriously proposed that every state survey provide for as many as fifteen duplicate collections of samples to be shipped off to colleges, in addition to a collection for the state government. Merrill records the sad fate of the 1,566 specimens of Jackson's "state cabinet" in Augusta which "were thrown promiscuously into boxes and otherwise in disarray"; most of these were transferred to Colby College in 1888 (Merrill, 1920, p. 132-133).

Following the New Hampshire survey, Jackson enjoyed a three-year hiatus from government geological work. In 1847 and 1848 he worked as a United States geologist in Michigan, but resigned for reasons that remain obscure (Woodworth, 1897, p. 79-81; Merrill, 1904, p. 414-415). With the Michigan survey, Jackson's career as a government geologist came to an end.

Almost simultaneously, a storm of priority disputes began over the discovery of the anesthetic value of ether and the invention of electrical telegraphy; these cannot be dealt with here except to say that the disputes became the focus of Jackson's

later life. Accounts of the priority dispute in Woodworth (1897) and Bouvé (1880) give credence to Jackson's claim, but for the modern reader, Jackson appears as a nineteenth-century Robert Hooke, claiming priority for discoveries about which he had published little or nothing.

By 1873, Jackson was overcome by an apparent mental illness which prevented any further work. He died in 1880 at Massachusetts General Hospital. Thomas Bouvé, writing the memorial remarks for the Boston Society of Natural History, said

> The truth is, Dr. Jackson was a man of great genius, and his intuitive perception of scientific truths remarkable; but from some peculiarities hard to comprehend, he often contented himself with enunciating what he recognized as a fact, without striving to substantiate it. (Bouvé, 1880, p. 46)

Woodworth, in a more expansive tone aided by distance in time, had this to say:

> Jackson was a genius. He had the inventive faculty; the habit of incessant investigation; the capacity of getting tangible, fruitful results; and the ability to suggest successful expedients to others. Geologists think of him as a geologist. (Woodworth, 1897, p. 85)

CONCLUSION

C. T. Jackson is not remembered today as a genius. Beginning with Merrill (1904) and continuing through Aldrich (1981), the historical assessment of Jackson's work has ranged from negative to neutral, at best. The present revision in the historical judgment of Jackson might appear at first glance to be an apology. But the apologist need argue only that Jackson fulfilled the political mandate for a "state survey" with notable efficiency, and with an attention to political realities which was uncharacteristic of the early state surveys viewed as a whole. Indeed, the apologist could argue that the meaning of "geological survey" has changed since the 1830's in a way that renders Jackson's work incommensurable with later "surveys." This is not, however, an apology. For to the question: why has Jackson been "assigned a lesser place in the pantheon of earlier investigators of the Northeast"? (Aldrich, 1981, p. 10), there is a reciprocal question: why be concerned about Jackson at all? The latter question is more easily answered than the first. The answer has to do with the revolution in the history of science proposed by Kuhn in *The Structure of Scientific Revolutions* (1970), a revolution that has yet to be carried through. Kuhn's book has been interpreted to suggest that the examination of the scientific process would benefit, indeed may not be possible without, an examination of "normal science." Normal science is, as Rudwick (1985) has put it, "the ordinary business of scientific research [which] is carried on within a shared or collective framework of methodological assumptions, heuristic maxims, routine procedures, observational and experimental standards, criteria of interpretative judgement, and much else besides." It encompasses the kind of discovery that results from empirical field work and experimentation. Simply put, knowing where to look, when to look, and for what to look leads to the kind of discovery that is characteristic of the field geologist in normal scientific practice. Investigating "normal science" as part of the Kuhnian research project has hardly begun.

Jackson, for the period of the state surveys, was doing normal science. The foregoing details of his survey of Maine are but a vignette of the normal science practiced by American geologists in the 1830's. As Daniels has written,

> Whatever the fate of natural-history theories, natural history descriptions are not so likely to be superceded as they are to be elaborated and refined. The history of scientific progress, therefore, has a place for those who wrote the early descriptions. . . . An over-zealous attention to scientific progress has obscured the entire nature of the early nineteenth century scientific community. (Daniels, 1968, p. 32)

Moreover, the various observational practices, assumptions, maxims, procedures, and the like applied by Jackson to the survey of Maine were dictated by a felicitous match of his so-called "Wernerian" commitments and mineralogical biases to the problems of reconnoitering the geology of Maine. In contrast, the Lyellian metatheory of the 1830's had little application to the geology of Maine, and a first survey by an accomplished biostratigrapher might well have produced far less of importance than Jackson's, had it been possible at all.

That Jackson was not a Lyell, a Hutton, a Murchison, or even a Werner is obvious enough. But his "lesser place in the pantheon" has not been assigned relative to these figures in the history of geology. Rather, it is relative to James Hall and the Rogers brothers who developed metatheories to explain what they found in New York and Ohio, Pennsylvania, and New Jersey. But the geology of Maine is daunting, both in terms of complexity and of accessibility, as any field geologist who has worked here will attest. The development and application of metatheories must be made with caution. The structure of the geology of Maine is exceedingly complicated, and correlation is difficult and as yet incomplete. As a type area for the phenomenon of post-glacial coastal submergence, Maine is unique in the United States. Accordingly, a survey-for-survey comparison of the early state geologists is misleading, for these geologists did not work on the same problems or under comparable constraints.

Jackson was a competent field geologist, accomplished at fine-scale interpretation, and politically attuned if not astute. But in terms of modern geology, he was "wrong" much of the time. If the ultimate goal of the history of geology is an object lesson – and who is to say that it is not? – there is object lesson enough in the story of C. T. Jackson.

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Lithotectonic Stratigraphy, Deformation, Plutonism, and Metamorphism, Greater Casco Bay Region, Southwestern Maine

Arthur M. Hussey II
Department of Geology
Bowdoin College
Brunswick, Maine 04008

ABSTRACT

The stratified rocks of the greater Casco Bay area include the Casco Bay Group, Cross River formation, and rocks previously mapped as the Bucksport Formation, all of probable Precambrian to Ordovician age. The Bucksport-type rocks are now correlated with the Sebascodegan formation, originally the Sebascodegan member of the Cushing Formation of the Casco Bay Group. Bounding sequences are the Late Ordovician-Devonian Kearsarge-central Maine sequence, Ordovician Penobscot and Benner Hill sequences, and Precambrian to Ordovician Merrimack Group.

The Casco Bay Group is interpreted to represent two distinct lithotectonic packages. One lies on the west side of the Flying Point fault and consists of the Nehumkeag Pond, Mount Ararat, Torrey Hill, and Richmond Corner members of the Cushing Formation. The other, east of the Flying Point fault, includes the remaining members of the Cushing Formation (Peaks Island, Wilson Cove, Merepoint, Bethel Point, and Yarmouth Island members), and the Sebascodegan, Cape Elizabeth, Spring Point, Diamond Island, Scarboro, Spurwink, Jewell, and Macworth Formations of the Casco Bay Group. The western package is referred to for discussion purposes as the Falmouth-Brunswick sequence, and the eastern package as the Saco-Harpswell sequence.

Major west-dipping thrust faults separate 1) the Kearsarge-central Maine sequence from the Falmouth-Brunswick sequence (the Hackmatack Pond fault), 2) the Falmouth-Brunswick sequence from the Saco-Harpswell sequence (the Beech Pond fault), and 3) the Sebascodegan formation of the Boothbay area from the Cross River formation (Boothbay thrust). The Cross River formation is a correlative of sedimentary rocks of either the Penobscot sequence or the Benner Hill sequence. The Cape Elizabeth Formation overlies different units of the Cushing Formation of the Saco-Harpswell sequence either because of regional tilting and erosion of the Cushing Formation pile prior to Cape Elizabeth deposition or because of lateral facies variations within the Cushing Formation.

The uppermost unit of the Casco Bay Group, the Macworth Formation, is conformable to the underlying Jewell Formation, and based on lithic similarity may correlate with either the Vassalboro Formation or the Berwick Formation, thus establishing a sedimentary continuum either between the Saco-Harpswell and Kearsarge-central Maine sequences or between the Saco-Harpswell sequence and the Merrimack Group.

Intrusion of granite, granodiorite, and diorite occurred from Early Devonian to Mississippian time. Post-tectonic diabase and basalt dikes of Triassic to Cretaceous age are the youngest rocks exposed in the region.

All metasedimentary rock sequences were metamorphosed and multiply deformed during either the Acadian orogeny or an earlier, pre-Taconic orogeny. Minor late recumbent folds of the Falmouth-Brunswick sequence may be related to forceful intrusion effects of the Acadian orogeny. Major thrust faults are probable effects of the Acadian orogeny. Strike-slip movement on faults of the Norumbega fault zone predate emplacement of Mississippian plutons but are later than Acadian regional metamorphism. Normal faults are post-metamorphic and are probably related to initiation of continental rifting in Late Triassic time.

INTRODUCTION

The purpose of this paper is to review the stratigraphy, structure, correlations, age, metamorphism, and plutonic igneous activity of the rocks of the Casco Bay region, southwestern Maine (Fig. 1). This paper discusses the problems of stratigraphic correlation, age interpretation, and structural analysis of the metamorphosed stratified rocks of the greater Casco Bay region. Relations to other bounding sequences (Merrimack Group, Kearsarge-central Maine sequence, and units to the east of the Casco Bay sequence) are reviewed.

The stratigraphic sequence of the study area includes the Casco Bay Group, the Cross River formation, and the Sebascodegan formation, part of which was previously mapped as the southern Maine continuation of the Bucksport Formation typically exposed in eastern central Maine. The interpretation of the Casco Bay Group presented in this paper is that it is divisible into two distinct lithotectonic packages, the Falmouth-Brunswick sequence consisting of members of the Cushing Formation exposed west of the Flying Point fault and the Saco-Harpswell sequence consisting of members of the Cushing Formation east of the Flying Point fault plus formations of the Casco Bay Group above the Cushing Formation (Table 1 and Fig. 2).

TABLE 1. CORRESPONDENCE OF LITHOTECTONIC PACKAGES TO ROCK STRATIGRAPHIC UNITS OF THE CASCO BAY GROUP.

Rock stratigraphic units	Lithotectonic packages
Casco Bay Group	
Macworth Formation	
Jewell Formation	
Spurwink Metalimestone	
Scarboro Formation	
Diamond Island Formation	
Spring Point Formation	Saco-Harpswell sequence
Cape Elizabeth Formation	
Sebascodegan formation	
Cushing Formation	
Wilson Cove member	
Bethel Point member	
Merepoint member	
Yarmouth Island member	
Peaks Island member	
Nehumkeag Pond member	
Mount Ararat member	Falmouth-Brunswick sequence
Torrey Hill member	
Richmond Corner member	

As a result of recent mapping, the rocks previously represented as the Bucksport Formation in the Boothbay area are here interpreted to be stratigraphic equivalents of the lithically similar Sebascodegan member of the Cushing Formation (Hussey, 1985). Because of the regional extent and great thickening of this unit in the Boothbay area, it is here separated from the Cushing Formation and raised to formational rank. It interfingers with the Cushing Formation in the Brunswick area.

The Macworth Formation is interpreted to be either the equivalent of the Vassalboro Formation of the Kearsarge-central Maine sequence or the Berwick Formation of the Merrimack Group.

STRATIGRAPHY

Complete descriptions of the groups, formations, and members in the discussion area are presented in an open-file report (Hussey, 1985). The names of new formations and members proposed there are used informally in this discussion (all place names, however, have been cleared and reserved for future formal use). The informality of a name is indicated by not capitalizing the rock-stratigraphic rank of the unit (for example, Cross River formation, Merepoint member). In the following discussion, only the more generalized character of these units will be given.

Falmouth-Brunswick Lithotectonic Sequence

The members of the Cushing Formation as earlier defined (Hussey, 1985) lying west of the Flying Point fault (Fig. 2) constitute a lithotectonic sequence distinct from the Cushing Formation east of the Flying Point fault. Referred to here as the Falmouth-Brunswick sequence, this package consists of the Nehumkeag Pond, Mount Ararat, Torrey Hill, and Richmond Corner members of the Cushing Formation. There are four reasons for separating this sequence from the rest of the Cushing Formation. 1) The Cushing Formation east of the Flying Point fault lacks the thin alternations of amphibolite and quartz-feldspar-biotite granofels typical of the Mount Ararat member, areally the most extensive unit of the Falmouth-Brunswick sequence. Amphibolites within the Cushing Formation east of the Flying Point fault are much thicker and more localized. 2) There is no correlation of lithologic sequences from one side of the Flying Point fault to the other. 3) The sequence west of the Flying Point fault, and not that to the east, is similar to metavolcanic sequences of pre-Silurian age in New Hampshire, and eastern and central Massachusetts. 4) Seismic reflection evidence suggests that the Falmouth-Brunswick sequence has been thrust in from the west.

The Nehumkeag Pond member consists of thin- to medium-bedded and massive, non-rusty-weathering, medium to light gray, quartz-plagioclase-K-feldspar-biotite-muscovite gneiss and granofels. It has been moderately migmatized and injected by non-foliated and weakly foliated pegmatite. Prior to metamorphism these rocks were felsic to intermediate pyroclastic volcanic rocks. Minor units within the member include thin but mappable lenses of marble and calc-silicate gneiss, rusty gneiss and schist, and massive amphibolite.

The Mount Ararat member, the most extensive member of the Falmouth-Brunswick sequence, consists of alternations of light gray, quartz-plagioclase-biotite gneiss and granofels, and dark gray amphibolite commonly interbedded on a scale of 2 to 10 cm. The Mount Ararat member includes a rusty pelitic

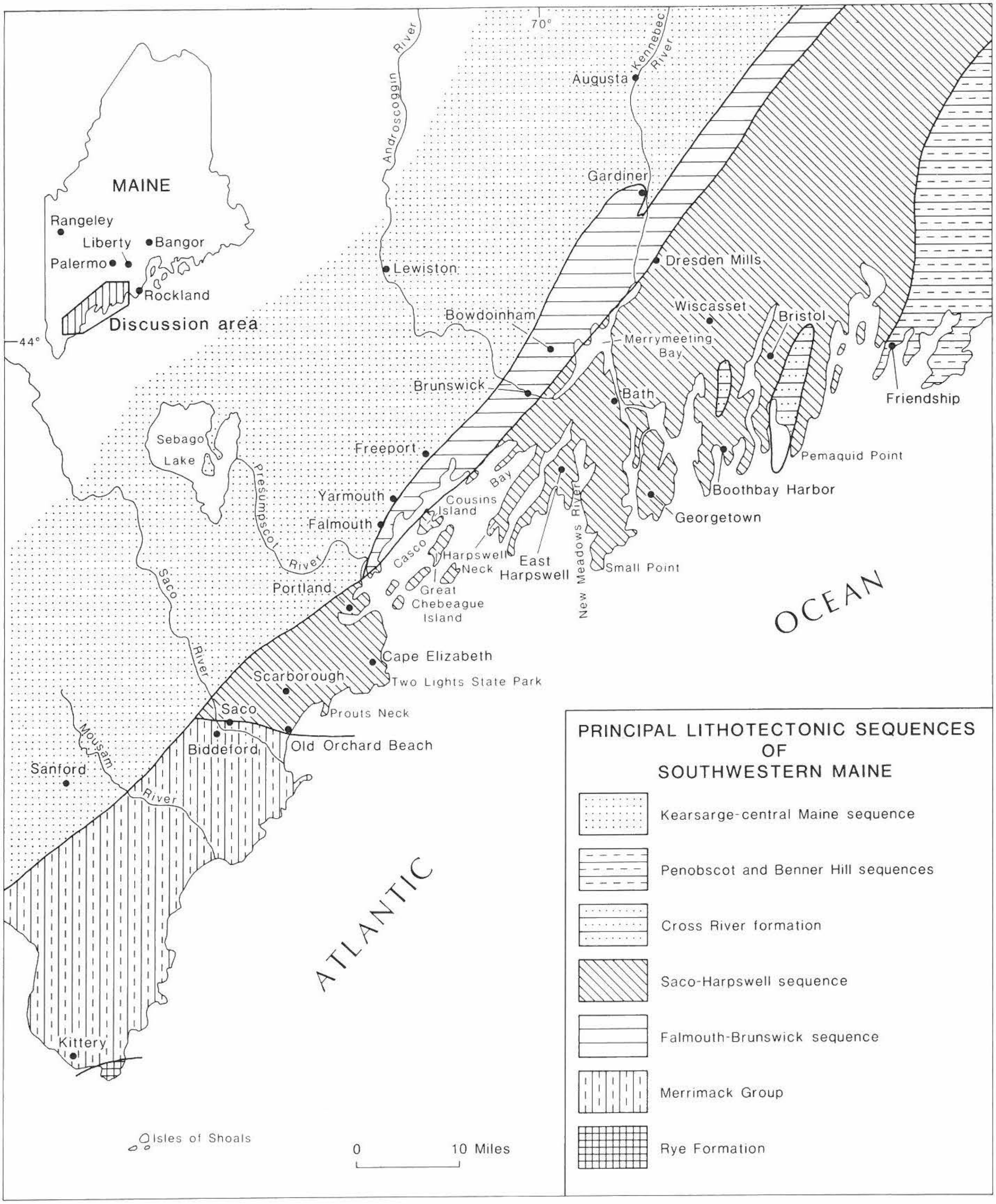

Figure 1. Map showing location of study area, principal lithotectonic sequences of southwestern Maine, and names of places referred to in the text.

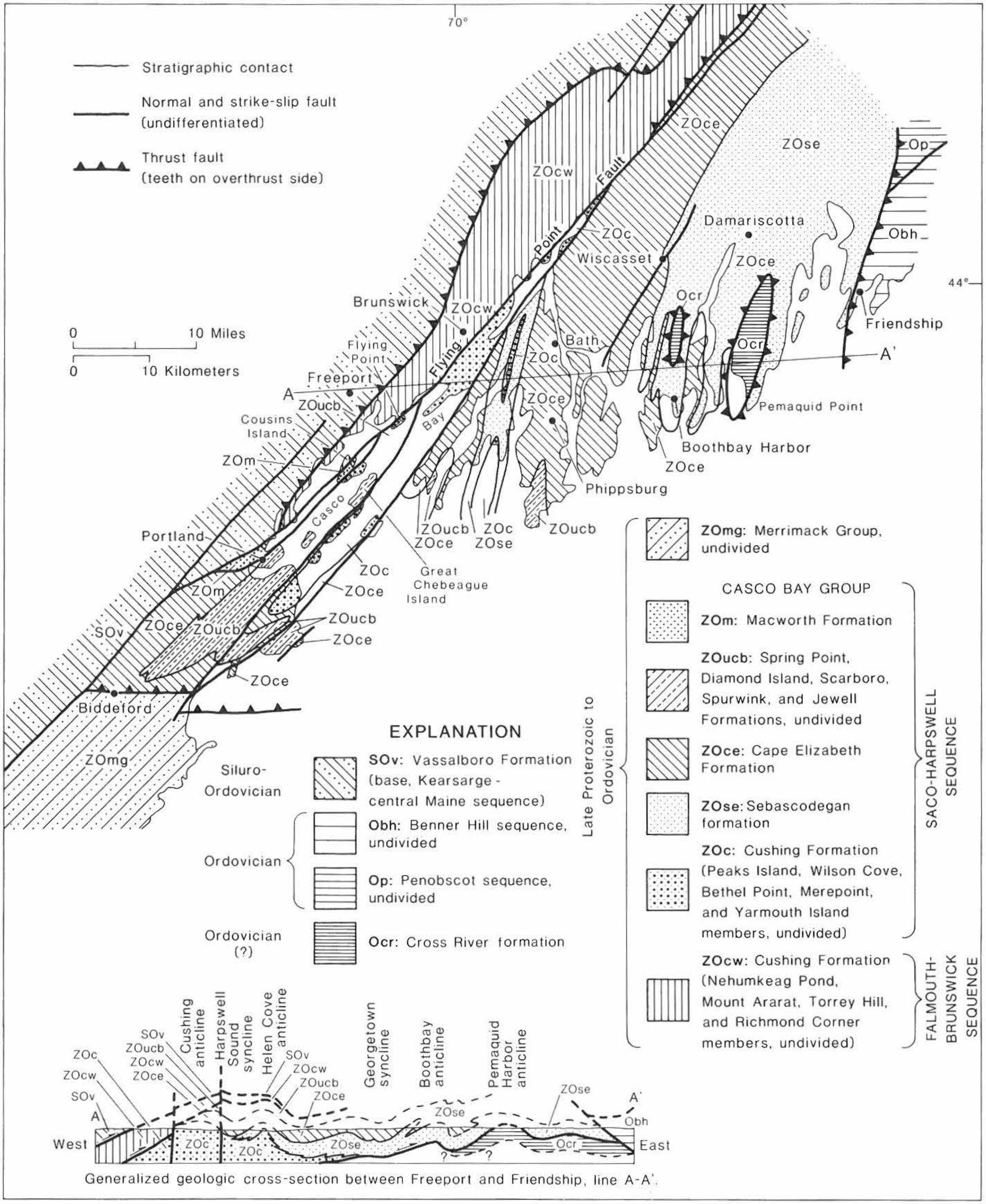

Figure 2. Generalized geologic map and cross section of the greater Casco Bay area, Maine. Plutonic rocks are omitted to emphasize lithotectonic relations.

schist lens and occasional mappable massive amphibolite lenses. The abundance of thin amphibolite beds distinguishes this unit from the Nehumkeag Pond member and from members of the Cushing Formation east of the Flying Point fault. Prior to metamorphism, the biotite gneiss/granofels and thin amphibolite layers were probably felsic and basic volcanic tuffs.

The Torrey Hill member is a thin unit consisting almost entirely of very sulfidic, graphite-bearing, sillimanite-rich schist. The Richmond Corner member is predominantly medium gray, biotite-garnet-quartz-feldspar, salt and pepper-textured schist. Interbedded pink coticule and zones of rusty-weathering, biotite-quartz-feldspar schist are common. Sillimanite is locally present in both rusty and non-rusty schist. The Torrey Hill and Richmond Corner members were pelitic sediments prior to metamorphism.

Contacts between formations of the Falmouth-Brunswick sequence have not been observed but are inferred to be conformable. Exposures close to the mapped position of these contacts lack all indications of structural discordance. No evidence for the stratigraphic sequence of these formations has been observed to date. The Hackmatack Pond fault (Pankiwskyj, 1978) separates the Falmouth-Brunswick sequence from the Vassalboro Formation of the Kearsarge-central Maine sequence. This fault coincides with a west-dipping reflector in the Quebec-western Maine seismic reflection profile (Stewart et al., 1986) in the Palermo area (north of the study area). The Beech Pond fault (Newberg, 1985) and the Flying Point fault (Hussey, 1985) separate the Falmouth-Brunswick sequence on the west from the Saco-Harpswell sequence on the east. The Flying Point fault is the contact southwest from Dresden Mills, and the Beech Pond fault is the contact northeast from Dresden Mills. The Beech Pond fault approximately coincides with the emergence of a west-dipping reflector on the Quebec-western Maine seismic reflection profile and is interpreted as a major west-dipping thrust. This fault is the principal boundary between the Falmouth-Brunswick and Saco-Harpswell lithotectonic sequences; south of Dresden Mills it is offset by the Flying Point fault.

Saco-Harpswell Lithotectonic Sequence

The Casco Bay Group (Hussey,1985) east of the Flying Point fault constitutes a second lithotectonic package, referred to here as the Saco-Harpswell sequence. This package consists of the Peaks Island, Yarmouth Island, Bethel Point, Mere Point, and Wilson Cove members of the Cushing Formation east of the Flying Point fault, and the Sebascodegan, Cape Elizabeth, Spring Point, Diamond Island, Scarboro, Spurwink, Jewell, and Macworth Formations.

The lower part of this sequence, the Cushing Formation, consists predominantly of quartzo-feldspathic metavolcanic rocks and volcanogenic metasedimentary rocks with minor amphibolite, and pelitic and calcareous metasedimentary rocks. In general, the Cushing Formation east of the Flying Point fault shows a variation from a volcanic-dominant sequence to a sedimentary-dominant sequence from west to east in its outcrop belt. The Peaks Island member consists of dominantly massive but occasionally thin- to medium-bedded felsic to intermediate pyroclastic metavolcanic rocks with minor volcanogenic metasedimentary rocks. The metavolcanic rocks range from rhyolite to dacite. Volcaniclastic structures such as relict phenocryst fragments and breccia blocks of varying felsic to intermediate volcanic composition are preserved in outcrop areas from South Portland to Harpswell Neck. Amphibolites are rare in this member.

The Wilson Cove member is the uppermost member of the Cushing Formation. It consists of black, locally very sulfidic garnet-rich biotite schist, with minor sulfidic amphibolite and two-mica schist. The Wilson Cove member interfingers with felsic metavolcanics of the Peaks Island member on Harpswell Neck. It is locally absent in the central parts of Casco Bay but is again present on Prouts Neck in Scarborough*.

The Merepoint member of the Cushing Formation is a sulfidic and feldspathic two-mica schist with minor sillimanite. It is exposed in the core of the Merepoint anticline in the northern part of Casco Bay.

The Yarmouth Island and Bethel Point members are exposed in the core of the Hen Cove anticline (Fig. 3). The lower of these, the Yarmouth Island member, consists of weakly thin-bedded to massive, felsic to intermediate metavolcanic rocks with interbedded sillimanite and/or staurolite-bearing feldspathic gneisses, and zones of calc-silicate gneiss and amphibolite up to 20 m thick. Gedrite and cordierite are common in the felsic to intermediate metavolcanic rocks.

The Bethel Point member is a uniform sequence of very sulfidic two-mica schist with minor thin beds of two-mica feldspathic quartzite. This member is correlated with the Merepoint member of the Cushing Formation on the basis that both are semipelitic to pelitic sulfidic schists.

The Sebascodegan formation was formerly the Sebascodegan member of the Cushing Formation, being the uppermost member of the formation in the East Harpswell area. It is raised to formational status here by its extension into the Boothbay Harbor area and its correlation with rocks formerly mapped as the Bucksport Formation (Osberg et al., 1985). These rocks apparently thicken to the east and are exposed over a broad enough area to warrant formational rank. Through facies change, the Sebascodegan formation grades into that part of the Peaks Island member of the Cushing Formation above the Merepoint member in the general vicinity of the coast just south of Brunswick.

In East Harpswell (Fig. 1), the Sebascodegan formation consists of interbedded thin- to medium-bedded, quartz-plagioclase-biotite gneiss and granofels, greenish gray calc-silicate gneiss, and sillimanite-bearing quartzo-feldspathic biotite gneiss and

*At present, the official spelling of the town is Scarborough. The original spelling of the formation named after that town by Katz (1917) was Scarboro, and the latter spelling is used throughout this discussion for referring to the formation.

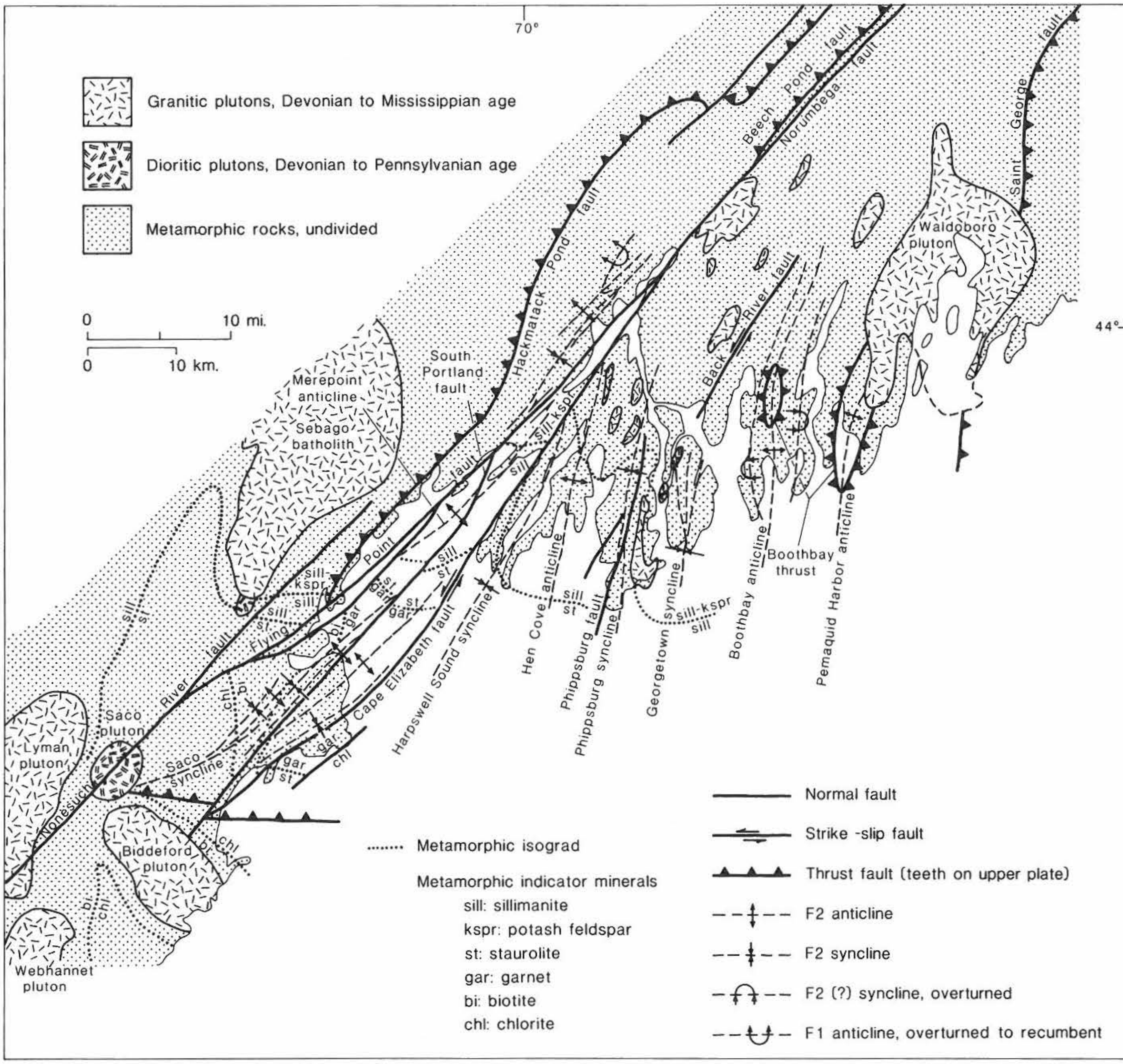

Figure 3. Structural, metamorphic, and plutonic map of the greater Casco Bay area, southwestern Maine.

granofels. Prior to metamorphism these were felsic to intermediate pyroclastic volcanic rocks, and calcareous and argillaceous feldspathic wackes reworked from them. The quartzo-feldspathic biotite gneiss and granofels locally preserve relict volcanic breccia structures similar to those of the Peaks Island member of the Cushing Formation. In the Harpswell area, the Sebascodegan formation includes lenses up to 50 meters thick of calc-silicate granofels or gneiss with very minor biotite granofels. It also includes sulfidic biotite granofels lenses up to 50 meters thick, and amphibolite lenses locally with thin marble.

The amphibolite occurs as three distinct lithologic types: 1) fine-grained, evenly foliated, salt and pepper-textured, plagioclase-hornblende amphibolite, probably of volcanic origin; 2) coarse-textured, hornblende-andesine-biotite amphibolite, the texture of which is suggestive of intrusive diorite parentage; and 3) association of mineralogically complex amphibolite, calc-silicate gneiss, and phlogopitic biotite schist. A characteristic lithic sequence of this latter type amphibolite is a central thin layer of calc-silicate bounded in order on each side first by cummingtonite-anthophyllite-hornblende amphibolite and then by bronze-colored, phlogopitic biotite schist. The

presence of calc-silicate in the assemblage and the calcic composition of the plagioclase (bytownite) in all the associated lithologies suggests a non-igneous parentage for type 3 amphibolites. They may represent calcareous or dolomitic sediments interbedded with shale. The amphibolite layers may be the product of metamorphic differentiation between these two lithologies.

Units of the Casco Bay Group above the Cushing Formation (Table 1) constitute a conformable sequence of pelitic and semipelitic metasedimentary rocks with minor metavolcanic rocks, ribbon metalimestone, and metapsammite. Areally, the most extensive and probably thickest unit of this part of the Casco Bay Group is the Cape Elizabeth Formation which directly overlies the Cushing Formation and Sebascodegan formation. The Cape Elizabeth Formation is a thin- to medium-bedded association of quartz- plagioclase-biotite-muscovite schist or phyllite, and muscovite-biotite schist or phyllite. A variety of aluminosilicate minerals are present. These include sillimanite, staurolite, garnet, and more rarely andalusite, depending on grade of metamorphism. Minor lithologies within the Cape Elizabeth Formation include amphibolite, which may be associated with calc-silicate and marble, sulfidic sillimanite or staurolite-rich schist, and coticule beds (pink garnet-biotite-quartz granofels). Amphibolite commonly forms non-mappable beds that are strongly boudined and extensively chloritized.

The Spring Point Formation consists primarily of metamorphosed basic and intermediate volcanic tuffs and flows. These are represented now by chlorite-spessartite phyllite, actinolite gneiss, and hornblende-garnet amphibolite locally with cummingtonite and rarely anthophyllite. The upper part of the formation, particularly in the Harpswell area, consists of thin-bedded, quartzo-feldspathic, biotite-muscovite granofels with local thin interbeds of chlorite-hornblende schist.

The Diamond Island Formation is a distinctive thin (35-50 m) unit of finely foliated sulfidic, black, quartz-muscovite-graphite phyllite. It rarely shows bedding.

The Scarboro and Jewell Formations are essentially identical in lithology, consisting of sulfidic and non-sulfidic phyllite or schist. Garnet, staurolite, chiastolite, and chloritoid are present at appropriate grades of metamorphism. Prior to metamorphism, these schists and phyllites were sulfidic and non-sulfidic shales. Both formations have local lenses up to several tens of meters thick of very greenish, chloritic phyllite interpreted to represent basic pyroclastics.

The Spurwink Metalimestone separates the Jewell and Scarboro Formations. It is a thin-bedded alternation of ribbony-appearing, gray, fine-grained marble and quartz-biotite phyllite. The quartzose phyllite beds are extensively boudined.

The Macworth Formation is the highest formation of the Casco Bay Group. It is a monotonous sequence of drab brownish gray, thin-bedded to thinly laminated, calcareous quartz-biotite phyllite. On the eastern edge of Cousins Island in Casco Bay, presumably at a higher grade of metamorphism, the formation consists of thin-bedded alternations of quartz-plagioclase-biotite granofels and greenish, calc-silicate granofels. These lithologies are closely similar to the Vassalboro and Berwick Formations at intermediate grade of metamorphism. On the north end of Great Chebeague Island in Casco Bay a sequence of drab, thinly laminated, grayish brown phyllite above the Jewell Formation is mapped and correlated as part of the Macworth Formation. This is a critical correlation inasmuch as it establishes the stratigraphic position of the Macworth Formation as the highest unit of the Casco Bay Group.

Recognition of the formations of the Casco Bay Group above the Cushing Formation in the area east of the New Meadows River (Fig. 1) is uncertain because of thickness changes, apparent facies changes, and repetition of amphibolites (all of which are possible correlatives of the Spring Point Formation) in the stratigraphic column. Near the western shore of the town of Phippsburg, on the east side of the New Meadows River, the writer (unpublished work) has mapped an amphibolite with associated calc-silicate skarn and marble. The same unit defines the northern part of the Phippsburg syncline (Fig. 3) south of Bath. A similar amphibolite without the calc-silicate skarn, but with coticule and rusty schist defines the Georgetown syncline (Fig. 3). A conspicuous, multiply deformed amphibolite of similar lithology also crops out in the Boothbay Harbor area (Hussey, 1986). Finally, an amphibolite with associated marble occupies a narrow belt approximately 3 km long in the west edge of the city of Bath. In all these belts, the amphibolite and associated rocks occur within typical Cape Elizabeth rocks, pelitic and psammitic. One or the other of these may correlate with the Spring Point Formation or they may simply be one or more lenses in the Cape Elizabeth Formation.

In the core of the Phippsburg syncline in the Small Point area south of Bath, there is a narrow belt of amphibolite (in part sulfidic) and associated calc-silicate gneiss and marble. Adjacent to this is a belt of rusty-weathering black phyllite. The black phyllite is identical to the Diamond Island Formation. The adjacent amphibolite is less like the typical Spring Point Formation as mapped at high metamorphic grade in the Orrs Island area (Hussey, 1971b) than are the amphibolites noted above. The rocks above and below the amphibolite/black phyllite sequence are indistinguishable from each other. These consist of sillimanite or andalusite-rich metapelites with thin intervals of Cape Elizabeth-like, quartz-feldspar-biotite-garnet schist. As a whole, these schists are not like the typical Cape Elizabeth lithology or the typical Scarboro-Jewell lithology. Associated with these aluminous pelites are lenses of metasedimentary rock types (thin-bedded calc-silicate and biotite granofels, garnet-biotite schist, very rusty-weathering, light gray muscovite schist, and calc-silicate marble) that are not present in either the Cape Elizabeth Formation or Scarboro-Jewell sequence elsewhere. If these differences are attributed to facies variations, then the correlation of the black phyllite with the Diamond Island Formation is reasonable. Correlation of the associated amphibolite with the Spring Point Formation would follow, and the amphibolite belts mentioned above would constitute a member within the Cape Elizabeth Formation. This correlation is currently favored.

Contact Relations. Contacts have been observed between the

Cape Elizabeth Formation and members of the Cushing Formation and between all other formations of the Casco Bay Group, except the Macworth and Jewell Formations. With the exception of the contact between the Cape Elizabeth and Cushing Formations (see discussion below), all are sharp and apparently conformable. The Macworth Formation is mostly fault-bounded with the Flying Point fault on the northwest side and an unnamed normal fault on the southeast (Fig. 2). The Torrey Hill member of the Falmouth-Brunswick sequence and the Vassalboro Formation occur on the northwest side of the Macworth Formation, and various units of the Saco-Harpswell sequence adjoin the formation on its southeast side (Fig. 2). On the north end of Great Chebeague Island in Casco Bay the actual contact between the Macworth and Jewell Formations is concealed by a short interval of mud-flat cover, but structural evidence suggests that the Macworth Formation is conformable with the Jewell Formation.

The Cape Elizabeth Formation overlies the Sebascodegan formation on both limbs of the Hen Cove anticline in the Harpswell area. On the east limb of the Cushing anticline it overlies either the Wilson Cove member of the Cushing Formation or, where that is absent, the Peaks Island member. On the west side of the Cushing anticline, the Wilson Cove member is everywhere absent and the Cape Elizabeth Formation overlies the Peaks Island member. Close to the contact on that side, and within the Peaks Island member, are two zones of very rusty-weathering white muscovite schist that are not seen on the eastern side of the Cushing anticline. On the east side of the Cushing anticline at Chimney Rock in Cape Elizabeth, the contact between the Cape Elizabeth Formation and the Peaks Island member of the Cushing Formation is occupied by a 1 m thick zone of quartz-feldspar, fine granule metaconglomerate suggesting the possibility of an unconformable contact. At other localities, the contact appears conformable. These relations, i.e., the Cape Elizabeth Formation overlying different members of the Cushing Formation and the apparent across-strike variations within the Cushing Formation, raise the following speculations about the nature of the contact between the Cape Elizabeth and Cushing Formations, and the relationship between the different members of the Cushing Formation (Fig. 4). The across-strike variations within the Cushing Formation may be explained by members of the Cushing Formation that have been tilted and are younging toward the east below the Cape Elizabeth Formation (Fig. 4a). This would suggest either a major unconformity or possibly a major thrust contact. Alternatively, the relations may reflect facies changes within the Cushing Formation from a volcanic pile on the west to an apron of reworked volcanic detritus on the east. In this case the Cape Elizabeth-Cushing contact is either conformable or locally disconformable (Fig. 4b). Critical field data to distinguish between these hypotheses are lacking; however, model b is favored because of the correlation of the Merepoint member of the Cushing Formation with the Bethel Point member.

Regional correlations and age of the units of the Casco Bay Group will be discussed in a following section.

Cross River formation

The Cross River formation is exposed in the cores of two antiforms in the Boothbay area. The principal lithology is extremely migmatized, very rusty-weathering, sulfidic, sillimanite-rich and locally graphite-rich gneiss and schist, with local beds of feldspathic biotite quartzite often represented only as disoriented rafts in the migmatite. At the top of the formation in the western of the two exposure belts, is a thin, but mappable, unnamed member consisting of non-migmatized, salt and pepper-textured amphibolite, and biotite granofels with large garnet porphyroblasts.

The Cross River formation is similar to sulfidic schists of both the Benner Hill sequence of Osberg and Guidotti (1974) and the Penobscot Formation of Bickel (1976) (Fig. 2). Both of these lie east of the Sebascodegan/Bucksport outcrop belt and are separated from the Cross River formation by the St. George fault, an east-dipping thrust (Fig. 3).

AGE RELATIONS AND CORRELATIONS

Paleontological Control

Age relations of the rocks of the greater Casco Bay area are poorly known. No fossils occur within rocks of the Casco Bay Group, Cross River formation, and Sebascodegan formation. The only paleontological control is for sequences bounding these packages. Strongly deformed brachiopods of Ordovician age (Boucot et al., 1972; Neuman, 1973) occur in a quartzite lens

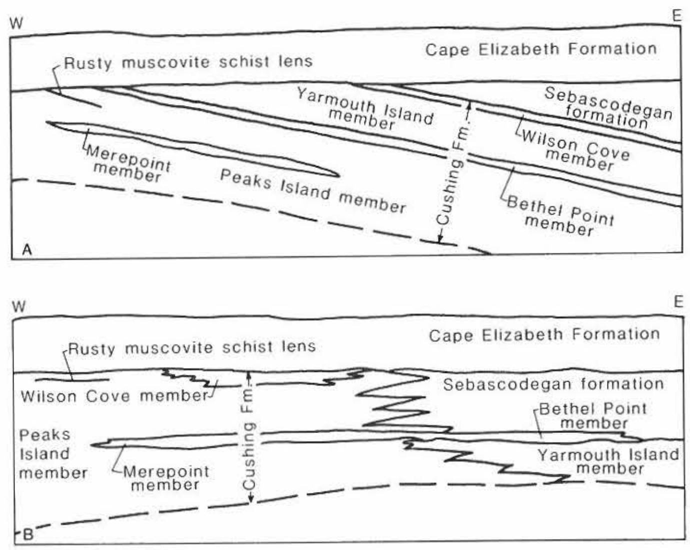

Figure 4. Alternative models for the relationship of the Cape Elizabeth Formation to the underlying units of the Cushing Formation. (a) Members of the Cushing Formation form a tilted sedimentary pile beneath the Cape Elizabeth Formation; (b) Members of the Cushing Formation are a series of sedimentary facies below the Cape Elizabeth Formation. Alternative B is preferred because it allows the suggested correlation of rusty schist on Merepoint, Brunswick, with the Bethel Point member of the Cushing Formation.

in the Benner Hill sequence of Osberg and Guidotti (1974) near Rockland, Maine. Graptolites of uncertain, but inferred Early Silurian age have been reported from the Vassalboro Formation of the Kearsarge-central Maine sequence (Osberg, 1980). Arenig (Early Ordovician) age graptolites have been recovered from the Cookson Formation, interpreted to be a correlative of the Penobscot Formation (Ruitenberg and Ludman, 1978). The usefulness of this limited paleontological control depends on the validities of inferences made regarding stratigraphic and structural relations of the different stratigraphic packages involved.

Radiometric Ages of Metamorphic Rocks

Relatively few radiometric age determinations are available for inferring the ages of the stratified rocks of the greater Casco Bay area (Table 2). Gaudette et al. (1983) report an Early Ordovician Rb/Sr whole rock age for the Mount Ararat member of the Cushing Formation in the Gardiner area (Fig. 1). Brookins and Hussey (1978) give an Early Ordovician Rb/Sr whole rock age for the Cape Elizabeth and Cushing Formations of the Casco Bay Group. They report an Rb/Sr whole rock age of 481 ± 40 Ma for the Cushing Formation, but this represents a mix of ages of the Mount Ararat member of the Cushing Formation of the Falmouth-Brunswick sequence and the Peaks Island member of the Cushing Formation of the Saco-Harpswell sequence. These members were not separately mapped at the time.

The interpretation of the whole-rock Rb/Sr ages of the metavolcanic and metasedimentary sequences has been the subject of much confusion. Brookins (pers. commun., 1986) believes these ages represent time of formation of the rocks, that is, time of sedimentation/diagenesis and volcanism. Others, (e.g., Lyons et al., 1982), argue that such radiometric ages are reset at least partially, if not entirely, by major episodes of metamorphism and therefore represent a time of orogeny and metamorphism, the time of formation of the rocks being somewhat earlier than this. This question must be resolved by radiometric age specialists before such age determinations can be applied to the geochronological problems of the lithotectonic sequences of the greater Casco Bay area.

Radiometric Ages of Intrusive Rocks

Radiometric ages of intrusive rocks (Table 2) establish a minimum age of Middle to Late Devonian for the stratified rocks of the greater Casco Bay area. Knight and Gaudette (1987) report a Late Devonian Rb/Sr age for the Waldoboro pluton (Fig. 3) which intrudes the Cross River and Sebascodegan formations. Brookins (pers. commun., 1976) obtained a whole rock Rb/Sr age of 385 Ma for weakly foliated and folded, generally concordant pegmatites of the Falmouth area, and 375 Ma for discordant, evenly-walled unfoliated pegmatite dikes in the same area. These ages suggest injection of the pegmatites during late stages of the Acadian orogeny.

A minimum age of Middle Ordovician is indicated for the Merrimack Group that bounds the Casco Bay Group on the south (Fig. 1). Gaudette et al. (1984) report a Middle Ordovician Rb/Sr age for the Exeter pluton that intrudes the Kittery and Eliot Formations of the Merrimack Group in southeastern New Hampshire. Zartman and Naylor (1984) give a Late Ordovician zircon age for the Newburyport pluton that intrudes the Kittery Formation in southeastern New Hampshire and adjacent Massachusetts. Gaudette et al. (1984) also indicate a minimum age of Middle Ordovician for the Rye Formation of southwestern Maine and southeastern New Hampshire. They report a 477 Ma age for a small diorite pluton that intrudes the Rye Formation on Appledore Island in the Isles of Shoals off Kittery.

Correlations and Ages of the Stratified Rocks of the Greater Casco Bay Area

Cape Elizabeth Formation. The Cape Elizabeth Formation traces into the Hogback Schist described by Perkins and Smith (1925) in the Palermo area and is probably equivalent to the high-grade sillimanite-bearing migmatites of the Passagassawakeag Gneiss of Bickel (1976). The following are some additional possible correlations of the Cape Elizabeth formation based on lithic similarity:

1) With the non-rusty schists of the Benner Hill sequence. In the Friendship area (Fig. 1) just east of the study area, numerous septa of Sebascodegan-like rocks in the southern end of the

TABLE 2. RADIOMETRIC AGES OF METAMORPHIC AND PLUTONIC ROCKS BEARING ON THE GEOCHRONOLOGY OF THE GREATER CASCO BAY AREA, SOUTHWESTERN MAINE

Rock unit	Method	Age (Ma)	Reference
		METAMORPHIC	
Mt Ararat m, Cushing Fm	Rb-Sr	494 ± 25	Gaudette et al. (1983)
Cushing Fm	Rb-Sr	481 ± 40	Brookins and Hussey (1978)
Cape Elizabeth Fm	Rb-Sr	485 ± 30	Brookins and Hussey (1978)
		PLUTONIC	
Diorite, Isles of Shoals	Rb-Sr	477	Gaudette et al. (1984)
Exeter pluton	Rb-Sr	473 ± 37	Gaudette et al. (1984)
Webhannet pluton	Rb-Sr	390 ± 10	Gaudette et al. (1982)
Webhannet pluton	Zr	403 ± 13	Gaudette et al. (1982)
Three Mile Pond pluton	Rb-Sr	381 ± 14	Dallmeyer and Van Breeman (1981)
Newburyport quartz diorite	Zr	450 ± 15	Zartman and Naylor (1984)
Biddeford pluton	Rb-Sr	344 ± 12	Gaudette et al. (1982)
Pegmatite, Falmouth area			
Discordant	Rb-Sr	375	Brookins (pers. commun. 1976)
Concordant	Rb-Sr	385	Brookins (pers. commun. 1976)
Sebago batholith	Zr	325 ± 3	Aleinikoff et al. (1985)
Lyman pluton	Rb-Sr	322 ± 12	Gaudette et al. (1982)
Waldoboro pluton	Rb-Sr	367 ± 4	Knight and Gaudette (1987)
Saco pluton	Rb-Sr	307 ± 20	Gaudette et al. (1982)

Waldoboro pluton crop out close to non-rusty schists that are lithically very similar to the Cape Elizabeth Formation in the Boothbay Harbor area (Hussey, 1972). These schists have been traced by Newberg (1979) into the outcrop belt of the Benner Hill sequence of Osberg and Guidotti (1974). As noted above, fossils from this sequence (Boucot et al., 1972; Neuman, 1973) indicate an Ordovician age for these rocks. If this correlation is valid, the Cape Elizabeth Formation would be of Ordovician age.

2) With the Eliot Formation of the Merrimack Group. The Cape Elizabeth Formation at low metamorphic grade in the Saco area is lithologically identical to the Eliot Formation, the middle unit of the Merrimack Group in southwestern Maine. This correlation would suggest a Late Precambrian to Ordovician age for the Cape Elizabeth Formation. This correlation might preclude a correlation with the Benner Hill sequence in view of the seemingly greater age of the Merrimack Group. The Merrimack Group is older than the 473 Ma Rb/Sr age obtained by Gaudette et al. (1983) for the Exeter pluton that intrudes the group in southeastern New Hampshire. Prior to intrusion, the Merrimack Group had been folded first by large-scale (?) recumbent folds and then refolded by upright folds that control the map-pattern of the formations of the group. In view of this structural history and the age of the Exeter pluton, they suggest that the Merrimack Group might be as old as Late Precambrian, and no younger than Middle Ordovician.

3) With the Rye Formation. The high-grade Cape Elizabeth Formation of the Georgetown area is lithically similar to metasedimentary rocks exposed on Appledore Island in the Isles of Shoals offshore of Portsmouth, New Hampshire. These rocks are correlated with the Rye Formation of southwestern Maine and southeastern New Hampshire, and possibly with the Nashoba Formation of eastern Massachusetts. Hepburn and Munn (1984) obtained an Rb/Sr Early Middle Ordovician metamorphic (?) age for the Nashoba Formation, very similar to the age reported by Brookins and Hussey (1978) for the Cape Elizabeth Formation (Table 2).

Upper Part of the Casco Bay Group. Units of the Casco Bay Group above the Cape Elizabeth Formation have been mapped in the Liberty area to the north of the study area by Pankiwskyj (1978) where they are exposed in a syncline southeast of the Norumbega fault. Due to plunge reversals, these rocks cannot be traced into the exposures at the southern end of the Casco Bay synclinorium. The upper units of the Casco Bay Group correlate with no other sequences in adjacent parts of the northern Appalachians. Their stratigraphic position, conformably above the Cape Elizabeth Formation, suggests a similar age for them, i.e., Early to Middle Ordovician at the youngest, and Late Precambrian at the oldest.

Sebascodegan formation of the Boothbay Harbor Area. The biotite and calc-silicate granofels sequence of the Boothbay Harbor area, now mapped as part of the Sebascodegan formation of the Casco Bay Group, closely resembles the Bucksport Formation of east-central Maine. With that formation they heretofore have been correlated (Osberg et al., 1985). These granofels also resemble the Kittery and Berwick Formations of the Merrimack Group and the Vassalboro Formation of the Kearsarge-central Maine synclinorium. Stewart (pers. commun., 1987) notes significant differences in bedding style and the amount of calc-silicate between the Sebascodegan formation rocks of the Boothbay Harbor area and the type Bucksport Formation near the town of Bucksport to the northeast and suggests they are not correlatable. Significantly, the Bucksport-like rocks are overlain conformably by the Cape Elizabeth Formation in the Boothbay Harbor area, just as the Sebascodegan formation is overlain by the Cape Elizabeth Formation in the eastern part of Harpswell, further strengthening the correlation of the calc-silicate and biotite granofels of the Boothbay Harbor area with the Sebascodegan formation. This poses the problem of how to separate the outcrop belt of the Sebascodegan formation from that of the Bucksport Formation on strike with it to the north east. The map pattern of this belt as it is known at this time (Osberg et al., 1985) does not readily allow this interpretation. This raises the question of whether they do correlate and that the Bucksport Formation is not of Silurian to Devonian age but of Late Precambrian to Ordovician age. This remains a major problem for resolution through future mapping and discussion.

Macworth Formation. The Macworth Formation is similar at its highest metamorphic grade to the Vassalboro Formation of the Kearsarge-central Maine sequence, although it is perhaps a bit less feldspathic and quartzose than the latter. It also resembles other calc-silicate and biotite granofels sequences (Berwick, Kittery, and Sebascodegan Formations) of the discussion area. At our present state of understanding, two possible correlations of the Macworth Formation should be entertained (Table 3).

1) The Macworth Formation correlates with the lower part of the Vassalboro Formation, perhaps parts not exposed northwest of the Norumbega fault (Table 3, alternative 1). This correlation would establish the age of the Macworth Formation as Middle to Late Ordovician. Because of the apparent stratigraphic conformity between the Macworth and Jewell Formations described above, a consequence of this correlation would be to establish a depositional continuum between the Casco Bay and Kearsarge-central Maine packages.

2) The Macworth Formation correlates with the Berwick Formation of the Merrimack Group and not the Vassalboro Formation, the latter being equivalent to the type Bucksport Formation to the northeast (Table 3, alternative 2). The Sebascodegan formation in Boothbay Harbor could conceivably be equivalent to the Kittery Formation and the Cape Elizabeth Formation to the Eliot Formation. The units of the Casco Bay Group between the Cape Elizabeth and Macworth Formations would have to pinch out to the southwest inasmuch as none of these lithologies are represented in the outcrop belt of the Merrimack Group. Such a pinch-out, at least for the Spring Point and Diamond Island Formations is already demonstrated (Hussey, 1971a); these two formations are not mapped around the south end of the Saco syncline (Fig. 3). This correlation would not establish any sedimentary relation between the Casco Bay Group and the Kearsarge-central Maine sequence, but would relate the

TABLE 3. PROPOSED STRATIGRAPHIC CORRELATION OF ROCKS OF THE GREATER CASCO BAY AREA.

Alternative 1: Macworth correlates with the Vassalboro Formation.

Southwestern Maine	Greater Portland Area	Harpswell	Boothbay Harbor	
Shapleigh Group	Vassalboro Fm Macworth Fm			Early Silurian to Late Ordovician
(fault)	CASCO BAY GROUP: Jewell Fm Spurwink Metals. Scarboro Fm Diamond Island Fm Spring Point Fm Cape Elizabeth Fm Cushing Fm (of Saco-Harpswell sequence) (fault)	Jewell Fm Spurwink Metals. Scarboro Fm Spring Point Fm Cape Elizabeth Fm Sebascodegan fm Cushing Fm (of Saco-Harpswell sequence)	Cape Elizabeth Fm Sebascodegan fm (fault) Cross River fm	Ordovician
MERRIMACK GROUP: Berwick Fm Eliot Fm Kittery Fm	(fault) Cushing Fm (of Falmouth-Brunswick sequence)			Ordovician to Late Proterozoic

Alternative 2: Macworth correlates with the Berwick Formation.

Southwestern Maine	Greater Portland Area	Harpswell	Boothbay Harbor	
Shapleigh Group	Vassalboro Fm			Early Silurian
(fault)	(fault)			
MERRIMACK GROUP: Berwick Fm Eliot Fm Kittery Fm	Macworth Fm CASCO BAY GROUP: Jewell Fm Spurwink Metals. Scarboro Fm Diamond Island Fm Spring Point Fm Cape Elizabeth Fm Cushing Fm (of Saco-Harpswell sequence)	Jewell Fm Spurwink Metals. Scarboro Fm Spring Point Fm Cape Elizabeth Fm Sebascodegan fm Cushing Fm (of Saco-Harpswell sequence)	Cape Elizabeth Fm Sebascodegan fm	Late Proterozoic to Early Ordovician
	(fault) Cushing Fm (of Falmouth-Brunswick sequence)		(fault) Cross River fm	Ordovician

Casco Bay Group to the Merrimack sequence. A choice between these two alternative correlations must await further field work and regional synthesis.

A correlation of the Vassalboro Formation with the Berwick Formation now appears to be untenable. Available information suggests that the Vassalboro Formation underwent both recumbent and upright deformation during the Acadian orogeny in Early Devonian time, whereas the Merrimack sequence was similarly deformed prior to the Middle Ordovician emplacement of the Exeter pluton.

Cushing Formation. Radiometric ages obtained by Gaudette et al. (1983) for the Mount Ararat member of the Cushing Formation (Table 2) indicate an Early Ordovician or older age for the Falmouth-Brunswick sequence.

The relations between the Cushing members of the Falmouth-Brunswick sequence and Cushing members of the Saco-Harpswell sequence is uncertain. Two alternatives must be entertained.

1) The two sequences are unrelated in space and perhaps, but not necessarily, in time to each other. As noted earlier, no correlations or evidence of stratigraphic superposition have been deduced that establish either a facies or relative time relationship between them. However, on a very speculative and tenuous basis, units of the Falmouth-Brunswick sequence, particularly the Mount Ararat member, show strong lithic similarities to several pre-Silurian metavolcanic sequences to the northwest and southwest including the stratified core rocks of the Oliverian domes and overlying Ammonoosuc volcanics of the Bronson Hill anticlinorium in western New Hampshire (Naylor, 1968), parts of the Massabesic Gneiss of southeastern New Hamsphire (Bothner et al., 1984), and the Monson Gneiss of the Pelham dome (Robinson et al., 1986). These rocks are also similar to Ordovician bimodal volcanics of the Miramichi anticlinorium in central western New Brunswick (Fyffe, 1982). The Falmouth-Brunswick sequence might then be a part of one of these sequences thrust in from the west over units of the Saco-Harpswell sequence.

2) The Falmouth-Brunswick sequence constitutes a basement terrane on which the Saco-Harpswell sequence was deposited. Contamination of subduction-produced intermediate volcanics by a felsic basement could account for the dominantly felsic nature of the volcanic rocks of the Saco-Harpswell sequence. The volcanic rocks in the Cushing Formation of the Saco-Harpswell sequence might conceivably be equivalent to the Ammonoosuc Volcanics, or be a separate sequence, depending on subduction models for Early Ordovician time.

These speculations naturally raise questions about the evolution of the area in terms of plate-tectonic activity – subduction, volcanic arc, and trench development, back-arc basin spreading, etc. It is, however, premature to develop these ideas further in view of uncertainties about ages of the rocks.

Cross River formation. The Cross River formation is correlated with either the Penobscot Formation or the rusty schist of the Benner Hill sequence, both of which are exposed on the east side of the St. George fault (Figs. 2 and 3). This would indicate a Cambrian to Ordovician age for the Cross River formation. The two exposures of the Cross River formation are windows through the Boothbay thrust, a major low-angle, west-dipping folded thrust, not the St. George fault (see section following).

STRUCTURE

Introduction

Major folds and faults that affect the rocks of the Casco Bay Group and surrounding sequences are shown in Figure 3. All sequences have been multiply folded, thrust-faulted, and normally faulted. Various interpretations for the major structure of the Casco Bay Group have been proposed in the past. North of the study area, Osberg (1974) referred to the belt as the Liberty-Orrington anticlinorium on the basis that older rocks (the Casco Bay Group) occur in the center of the structure. On the other hand, the structure of the south end of the belt, in the Portland-Boothbay Harbor area, was referred to as the Casco Bay synclinorium (Hussey, 1968) on the basis of the overall synclinorial pattern of the upright folds of the Casco Bay Group. The presence of older rocks supposedly surrounded by younger sequences was explained by postulating a major synclinorially folded klippe for the Casco Bay Group (Hussey, 1985; Osberg et al., 1985). Now, in light of the seismic reflection studies across the northern part of the study area (Stewart et al., 1986), the regional structure of this belt is herein reinterpreted to be dominated primarily by several major west-dipping thrust faults. This regional picture is complicated by normal and strike-slip faulting of the Norumbega fault system and multiple folding of the stratified rocks.

Folds

Two fold systems of regional extent affecting the Saco-Harpswell sequence are recognized. The older system, designated F_1, is characterized by nearly isoclinal recumbent folds. The younger, designated F_2, deforms the older folds and consists of upright to slightly overturned tight to open folds, locally with well-developed axial planar schistosity. Although large-scale recumbent folds such as postulated by Osberg (1980) and Eusden et al. (1987) for the Kearsarge-central Maine synclinorium cannot be demonstrated, mesoscopic-scale recumbent folds are locally observed in outcrop. At Chimney Rock in Cape Elizabeth, downward-facing F_2 folds (D. W. Newberg, pers. commun., 1982), deform the Cape Elizabeth-Peaks Island contact. This locality clearly establishes the stratigraphic sequence between these units; graded beds in the Cape Elizabeth Formation within a half meter of the contact indicate that the Cape Elizabeth Formation was deposited on top of the Cushing Formation. On the eastern side of Small Point, tight recumbent F_1 folds of thin calc-silicate and biotite granofels beds have been refolded by relatively open, upright F_2 folds. The Cape Elizabeth Formation at Two Lights State Park in Cape Elizabeth is deformed by tight F_1 recumbent isoclines and gentle, open F_2 folds (Fig. 5). On the west side of Small Point thin calc-silicate beds in the Cape Elizabeth Formation show small, mesoscopic-scale, recumbent F_1 folds with a well-developed fracture cleavage deformed by small-scale F_2 upright folds (Fig. 6a). The fracture cleavage of the calc-silicate beds and adjacent psammitic beds of the Cape Elizabeth Formation are deformed by an intermediate set of small-scale, tight, recumbent folds. The limbs of these tight isoclines (Fig. 6b) show marked pressure-solution thinning of the quartz-rich bands between cleavage planes. The relatively thicker quartz laminae in the hinges give a pseudo-bedded appearance. Quartz veins in adjacent mica schists of the Cape Elizabeth Formation at Small Point preserve evidence of similar stages of multiple deformation (Fig. 6c).

The map pattern of the formations of the Casco Bay Group is controlled by F_2 folds in addition to the various faults. F_2 folds are much larger in scale than any observed or mapped

recumbent folds. F_2 parasitic folds are most commonly preserved in the Cape Elizabeth Formation. In the Small Point area, F_2 folds are relatively open and deform the schistosity defined by muscovite. Locally, biotite is developed parallel to F_2 axial planes. In the Harpswell area, F_2 parasitic folds are tighter than at Small Point, and they have a well-developed biotite as well as muscovite schistosity (Hussey, 1971b). In the Portland and Cape Elizabeth areas, F_2 folds are locally very gentle, open structures (Hussey, 1971a).

Fold systems of the Falmouth-Brunswick sequence are not as well known, due in part to the scarcity of outcrop of units of the sequence and in larger part to the fact that much of the sequence is strongly migmatized. F_1 and F_2 fold systems are inferred from the map pattern of the units of the Cushing Formation west of the Flying Point fault. F_1 folds are represented by the refolded hinges of marble and sulfidic gneiss mapped by Newberg (1981) in the Bowdoinham area (Fig. 3). F_2 folds are the north-northeast-trending folds (Fig. 3) that are generally congruent with F_2 fold systems of the Kearsarge-central Maine sequence (Osberg et al., 1985). In the area of most extensive migmatization, mesoscopic folds are strongly overturned

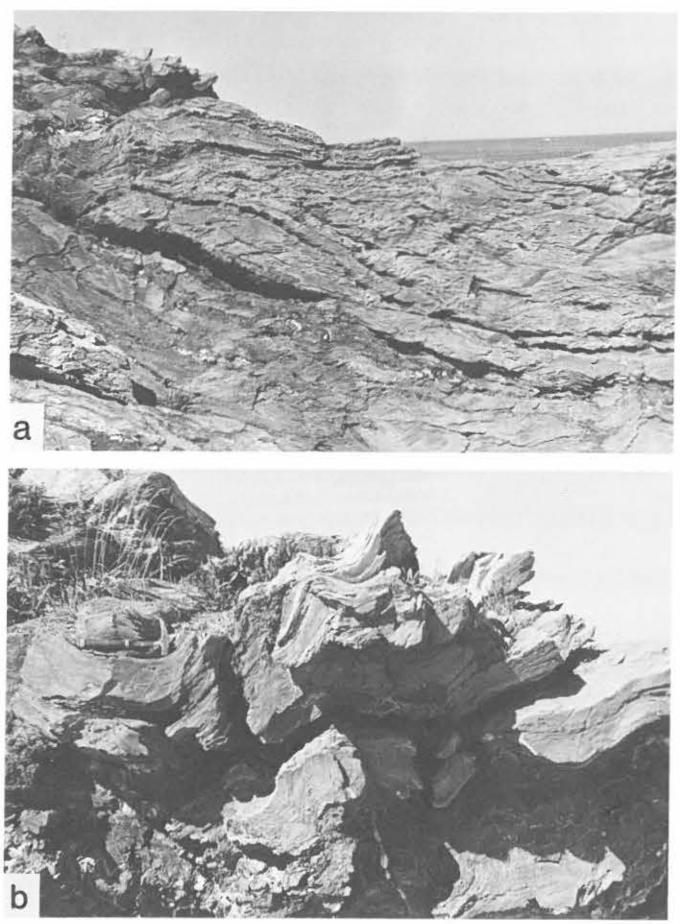

Figure 5. Multiple deformation of the Cape Elizabeth Formation, Two Lights State Park, Cape Elizabeth. (a) Recumbent F_1 fold. (b) Slightly overturned F_2 folds that refold bedding and spaced cleavage of F_1 folds.

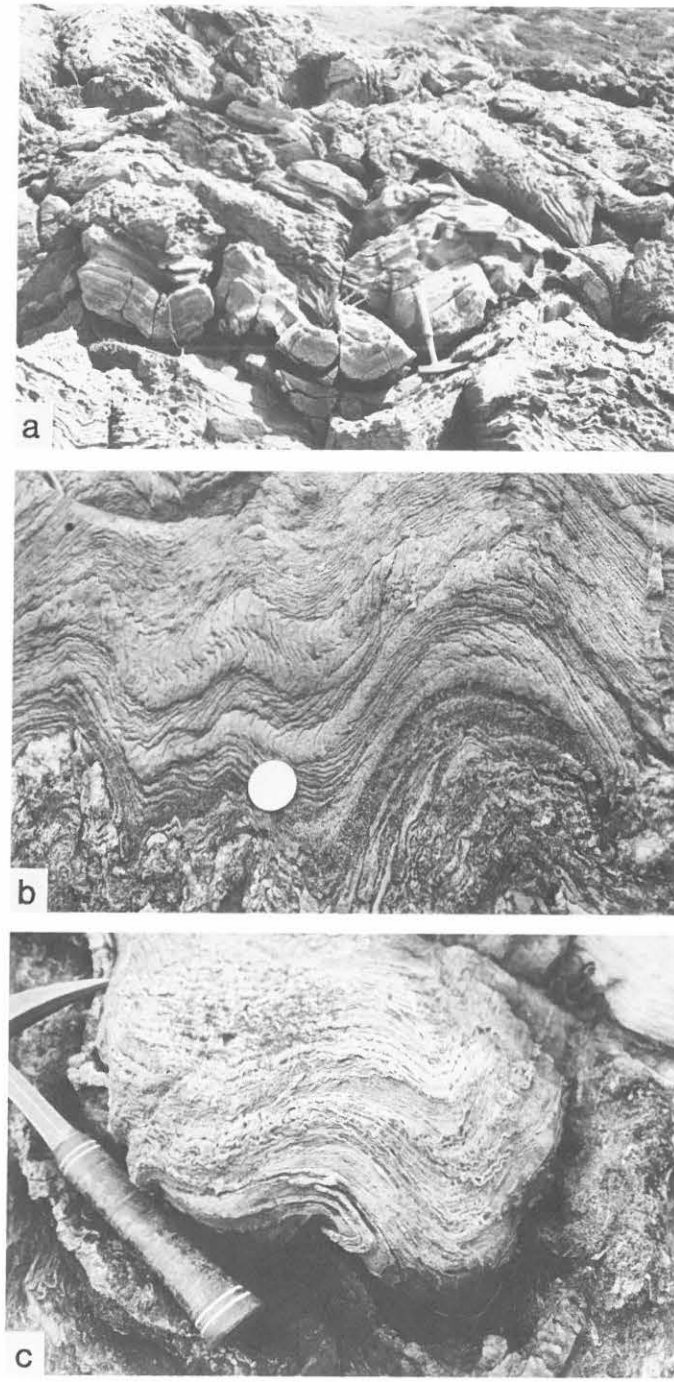

Figure 6. Multiple deformation of the Cape Elizabeth Formation, Small Point, Phippsburg. (a) Mesoscopic-scale recumbent fold (F_1) of calc-silicate bed refolded by relatively open upright F_2 folds. (b) Small-scale open upright F_2 folds. Spaced cleavage (originally parallel to the axial planes of F_1 recumbent folds) has been intensely and recumbently folded by an intermediate deformation (designated F_{1A}) seen only at this locality. Pressure solution has resulted in the removal of quartz in, and hence extreme attenuation of, the limbs of F_{1A}. At a distance the relatively thickened hinges of F_{1A} folds have the appearance of thin psammitic beds. Note dime for scale. (c) Multiply folded quartz vein in metapelite shows two stages of recumbent folding (F_1 and F_{1A}) openly refolded by F_2.

to recumbent or reclined (Fig. 7). Most commonly, these folds deform the migmatite stringers and thin concordant pegmatite lenses suggesting that rather than being older F_1 structures, they are younger structures related possibly to local gravity flattening during migmatization. These folds are very similar to late F_3 folds developed around the northeastern margin of the Sebago batholith during its emplacement (Lux and Guidotti, 1985) and around the northeast edge of the Mooselookmeguntic pluton in the Rangeley area for which Moench and Zartman (1976) suggested an origin by gravitational flattening due to heating as the pluton was emplaced. Some of the reclined and recumbent folds of the Falmouth-Brunswick sequence are strongly compressed and may represent older F_2 or possibly F_1 folds flattened during migmatization.

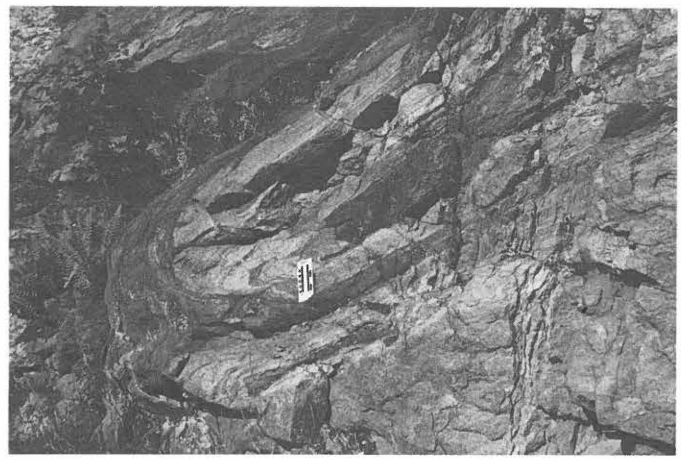

Figure 7. Recumbent F_3 folds of the Mount Ararat member of the Cushing Formation (Falmouth-Brunswick sequence), I-95 roadcut near the Bowdoinham exit.

Faults

Major faults in the study area are shown on Figure 3. They include normal and strike-slip faults associated with the Norumbega fault zone, and major thrust faults that juxtapose rocks of contrasting depositional environments against one another.

The major thrust faults include the Hackmatack Pond and Beech Pond faults, (Pankiwskyj, 1978; Newberg, 1985), the Boothbay thrust here named for its exposure in two windows in the Boothbay-Bristol area, and the St. George fault (Bickel, 1976). The Beech Pond and Boothbay thrusts are represented as distinct reflectors on the seismic reflection survey across the Kearsarge-central Maine synclinorium and coastal lithotectonic belt (Stewart et al., 1986). All dip gently to the northwest. The Hackmatack Pond fault brings the Siluro-Ordovician Vassalboro Formation, the lowest unit of the Kearsarge-central Maine sequence, eastward over the Late Proterozoic-Ordovician rocks of the Falmouth-Brunswick sequence. The Beech Pond fault brings the Falmouth-Brunswick sequence eastward over the Late Proterozoic-Ordovician Cape Elizabeth Formation of the Casco Bay Group. The Boothbay thrust brings the Late Proterozoic-Ordovician units of the Casco Bay Group, including the Sebascodegan formation, eastward over rusty schist and migmatite of the Ordovician(?) age Cross River formation. This fault surfaces in two antiforms only in the Boothbay-Bristol area and is cut out by the St. George fault. The St. George fault thrusts the Penobscot and Benner Hill sequences westward over the the Sebascodegan formation and the Bucksport Formation north of the the study area.

The Norumbega fault system is represented in the study area by the Cape Elizabeth, Flying Point, and Nonesuch River faults. These are post-metamorphic faults, the major movement on which, however, predates intrusion of the Devonian-Mississippian Lyman, Biddeford and Saco plutons. The Nonesuch River fault is traced from its junction with the Flying Point fault southwest of Portland to the Maine-New Hampshire border near Rochester, New Hampshire where it can be traced into the Campbell Hill fault of Lyons et al. (1982). The Nonesuch River fault is marked by the straight course of the upper Nonesuch River, silicified zones with drusy quartz boxwork, offset metamorphic isograds, and local contortion of schistosity of the Cape Elizabeth Formation. Hussey and Newberg (1978) suggest right-lateral strike-slip movement of up to 30 km. Movement may, however, have been left-lateral as suggested below for the Flying Point and Cape Elizabeth faults. This movement occurred prior to the intrusion of the Saco and Lyman plutons in Mississippian time inasmuch as these plutons are not significantly offset along the faults. Renewed minor dip-slip movement after pluton emplacement is suggested by a drainage and topographic lineament that does not offset the pluton-wall rock contacts. This lineament extends at least 8 km northeast of where the Flying Point fault meets the Nonesuch River fault.

The Flying Point fault extends from its junction with the Nonesuch River fault northeast through Flying Point in Freeport, to the Bowdoinham area where it merges with the Cape Elizabeth fault to form the Norumbega fault. It separates rocks of the Vassalboro Formation and the Falmouth-Brunswick sequence on the west from the Saco-Harpswell sequence on the east. It forms the boundary between high-grade migmatized members of the Cushing Formation of the Falmouth-Brunswick sequence on the west and significantly lower-grade units of the upper part of the Casco Bay Group on the east. This is traceable as far north as the southern end of Merrymeeting Bay. The fault offsets the migmatization front by approximately 40 km suggesting a major component of left-lateral strike-slip movement. The amount is unknown due to uncertainties as to the dip of the migmatite front. If very gentle, the major movement may actually be dip-slip normal, with the east side of the fault down-dropped. Splays of the fault are exposed at Johns Point on Flying Point, Freeport. Here rocks of the Cushing Formation of both sequences have been brecciated, and the foliation contorted and cut by numerous slickensided surfaces. The latest movement suggested by the slickensides is dip-slip normal.

The Cape Elizabeth fault has been traced by Hussey (1971a,b)

and Newberg (1981) from the Old Orchard Beach area to the Dresden Mills area where it merges with the Flying Point fault. Pankiwskyj (1978) traces a single fault from there to the northeast into the Norumbega fault of Stewart and Wones (1974) in the Bangor area. Near its southern end, along the shore in Cape Elizabeth, the Cape Elizabeth fault is actually a kilometer-wide zone of high-angle fault slices, locally filled with gouge and breccia. The principal break is marked by a series of 5 to 10 m wide pods of milky quartz traceable for about 2 km. In the area of Harpswell Neck, Harpswell, the contact between the Cape Elizabeth and Cushing Formations and the sillimanite/staurolite isograd are offset about 4 to 5 km suggesting left-lateral strike-slip movement, but as with the Flying Point fault, this may be significantly less, and possibly mostly dip-slip, if the contact and isograd surfaces are both very gentle in overall dip.

Numerous normal faults have been mapped in the study area, the larger of which are shown in Figure 3. They are of three general trends: 1) north to northeast, parallel to regional strike of the the metasedimentary rocks; 2) northwest; and 3) east-northeast. Faults of the latter two directions are generally traceable for less than a kilometer, and result in offsets of a few tens of meters. Normal faults parallel to regional strike may involve dip-slip movement of a few hundred meters. Among the more prominent of these north-northeast-trending faults are the Phippsburg, South Portland, and Back River faults (Fig. 3) and the boundary fault on the southeast side of the Macworth outcrop belt. The existence of the Back River fault is clearly demonstrated by the offset of a thin amphibolite lens within the Cape Elizabeth Formation and a granite orthogneiss, and by the retrograding of biotite to chlorite in exposures of the Cape Elizabeth Formation along the trace of the fault.

Small, unmapped faults are numerous throughout the region. Along the shore, these frequently control small-scale topographic indents of the shoreline. They are commonly marked by thin, very rusty-weathering gouge and breccia zones, the susceptibility of which to wave erosion and frost action produces the indentations.

METAMORPHISM

The stratified rocks of the the study area have been metamorphosed in a low-pressure (Buchan-type) facies series with grades of metamorphism ranging from low greenschist facies to upper amphibolite facies. Pelitic rocks are characterized at intermediate grade by the presence of andalusite, cordierite, and staurolite. At lower grades, the biotite zone is only poorly developed, biotite appearing in abundance essentially with the appearance of almandine garnet. At the highest grades of metamorphism, the pelitic rocks are strongly migmatized with a mineral assemblage characteristic of the sillimanite-K-feldspar grade of metamorphism. To the north of the study area in lenses of metapelite associated with the Falmouth-Brunswick sequence, Pankiwskyj (1978) reports the presence of microscopic kyanite, suggesting transition to a medium-pressure facies series in that area. Kyanite also occurs in rocks of the Kearsarge-central Maine sequence just southeast of Sebago Lake (Thomson and Guidotti, 1986).

Pelitic rocks of the Small Point area, now at andalusite to sillimanite grades of metamorphism, preserve 2 to 4 cm long pseudomorphs of muscovite after chiastolite(?) (Fig. 8). This suggests an earlier, intermediate grade, low-pressure facies series metamorphism. The present assemblage includes poikiloblastic fresh andalusite up to 5 cm in diameter, twinned staurolite, biotite porphyroblasts, minor cordierite, and sillimanite in small quantities intergrown with the biotite, in addition to the muscovite pseudomorphs.

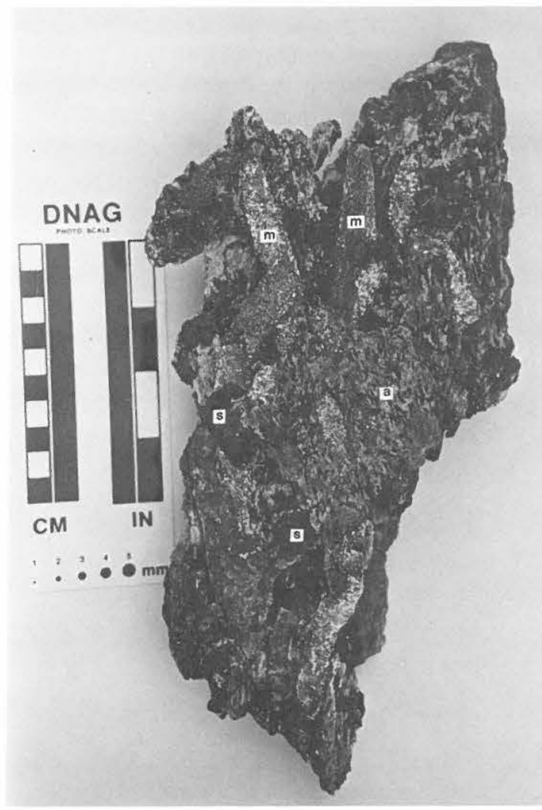

Figure 8. Multiple metamorphism of metapelite, Cape Elizabeth Formation, Small Point, Phippsburg. m: pseudomorphs of muscovite after first generation chiastolite; (a) second generation poikiloblastic pink andalusite with biotite inclusions; s: second generation staurolite porphyroblasts.

Retrograde metamorphism has variably affected the entire study area, and is expressed in the partial to locally complete chloritization of biotite, garnet, and staurolite, and the replacement of chiastolite by muscovite. Chloritization is particularly strong in the quartzo-feldspathic rocks of the Cushing Formation of both sequences, and the Cape Elizabeth Formation adjacent to the Flying Point, Back River, and Cape Elizabeth faults.

PLUTONISM

Intrusive rocks constitute about 25% of the outcrop in the study area. Plutons range in size from the Sebago batholith, areally the largest in the state of Maine, to the small elongate to irregular shaped plutons of the Phippsburg-Georgetown area (Fig. 3). Most are calc-alkaline biotite granite and two-mica granite. Two small quartz diorite plutons are present in Georgetown. Diorite forms the Saco pluton in the Saco area and a small unnamed pluton at the south end of the Sebago batholith northwest of Portland. Minor intrusives include granitic pegmatites which are very common in the areas of extensive migmatization, and diabase or basalt dikes. Pegmatites occur as irregular masses, lenses, and dikes. Some are folded and have a moderate foliation; others are massive. Exotic minerals (e.g., gemmy tourmaline, spodumene, purple apatite, rose quartz, etc.) characteristic of pegmatites in the central Maine area are lacking in most of the pegmatites of the Topsham-Georgetown area, notable exceptions being the large pegmatites of the Consolidated Feldspar Company quarry in Georgetown and the Fisher quarry in Topsham (Rand, 1957). The diabase and basalt form post-metamorphic dikes and sills up to 30 meters in thickness. They are ubiquitous throughout the area, although not abundant.

A few premetamorphic dikes have been mapped. They are mostly of felsic to intermediate composition, are unfolded, but are metamorphosed in harmony with the country rock. One dike consisting of light green, buff-weathering, calcite-green mica granofels, possibly representing an altered ultrabasic, intrudes the Cape Elizabeth Formation at the southern end of Bailey Island in Harpswell. Conformable lenses of chlorite-hornblende schist and occasionally anthophyllite-cummingtonite amphibolite occur sporadically in the Cape Elizabeth Formation. They are characteristically boudined and probably represent basic to ultrabasic sills intruded prior to both metamorphism and deformation.

TIME OF DEFORMATION, METAMORPHISM, AND PLUTONISM

Plutonism

Intrusion of plutons (other than the minor premetamorphic sills and dikes and post-tectonic basic dikes and sills) occurred from Early Devonian to Pennsylvanian time. Radiometric ages reported by Gaudette et al. (1982) and Aleinikoff et al. (1985) (Table 2) indicate intrusion of the Biddeford pluton (Fig. 3) during Middle Mississippian time, the Lyman pluton and Sebago batholith during late Mississippian time, and the Saco pluton possibly during the Pennsylvanian. The Pennsylvanian age of the Saco pluton is unusual because of the very strong pervasive lineation and foliation throughout the pluton and the metamorphic alteration of the primary igneous mineralogy. The pluton had earlier been interpreted to be a syntectonic premetamorphic Acadian intrusive (Hussey, 1985). The Webhannet pluton, which intrudes the Merrimack Group just to the south of the study area, was intruded during Early Devonian time. Pegmatites of the Falmouth-Yarmouth area were intruded during Late Devonian time. Granite and granodiorite plutons in the Augusta area north of the study area give radiometric ages indicating emplacement during Middle Devonian time (Dallmeyer and Van Breeman, 1981). McHone and Trygstad (1982) report a Middle Jurassic K/Ar age for a dolerite dike in Harpswell (Table 2). They indicate that similar dikes in New England give ages ranging from Late Triassic to Cretaceous.

Thrust-Faulting

The Hackmatack Pond fault which separates the Late Ordovician-Early Devonian Kearsarge-central Maine sequence from the Late Proterozoic-Ordovician Falmouth-Brunswick sequence is a result of the Acadian orogeny in Early Devonian time. The Boothbay thrust and the St. George fault are both locked by the Late Devonian age Waldoboro pluton. The youngest sequence affected by the Boothbay thrust is the Cross River formation of probable Ordovician age. Thus, although it may have formed during the Acadian orogeny, the Boothbay thrust may be as old as Ordovician. The St. George fault just east of the study area offsets the Penobscot Formation of Ordovician(?) age, and probably the Ordovician Benner Hill sequence (although the actual relations are obscured by the Waldoboro pluton). The continuation of this fault to the northeast into eastern Maine cuts the Flume Ridge Formation of possible Siluro-Devonian age. The St. George fault would appear to be an Acadian structure. The Beech Pond fault, which thrusts the Late Proterozoic-Ordovician Falmouth-Brunswick sequence eastward over similar age units of the Saco-Harpswell sequence, is either of Acadian or older age.

Normal and Strike-Slip Faults

Major movement on strike-slip faults of the area occurred between Early Devonian and Late Mississippian time. Strike-slip movement predates the emplacement of Mississippian to Pennsylvanian age plutons. The Nonesuch River fault, which merges with the Flying Point fault (Fig. 3) is locked by the Lyman pluton. The Cape Elizabeth fault is locked by the Biddeford and Saco plutons. However, since the Nonesuch River and Flying Point faults involve movement of the Late Ordovician-Early Devonian Kearsarge-central Maine sequence against pre-Silurian rocks of the Falmouth-Brunswick and Saco-Harpswell sequences, movement must post-date the Early Devonian.

Movement on northwest- and east-northeast-trending cross faults is inferred to post-date development of Acadian structures. However, no critical data are available to refine more closely the age of this movement.

Folding

The Vassalboro Formation is the oldest unit of the conformable formations of the Kearsarge-central Maine sequence that

includes units as young as Early Devonian. Clearly, then, both F_1 and F_2 fold systems of the Kearsarge-central Maine sequence were produced during the Acadian orogeny in Early Devonian time. Times of deformation of the rocks of the Falmouth-Brunswick and Saco-Harpswell sequences are not critically constrained by existing field relations, stratigraphic and structural interpretation, or radiometric age determinations. If, as previously discussed, the Macworth Formation, which is the uppermost unit of the Casco Bay Group, is conformable or disconformable to the Kearsarge-central Maine sequence, then both F_1 and F_2 of the Casco Bay sequence correlate with F_1 and F_2, respectively, of the Kearsarge-central Maine sequence. They would thus be of Acadian age. If, on the other hand, the Macworth Formation correlates with the Berwick Formation, the age of the deformation would be Late Precambrian to Early Ordovician. The constraint here is the 473 Ma age of the Exeter pluton which Gaudette et al. (1984) regard as having been posttectonically intruded into the Merrimack Group.

Strongly overturned folds (F_3) of the Falmouth-Brunswick sequence developed in the areas of intensive migmatization are interpreted to be younger than the F_2 folds of adjoining sequences. They have a similar geometry to the northwest-trending late folds around the northeast side of the Sebago batholith for which Hussey et al. (1986) and Aleinikoff et al. (1985) suggest a Mississippian age. The latter folds are interpreted to have been formed synchronously with the migmatization of the Kearsarge-central Maine sequence and emplacement of the batholith. The ages of F_1 recumbent folds and F_2 upright folds of the Falmouth-Brunswick sequence are unclear. F_2 folds appear to be generally conformable with F_2 folds of the Kearsarge-central Maine sequence which are clearly effects of the Acadian orogeny in Early Devonian time. This suggests that F_2 folding of the Falmouth-Brunswick sequence is likewise related to the Acadian orogeny.

No evidence of regionally extensive Alleghenian deformation has been noted.

Metamorphism

Regional metamorphism of the Kearsarge-central Maine sequence was accomplished in several stages during the Acadian orogeny (Guidotti, 1985). Mississippian ages for metamorphic minerals of rocks in the area north and northeast of the Sebago batholith reported by Lux and Guidotti (1985) are probably best interpreted as the age of a relatively extensive thermal aureole around the Sebago batholith. In a similar but much narrower aureole in the Vassalboro Formation around the Lyman pluton, also of Mississippian age, the Vassalboro Formation has been recrystallized. In particular, a schistosity defined by biotite has been recrystallized parallel to the contacts with the pluton, and to axial planes of small-scale F_3 folds. Approximately 1 km away from the pluton, outside the aureole, the schistosity has its normal north-northeast trend parallel to axial planes of F_2 folds. These relations argue for an Acadian age for the regional metamorphism of the Kearsarge-central Maine sequence.

The age of metamorphism of the Saco-Harpswell sequence is very uncertain, and parallels the arguments for age of folding discussed above. If the Macworth Formation is part of the Vassalboro Formation, the age of metamorphism is Acadian. If it is correlated with the Berwick Formation of the Merrimack Group, then metamorphism would predate the intrusion of the Exeter pluton, and would be Early Ordovician or older. The Early Middle Ordovician Rb/Sr ages (Table 2) for the Cape Elizabeth and Cushing Formations are subject also to varying interpretations, consequently not giving clear-cut constraints for time of metamorphism. They may represent ages of deposition/diagenesis and volcanism (Brookins and Hussey, 1978), or they may represent ages either completely or partially reset by metamorphism (H. E. Gaudette, pers. commun., 1986).

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Definition, Problems, and Reinterpretation of Early Premetamorphic Faults in Western Maine and Northeastern New Hampshire

Robert H. Moench
U.S. Geological Survey
Denver, Colorado 80225

Kost A. Pankiwskyj
Department of Geology and Geophysics
University of Hawaii
Honolulu, Hawaii 96822

ABSTRACT

Faults that originated prior to tight Acadian folding and formation of slaty cleavage are fundamental structural elements of the northwest limb and axial zone of the Kearsarge-central Maine synclinorium. They occur mainly, but not entirely, southeast of the Silurian tectonic hinge that marks the northwest margin of the synclinorium's ancestral sedimentary basin. The faults, termed early premetamorphic faults, commonly excise more than 3 km of section. Because the faults are almost cryptic in outcrops and are difficult to recognize, they have been controversial since three were first described in 1970. The following major early faults and associated features are now recognized in western Maine and adjacent New Hampshire. (1) The Hill 2808 and Mahoosuc faults are interpreted as a single basin-margin, southeast-down normal fault. (2) The Barnjum fault is an inferred southeast-down normal fault contained southeast of the hinge within the Silurian-Devonian basin sequence, but postulated to flatten northwestward across the hinge where it dislodged the shoreward facies of these deposits approximately along their basal unconformities. The gabbroic Sugarloaf pluton is inferred to have been emplaced as a thin sheet along part of the gently dipping portion of the fault. Smaller gabbro bodies also occur along or near other early faults. (3) The Plumbago Mountain fault, also within the northwest limb of the synclinorium, is interpreted as a steeply dipping, southeast-down normal fault near its northeast end, but as a gently dipping rupture farther southwest, where it is sinuous in plan, owing to younger deformation. In the Mahoosuc Range, New Hampshire, the Plumbago Mountain fault outlines a large inferred window containing small, synclinal, inferred outliers. (4) The Blueberry Mountain and Winter Brook faults are inferred to mark the margins and sole of the Rumford allochthon, a 30 $\times$ 70 km feature mapped in the axial zone of the Kearsarge-central Maine synclinorium. Within the allochthon is the Bald Mountain fault, interpreted as an internal rupture. The faults cut early folds, but, on the basis of geometric relations between fold plunges and fault displacements seen near the ends of the Hill 2808 and Barnjum faults, early faulting and folding are interpreted to have been contemporaneous.

On the basis of the distribution of inferred outliers and windows, we interpret the Plumbago Mountain, Blueberry Mountain, and Winter Brook faults to delineate younger-over-older features. An older-over-younger thrust interpretation is not ruled out, however, because the windows of our interpretation can be turned "inside out" to form outliers without changing map patterns. The alternate thrust interpretation requires probably several tens of kilometers of horizontal transport that should be expressed by the juxtaposition of different sedimentary facies, but no major facies juxtapositions are seen. Our preferred younger-over-older interpretation is satisfied by relatively small horizontal displacements.

The early faults and folds are postulated to have formed by giant-scale slumping into the ancestral sedimentary basin. On the basis of abrupt thickening of Silurian units on the southeast sides of the Hill 2808 and Barnjum faults, slump-faulting and folding probably began in Silurian time; it culminated, however, after deposition of the Lower Devonian Seboomook Formation but prior to Acadian compression. Culmination might have been triggered by emplacement of gabbroic magma along or near the faults.

INTRODUCTION

During his 1961-65 mapping in the Phillips, Rangeley, and Rumford 15-minute quadrangles, Moench (1970, 1971, 1973) delineated three major but almost cryptic longitudinal premetamorphic faults, shown as the Hill 2808, Barnjum, and Blueberry Mountain early premetamorphic faults on Figure 1a. Within the area that was mapped at that time, Moench (1970) showed that the Hill 2808 fault, together with a major hanging-wall syncline and an anticline farther southeast, defines a geometrically related fault-fold unit. Because the northeast end of the Hill 2808 fault is exposed within the mapped area, it was possible to demonstrate that displacement increases southwestward from zero to at least 3 km in the direction of plunge of the hanging-wall syncline and in the direction of increasing structural relief in the syncline-anticline pair (Moench, 1970, Figs. 2, 10; 1973, Fig. 4). A similar relationship is exposed in relation to the Barnjum fault. On this basis Moench (1970) inferred that a genetic relationship exists between faulting and folding. He further postulated that the entire major fault-fold-cleavage pattern originated by slumping into the ancestral sedimentary basin, and that slumping was accompanied by tectonic dewatering (Moench, 1970, Fig. 14, p. 1488-1492; 1973, Fig. 5, p. 336-339).

Since then, mapping has been completed or revised at 1:62,500 through a much larger area (Osberg et al., 1985; Moench et al., 1982; Moench and Pankiwskyj, 1988; Moench, 1984), and three additional premetamorphic early faults have been mapped, shown as the Mahoosuc, Plumbago Mountain, and Winter Brook early premetamorphic faults on Figure 1a. As a result, we now have a much-improved grasp of how the faults relate to the geologic framework and paleotectonic setting of a large region, and a better basis for addressing controversial aspects of the faults. Some of our colleagues have expressed disbelief of some (or all) early faults, on the grounds that the stratigraphic units on opposite sides were incorrectly identified. Others have challenged the inferred younger-over-older relationships and the postulated down-to-basin slump origins.

Finally, since the early 1970's, John Maxwell's dewatering hypothesis for the origin of slaty cleavage (Maxwell, 1962), applied by Moench (1966, 1970, 1973) to western Maine, has been largely supplanted by the multi-authored pressure-solution hypothesis (see discussions and examples in Borradaille et al., 1982). Accordingly, whereas Moench previously invoked a model of basin-controlled slumping and tectonic dewatering *without* major orogenic compression, he now favors a mechanism of pre-Acadian, basin-controlled, slump-faulting and open-folding, but *followed* by Acadian compression, which deformed the faults, greatly tightened and amplified the folds, and produced the slaty cleavage.

Readers are referred to Moench et al. (1987) and Moench (in press) for discussions of the newly-recognized Piermont allochthon of northwestern New Hampshire and adjacent areas. The Piermont allochthon is pertinent to this paper because it is interpreted as an Acadian thrust sheet whose contents accumulated on a postulated Silurian platform along and immediately northwest of the Silurian tectonic hinge of Figure 1b. The platform is proposed to have been compressed almost out of existence by Acadian compression.

The focus of this paper is on the stratigraphic and structural features that bear on the interpretation of the mapped early faults. The paper is an outgrowth of published mapping, the cited early papers, several talks, and a guide article that was prepared for a Maine Geological Society field trip that was run on July 28, 1985. Figure 1 is simplified from the maps of Moench et al. (1982), Moench and Pankiwskyj (1988), and Moench (ed.), (1984). Readers are referred to these published maps for sources of detailed information and mapping credits. We are grateful to Dwight Bradley, Bob Marvinney, and Phil Osberg for their constructive reviews.

STRATIGRAPHY

The area of Figure 1 is underlain by 7-8 km of Lower Devonian and Silurian metasedimentary strata, and by 5 km or more of Cambrian? to Ordovician metasedimentary and metavolcanic rocks. These rocks have been divided into many formations which, for simplicity, are grouped on Figure 1 into six sequences according to age. The thin Silurian shelf and shoreline facies are singled out to illustrate part of the southeastern margin of the Silurian source area. Brief descriptions of units are provided in the map explanation. A synthesis of sedimentation is provided in the pamphlet that accompanies the geologic map of western interior Maine (Moench and Pankiwskyj, 1988). As shown on the index to Figure 1, the area spans most of the Kearsarge-central Maine synclinorium, but it overlaps the north end of the Bronson Hill anticlinorium and the southeast side of the Boundary Mountains anticlinorium.

The Silurian tectonic hinge (Fig. 1b) is the approximate axis of abrupt northwestward thinning from about 5 km of Silurian clastic deposits of a conformable basin sequence southeast of the hinge to much thinner shelf and shoreline deposits to the northwest (Moench and Boudette, 1987; Hatch et al., 1983).

Northwest of the hinge, lower contacts change from conformity to unconformity. Whereas only a Silurian graptolite fauna is known in the basin sequence (Pankiwskyj et al., 1976), Silurian deposits northwest of the hinge locally contain a varied, relatively shallow-water shelly fauna (Boucot and Thompson, 1963; Boucot and Heath, 1969; Moench and Boudette, 1970). As already noted in the introduction, Moench (in press) has proposed that the interval between the hinge and the shelf-shore facies was a wide Silurian platform, with sub-basins.

The Silurian tectonic hinge is expressed most conspicuously by the Lower Silurian Rangeley Formation, which thins from about 3 km south of Rangeley village to about 1/2 km in the Kennebago Lake area. Upper Silurian shelf or shoreline facies with disconformable or unconformable lower contacts northwest of the hinge are represented by the Fitch Formation of western New Hampshire (Billings, 1956), the Hardwood Mountain Formation of the Moose River synclinorium, northwestern Maine (Boucot and Heath, 1969), and The Forks Formation (upper Silurian?) at the northeast corner of Figure 1a (Marvinney, 1984). These units are approximately equivalent to the Madrid Formation of the basin sequence. Lower Silurian shelf facies are represented by fossiliferous quartz conglomerate and calc-silicate rock of the Rangeley Formation (basal member C) within the northeastern end of the Kennebago outlier, where the lower contact is faulted, and outside the outlier just southeast of Kennebago Lake, where underlying Silurian deposits are conformable but thin (Moench and Boudette, 1970, 1987). Lower Silurian shoreline facies are represented by massive, lenticular-bedded orthoquartzite and quartz conglomerate assigned to the Clough Quartzite (Moench and Boudette, 1987) along the northwest margin of the Kennebago outlier, mapped both inside and outside the allochthon (Fig. 1a,b). Exposures of Clough Quartzite near the northwest side of the allochthon are underlain by a thin unit of pelitic rocks, probably member B of the Rangeley Formation; northwest of the allochthon these deposits rest unconformably on pre-Silurian rocks. The lower contact of the type Clough Quartzite in western New Hampshire is an unconformity (Billings, 1956).

In areas of low- to high-grade metamorphism, but not including the migmatitic gneisses (Fig. 1b), bedding is well preserved and provides the basis for defining formation characteristics and younging directions, and the stratigraphic omissions and truncations produced by the early faults. The actual faults, however, are very inconspicuous in outcrop (Moench, 1970, Figs. 7-9, 12, 13) and can be missed if careful attention is not paid to the differences between individual rocks and sequences of rocks on both sides. It is emphasized that widely separate parts of the stratigraphic section may be remarkably similar (for example, pelitic parts of the Rangeley and Carrabassett Formations) and can be misidentified if the sequence from one unit to another, if exposed, is not recognized. This is a particularly difficult but not unsolvable problem in areas of migmatitic gneiss. Although detailed bedding features and sedimentary evidence of younging directions are lost in areas of migmatitic gneiss, the large sequences can be seen and are identified accordingly on Figure 1a. (See Moench and Hildreth, 1976, and Moench and Pankiwskyj, 1988, for descriptions of the migmatitic gneisses.)

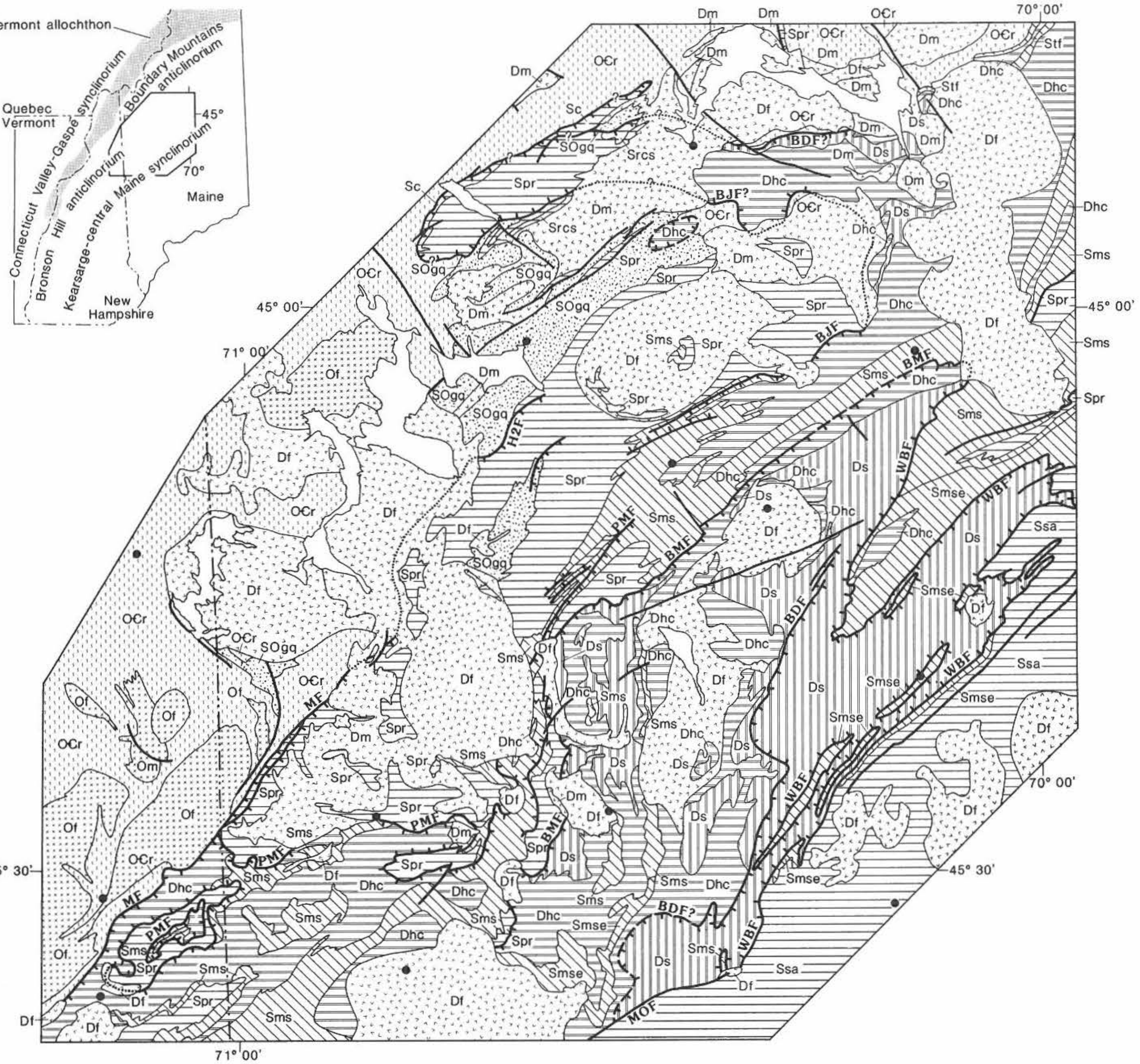
STRUCTURAL AND METAMORPHIC CHRONOLOGY

The complex structural and metamorphic pattern of western Maine is the product of events that produced the early faults and associated folds, followed by Acadian compression and subsequent plutonism. A chronology that accords with our interpretation of known relationships and available isotopic age data includes the following events:

1) Early faulting and associated open folding: Early faults and folds are inferred to be coeval; their origin was possibly accompanied by incipient development of slaty cleavage by tectonic dewatering, but no major metamorphism, with the possible exception of areas near gabbroic intrusions that were emplaced along early faults, particularly the Sugarloaf pluton along the Barnjum early fault. Earliest movements probably accompanied Silurian sedimentation and the latest movements occurred after deposition of the Lower Devonian Seboomook Formation. The age of the Sugarloaf pluton, approximately dated at 406 ± 12 Ma (data in Moench, 1984; K-Ar biotite method), suggests that latest movement on the Barnjum fault occurred no later than about 405 Ma.

2) Acadian compression: This compression caused tight folding, formation of slaty cleavage, and greenschist-facies or lower-rank metamorphism (M_1). This event greatly tightened and amplified the early folds, produced new folds, and deformed the early faults. This probably occurred between about 405 and 400 Ma, as shown by truncation of Acadian folds and cleavage by the Lexington batholith, dated at 399 Ma (Gaudette and Boone, 1985; Gaudette, pers. commun., 1985; Rb-Sr whole-rock (399 ± 6 Ma) and mineral (399 ± 3 Ma) isochrons).

3) Localized, overlapping multiple deformation and metamorphism produced by post-Acadian Devonian to Carboniferous plutonism: This produced the metamorphic zoning and many of the mapped structural "twists and turns" shown on Figure 1. The tie between the origin of superposed folds, late schistosity, metamorphic zoning, and the emplacement and cooling history of the Mooselookmeguntic batholith and other plutons has been demonstrated by Moench and Zartman (1976) on the basis of structural and petrographic observations. The metamorphic zoning of Holdaway et al. (1982) is similar in principle, but different in application. Each new thermal event can be labeled according to M_{1+n}, where 1 is the regional Acadian event, and n is each successive plutonic emplacement; chronology cannot be completed until all major plutons are dated. The oldest recognized metamorphism ($M_{2?}$) and deformation was produced by the 400 Ma Lexington batholith and possibly the Redington and Phillips plutons. An intermediate dynamothermal event was produced by two-mica granite of the Mooselookmeguntic batholith, dated at 371 ± 6 Ma (Moench and Zartman, 1976; Rb-Sr whole-rock; new constants). The youngest known high-rank dynamothermal pattern is related to the Carboniferous Sebago batholith in southwestern Maine (south of area of Figure



PLUTONIC ROCKS

Dm Df — Devonian rocks – Dm, mafic; Df, felsic and intermediate

Om Of — Ordovician rocks – Om, mafic; Of, felsic

STRATIFIED METAMORPHIC ROCKS

Lower Devonian

Ds — Younger sequence – Seboomook Fm: mainly gray pelitic schist and graded metasandstone. Lower contact conformable.

Dhc — Older sequence – Hildreths, Carrabassett, Littleton Fms. Hildreths: volcaniclastic metagraywacke, calc-silicate rock; Carrabassett: massive gray pelitic schist; Littleton: gray pelitic schist, metasandstone. Lower contacts conformable.

Silurian and Silurian?

Stf Sms Smse — Younger sequence – Stf, The Forks Fm: calcareous sandstone, conglomerate. Lower contact unconformable. Sms, Madrid and Smalls Falls Fms. Madrid: metasandstone, gray schist, basal calc-silicate rock; Smalls Falls: black sulfidic schist, quartzite, upper calc-silicate rock. Smse, eastern facies, Madrid and Smalls Falls: lacks calc-silicate rock. Lower contacts of Sms, Smse conformable.

Sc Srcs Spr Ssa — Older sequence – Sc, Clough Quartzite: orthoquartzite and quartz conglomerate. Lower contact unconformable. Srcs, Rangeley Fm, mbr C shelf facies: slate, sandstone, fossiliferous quartz conglomerate and calc-silicate rock. Spr, Perry Mountain and Rangeley Fms. Perry Mountain: interbedded pelitic schist, quartzite; Rangeley: gray sulfidic schist, metasandstone with quartz metaconglomerate (mbr C), quartz-rich polymictic conglomerate (mbr B), polymictic boulder to pebble conglomerate, meta-arkose (mbr A). Lower contacts conformable. Ssa, Sangerville Fm: green to gray pelitic schist, volcaniclastic metagraywacke, fine grained metaconglomerate, thinly bedded metalimestone. Lower contact conformable.

Lower Silurian? and Upper Ordovician?

SOgq — Transitional sequence – Greenvale Cove and Quimby Fms. Greenvale Cove: laminated feldspathic metasiltstone, metasandstone, calc-silicate rock; Quimby: gray to black sulfidic schist, volcaniclastic metagraywacke, lower felsic metavolcanic rocks. Contacts conformable.

Middle Ordovician, Lower Ordovician, Upper Cambrian?

O€r — Ammonoosuc Volcanics, Partridge and Dead River Fms. Ammonoosuc: silicic to basaltic metavolcanic and related metasedimentary and exhalative rocks. Lower contact possibly unconformable; Partridge: black sulfidic schist or slate, minor greenstone. Intertongues with Ammonoosuc; Dead River: thinly interbedded pelitic schist, quartzite.

Figure 1a. Simplified geologic map of western Maine and adjacent New Hampshire showing stratigraphic and plutonic units, and distribution of early premetamorphic faults. Modified from Moench and Pankiwskyj (1988) and Moench (ed.), (1984).

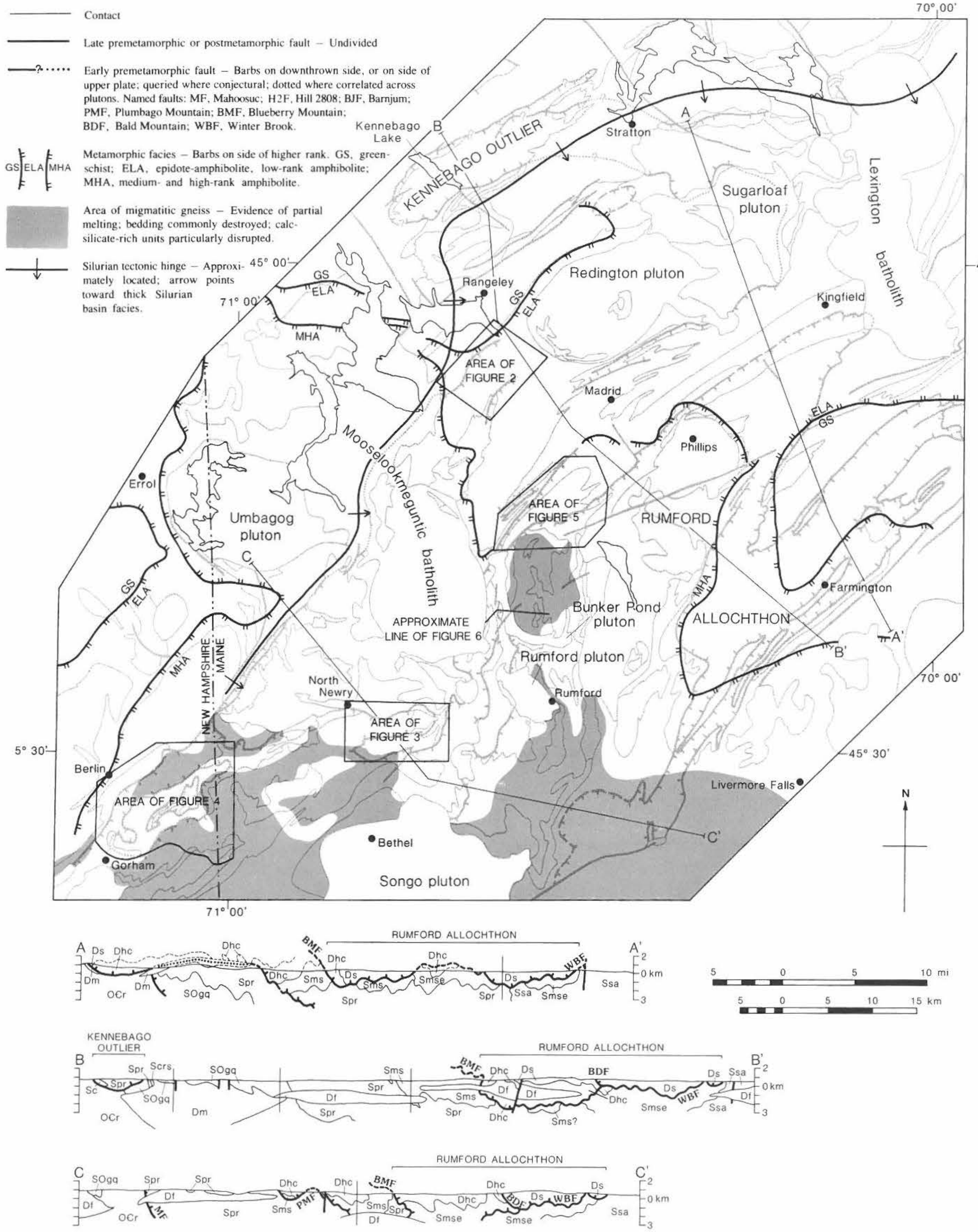


Figure 1b. Metamorphic isograds, pluton names, allochthons, and geographic names for same area as (a).

1), dated at about 325 Ma by two methods (Aleinikoff et al., 1985; Lux and Guidotti, 1985); this pattern probably overlaps the southern margin of the area of Figure 1.

HILL 2808 AND MAHOOSUC EARLY FAULTS

These features mark significant parts of the boundary between pre-Silurian and Siluro-Devonian rocks and are inferred to correlate across the Mooselookmeguntic batholith. This correlation is based on the structural and stratigraphic continuity seen in roof pendants of the batholith. The two faults have somewhat different characteristics, however, which are here interpreted in terms of different depth environments.

The Hill 2808 early fault dips 50-68 degrees southeast, as measured at four outcrops (Moench, 1971), and younger rocks on the southeast are faulted against older rocks on the northwest. These relationships indicate that it is a normal fault. The fault ends at the southeast corner of Rangeley Lake, and it is truncated on the southwest by the Mooselookmeguntic batholith.

The location of the important sequences that demonstrate the existence of the fault are shown in Figure 2 (areas 1a and 1b); detailed descriptions are provided by Moench (1970, p. 1475-1482; 1971). Area 1a is in the garnet and retrograded staurolite metamorphic zones. The sequence includes the Quimby Formation and type localities of the Greenvale Cove and Rangeley Formations, along and near Maine Highway 4, 1.5 to 3.5 miles south of Rangeley (Moench and Boudette, 1987). Here, an apparently unfaulted, southeast-younging sequence approximately 3-3.5 km thick is exposed across the northeastern terminus of the Hill 2808 fault. The rocks include the upper part of the Quimby Formation (1 km thick), the Greenvale Cove Formation (200 m thick), and three members of the Rangeley Formation (total about 3 km thick).

Structurally, the sequence at area 1a (Fig. 2) constitutes the northwest limb of the Mountain Pond syncline; this limb is truncated farther southwest by the fault. Near the arrow within the northwest limb of the syncline (Fig. 2), minor folds plunge 30-70 degrees to the southwest (Moench, 1971). This is the area of most abrupt loss of section toward the southwest along the fault. Farther southwest, plunges become gentler and loss of section is more gradual; here it takes place mainly by truncation of northeast-plunging, folded, pre-Rangeley strata northwest of the fault. Near the arrow at the nose of the Brimstone Mountain anticline, due southeast of the steepest southwest plunges of the Mountain Pond syncline, minor folds plunge 10-30 degrees northeast. Again, plunges become gentler farther southwest.

Area 1b (Fig. 2) lies astride the sillimanite isograd of the Mooselookmeguntic batholith. The important stratigraphic sequence is along the crest and steep eastern slope of Hill 2808 (Moench, 1971). Here, the uppermost beds of the Rangeley Formation (member C) are faulted against the uppermost beds of the Quimby Formation; approximately 3.5 km of strata are missing. The Quimby Formation, exposed in a narrow, northwest-younging belt just northwest of the fault, is overlain to the north-

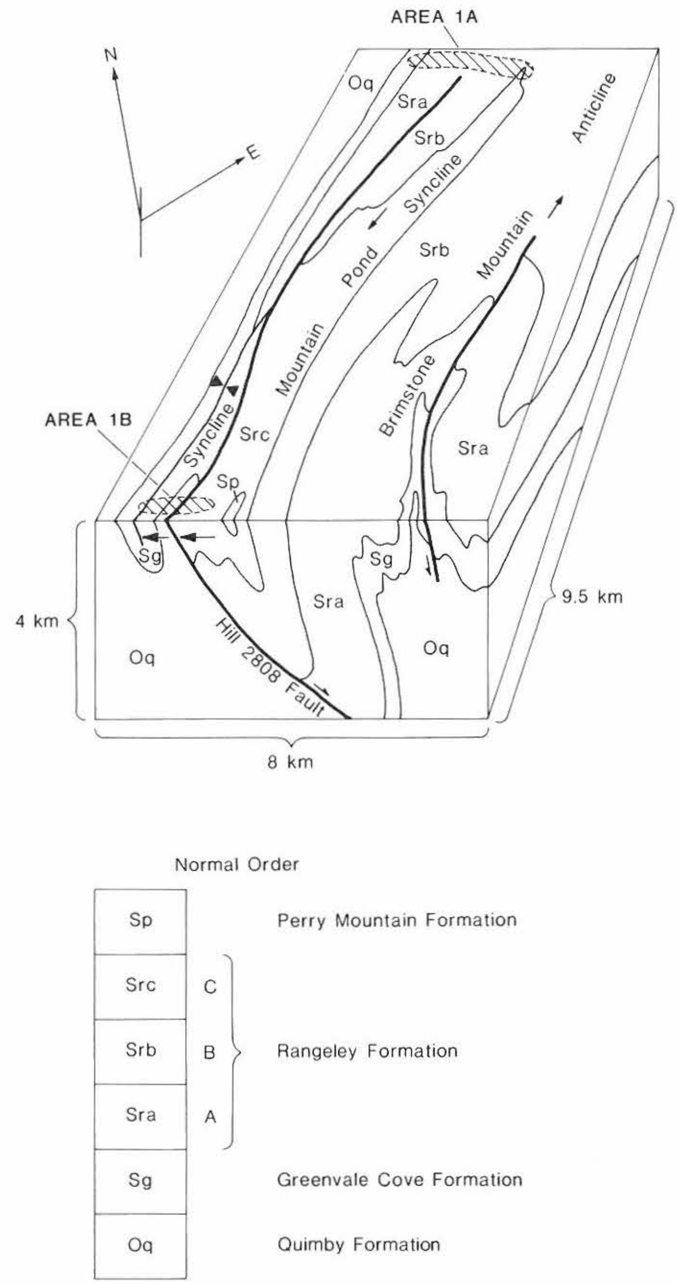


Figure 2. Block diagram showing inferred relationships between the Hill 2808 fault, the Mountain Pond syncline, and the Brimstone Mountain anticline. Restored after removal of the Mooselookmeguntic batholith. Arrows on cross-sectional view show topping direction; arrows on map surface show plunge direction of minor folds. Location shown on Fig. 1b.

west by the Greenvale Cove Formation, which is exposed in a syncline that is truncated by the fault to the northeast. Although graded beds top southeast within two or three meters of the fault, the larger sequences on both sides of the fault (about 400 m) top to the northwest, as indicated on the face of Figure 2.

As seen in outcrops in area 1b and thin sections, the fault is a sharp, pre-cleavage feature that shows evidence of sand-

stone disaggregation and mixing (Moench, 1970, Figs. 7-9, esp. 9b) but no conspicuous evidence of cataclasis. The wall rocks appear to have been poorly lithified during faulting.

The fact that the most abrupt southwestward loss of section along the fault coincides with the steepest southwest plunges within the syncline to the southeast and with the steepest northeast plunges within the anticline farther southeast is evidence that the fault and folds are temporally and dynamically related. The Mountain Pond syncline may be called a hanging-wall syncline, and the fault and both folds to the southeast may be called a fault-fold unit (Moench, 1973).

It is further noteworthy that member A of the Rangeley Formation, composed of metamorphosed sandstone and conglomerate, is much thicker (1.2 km) southeast of the Hill 2808 fault than it is anywhere to the northwest or to the southeast of the crest of the Brimstone Mountain anticline. By analogy with Gulf Coast structure (Moench, 1970, 1973), the fault may thus have begun to form in the Silurian as a growth fault. The Brimstone Mountain anticline is inferred to have begun to form as a rollover, by back-tilting toward the hanging wall of the presumably listric Hill 2808 normal fault. The Mountain Pond syncline may have formed by a combination of back-tilting on the northwest limb of the anticline and drag adjacent to the fault. During the Acadian, both folds were greatly tightened and overturned slightly to the northwest. If the fault originally dipped moderately, say 60 degrees, southeast, it would have been steepened by horizontal compression, but overturning would have had the opposite effect.

The Mahoosuc early fault juxtaposes the Rangeley, Perry Mountain, and Littleton Formations on the southeast against the Dead River and Partridge Formations, the Ammonoosuc Volcanics, and Ordovician granite (Fig. 1a). For a distance of about 13 km northeast of the state line, the Mahoosuc fault contains an inferred slice of extremely rodded polymictic, boulder- to pebble-metaconglomerate assigned to member A of the Rangeley Formation. The clasts, which include metamorphosed volcanic, sedimentary, and granitic rocks, and vein quartz, are round to sliver-like in plan view, but they are rodded almost vertically (consistent plunge about 80 degrees NE). The rodding might be Acadian, however, because less extreme conglomerate rodding of similar orientation is seen well away from known early faults.

Although the metaconglomerate might be interpreted to rest unconformably on pre-Silurian rocks, features seen at one outcrop indicate that its northwestern contact is a premetamorphic fault. The outcrop is in a brook at elevation 1860 ft, about 2,000 ft (600 m) S17°E of the top of Deer Hill in the Old Speck Mountain 15-minute quadrangle. Here, a bed of metaconglomerate about 3 m thick youngs northwest, as shown by normal grading, and is sharply juxtaposed at its top against rusty-weathering, black, sulfidic schist and metasiltstone of the Partridge Formation, which also youngs northwest. The contact is knife-sharp, shows no evidence of cataclasis, and is crossed by schistosity.

The following are other important characteristics of the Mahoosuc fault. 1) A zone as much as 1/2 km wide of sharply interlaminated feldspathic quartzite and pelitic schist extends along the southeast side of the metaconglomerate slice; granoblastic textures are seen in thin sections. This zone, interpreted as extremely flattened Rangeley and Perry Mountain Formations, might have originated by an unknown process during normal faulting; alternatively, it might be a product of buttressing of viscoplastic Silurian rocks against more competent pre-Silurian rocks during Acadian compression. 2) Several small bodies of metamorphosed gabbro, now amphibolite, are exposed along the fault or nearby within the hanging wall.

If the Hill 2808 and Mahoosuc early faults are correctly correlated across the Mooselookmeguntic batholith, they represent a single feature that extends at least 100 km along the boundary between pre-Silurian and Siluro-Devonian rocks in western Maine and northern New Hampshire. Noting that the fault lies close to or on the approximate trace of the Silurian tectonic hinge, Moench (in press) proposed that it is a major, basin-margin normal fault. As such, it was probably active in Silurian time; its latest movements, however, occurred after the early Devonian formation of the Plumbago Mountain early fault, which is truncated by the Mahoosuc fault. Under this interpretation, the Hill 2808 segment is inferred to represent a relatively shallow expression at the northeastern extremity of the fault. At the present level of exposure, the wall rocks of the Hill 2808 fault were semilithified. Farther southwest, deeper levels of penetration are represented where the fault cuts hard pre-Silurian granite. The rather straight trace of the entire length of the Mahoosuc-Hill 2808 fault indicates that it originally dipped rather steeply and penetrated to deeper levels than the Blueberry Mountain, Plumbago Mountain, and Winter Brook faults, whose traces are extremely sinuous.

BARNJUM EARLY FAULT, GABBROIC INTRUSIONS, AND KENNEBAGO OUTLIER

The Barnjum fault extends from about 3 km northwest of Madrid village at least to the south end of the gabbroic Sugarloaf pluton (Fig. 1a,b). In a manner that is analogous to the Hill 2808 fault and associated folds, the Barnjum fault is interpreted to be a normal fault, southeast side down. The southwest end of the fault is about 3 km northwest of Madrid village; displacement increases northeastward in the direction of plunge of a prominent hanging-wall syncline and is greatest at the south end of the Sugarloaf pluton, where the Carrabassett Formation is juxtaposed against the Perry Mountain Formation; the Smalls Falls and Madrid Formations have been excised. The fault cannot end at its point of maximum displacement, however, where two formations and unknown amounts of two others are missing. Because no evidence of displacement is seen on-strike immediately to the northeast, the Barnjum fault is inferred to project northward, as shown, across the Sugarloaf pluton. The Sugarloaf pluton, a rather thin (< 1 km) gently dipping sheet (Carnese, 1981; Carnese et al., 1982), intervenes between northeast-trending belts of Ordovician and Silurian rocks to the southwest and a hook-shaped belt of conformable Lower Devo-

nian formations to the northeast. Southeast of Stratton, parts of the Carrabassett, Hildreths, and Seboomook Formations exposed in the hook-shaped belt lie discordantly above the Dead River Formation. Here, all of the transitional and Silurian sequences are missing (Fig. 1a,b). Although Boone (1973, Fig. 7) has interpreted these relationships in terms of stratigraphic onlap across a regional unconformity, we prefer the fault interpretation because several different formations are juxtaposed on both sides of the boundary, or would be if the Sugarloaf pluton were removed.

We propose that the Sugarloaf pluton was emplaced along the flattened, northern extension of the Barnjum fault, as shown above ground in section A-A′ (Fig. 1b). Boone (1973) has shown that the Sugarloaf gabbro is older than the granitic Lexington batholith, in accord with the isotopic age data already cited. Permissively, the gabbro might have intruded while the Barnjum fault was active.

It is tempting to correlate the Barnjum fault with the lower contact of the Kennebago outlier of Silurian rocks (Fig. 1a,b). The outlier is an open, synclinal body that contains Early Silurian inferred shoreline deposits of the Clough Quartzite (Moench and Boudette, 1987, p. 275-277) along its north side, and fossiliferous Early Silurian shelf metalimestone near its eastern end. The Clough Quartzite, underlain by a thin unit of dark slate, also occurs in outliers, too small to show on Figure 1, just northwest of the Kennebago outlier. The Clough Quartzite of the Kennebago outlier is overlain by slate and metasandstone of the Rangeley (member C) and Perry Mountain Formations. Because the Clough Quartzite (and a body of Rangeley polymictic conglomerate exposed at the southwest end of the outlier) is exposed at or near the lower boundary of the outlier, this boundary can be interpreted as a major unconformity. The boundary, however, truncates almost at right angles across a steeply dipping, southeast-younging sequence that contains, immediately southeast of the outlier, conformable beds of the Greenvale Cove and Rangeley Formations. At this locality, the Rangeley Formation is thin (less than 500 m), but all three members are represented (Moench and Boudette, 1987, p. 275-277), including fossiliferous quartz conglomerate and metalimestone in member C.

Tentatively, we propose that the lower contact of the Kennebago outlier is a fault, because it is difficult to reconcile the close juxtaposition of rocks that lie above a right-angle unconformity with rocks of the same age that lie within a conformable sequence. On the assumption that the Kennebago outlier is allochthonous, we favor a model of southeastward sliding, rather than northwestward thrusting, in order to place the site of a major Silurian unconformity farther northwest from the known Silurian conformity.

We favor an interpretation in which the Barnjum early fault has an approximately cymoid shape in cross section, approximately as depicted on Figure 1b (section A-A′, and northwest end of section B-B′). According to this interpretation, north of the Redington and Sugarloaf plutons it is sub-horizontal, though broadly warped, and marks the sole of a widespread allochthon composed of gently folded Silurian and Lower Devonian units that were deposited perhaps several kilometers farther northwest. We infer that it dislodged these deposits along or near the basal unconformities, and that it passed southeastward into the thick basin sequence. If, contrary to our belief, the Kennebago outlier is autochthonous, the fault may have originated within the Devonian sequence north of the area of Figure 1 and sliced downward through thin Silurian deposits to the pre-Silurian sequence exposed below the fault southeast of Stratton.

Near the southeast sides of the Redington and Sugarloaf plutons, the fault is inferred to roll southeastward to a steep southeasterly dip; farther southeast it is inferred to flatten in depth and disappear with diminishing displacement within the thick basin sequence. Displacement also diminishes southwestward to zero west of Madrid village (Fig. 1a,b). According to our interpretation, the fault is a flat-lying detachment feature in northwestern areas, attached in the southeast, and originated much as a rug might slide down an incline, when nailed to the lower edge and one side of the incline. The fault was a locus for the intrusion of gabbroic magma which, speculatively, might have triggered movement.

The fault-fold relationships exposed along the segment of the Barnjum fault that lies between Madrid and the Sugarloaf pluton suggest that faulting began by a listric growth-fault mechanism, as proposed for the Hill 2808 fault and related folds. The fact that the thickest known deposits of the Silurian Smalls Falls Formation (composed of strongly sulfidic-graphitic, closed-basin deposits of types that are common in extensional basins) are exposed between the fault and Madrid village, and that the Smalls Falls Formation thins abruptly to the north across the fault, suggests that Barnjum faulting began during the deposition of the Smalls Falls Formation.

Osberg et al. (1985) depict the Barnjum fault southeast of the Redington and Sugarloaf plutons as a normal fault, southeast side down (section F-F′), and they show southeast transport on the fault between the Devonian and pre-Silurian sequence just east of Stratton (section E-E′), in approximate accord with our interpretation. Unlike the interpretations by Osberg et al. (1985), our model attempts to relate these features as a single surface of dislocation. We favor southeast transport by gravity sliding, as opposed to northwest-directed thrusting, because southeast-down normal displacement more easily explains the presence of younger rocks on the southeast side of the fault, where mapped south of the Redington pluton, as well as the relation between fault displacement and synclinal plunge in that area. A thrust model would require that the fault cut down-section in the direction of transport and cannot explain the plunge-displacement relationships.

PLUMBAGO MOUNTAIN EARLY FAULT

The northeast end of the Plumbago Mountain fault is mapped about 7 km southwest of Madrid village, where northwest-younging rocks of the Madrid and uppermost Smalls Falls Formations abut northwestward against southeast-younging basal

rocks of the Smalls Falls Formation (Moench, 1971); here, about 700 m of strata are missing. Displacement increases to the southwest, where the fault is mapped in and out of the septum between the Mooselookmeguntic batholith and the Bunker Pond pluton, and from there almost to Gorham, New Hampshire, for a total length of at least 70 km. Regionally, younger rocks on the southeast abut older rocks on the northwest, but this relationship is modified greatly by younger folding along the southwestern segment of the fault; there, structural relationships suggest that younger-over-older relationships occur vertically across a sub-horizontal dislocation. These relationships suggest that the fault originated as a southeast-down normal fault. It was probably listric, with a steep original dip near its northeast end and a gentle dip west of Rumford.

Outcrops of the Plumbago Mountain fault were found: 1) at an elevation of 2280 ft on the north shoulder of Plumbago Mountain, about 630 ft (250 m) west of the boundary between the Rumford and Old Speck Mountain quadrangles, Maine; 2) at an elevation of 1860 ft on the southeast side of an unnamed hill, 11,800 ft (3300 m) azimuth N60°W from the southeast corner of the Old Speck Mountain 15-minute quadrangle; and 3) at an elevation of 1600 ft in a tributary to Peabody Brook, near the west side of the Shelburne 7.5-minute quadrangle, New Hampshire; the outcrop is about 1300 ft (400 m) southwest of Giant Falls, which is labeled on the quadrangle map. At all localities the fault can be placed within a few centimeters, shows no evidence of cataclasis, and can easily be mistaken for a sharp normal contact.

The relationships shown on the face of Figure 3 (near the indicated younging directions) correspond to sequences exposed at North Newry (Fig. 1a,b) along the Bear River and Highway 26 in the Old Speck Mountain quadrangle. The rocks are in the sillimanite zone, but they are well bedded and display primary features. The fault surface can be placed within a few meters, but is not exposed. The Rangeley Formation north of the fault youngs north (left on Fig. 3), as shown by graded bedding. Here, member B of the Rangeley Formation, composed of dark, rusty-weathered pelitic schist, metasandstone, and polymictic metaconglomerate, is overlain to the north by member C, composed of schist, metasandstone, and quartz metaconglomerate. South of the fault is a south-younging sequence that includes two members of the Madrid Formation (thinly-bedded, calc-silicate rock overlain by thickly-bedded, feldspathic metasandstone and gray pelitic schist), and the Carrabassett Formation (gray pelitic schist showing faint to conspicuous cyclic graded bedding).

The top of the block (Fig. 3) corresponds to the map pattern in nearly 70 square km across the boundary between the Rumford and Old Speck Mountain quadrangles. The sinuous trace of the fault is produced by a late synform-antiform pair thought to plunge southwest at a low angle. Because of complex intersecting deformation, however, the actual plunge of this feature is unknown, and probably cannot be determined by studies of minor folds. The simplest interpretation of this feature is that the fault originally dipped gently southeastward and was later warped to produce the observed map pattern. It is noteworthy that if the upper plate of Lower Devonian rocks were removed, the Rangeley-Perry Mountain Formation contact that is mapped within the antiform in the Old Speck Mountain quadrangle prob-

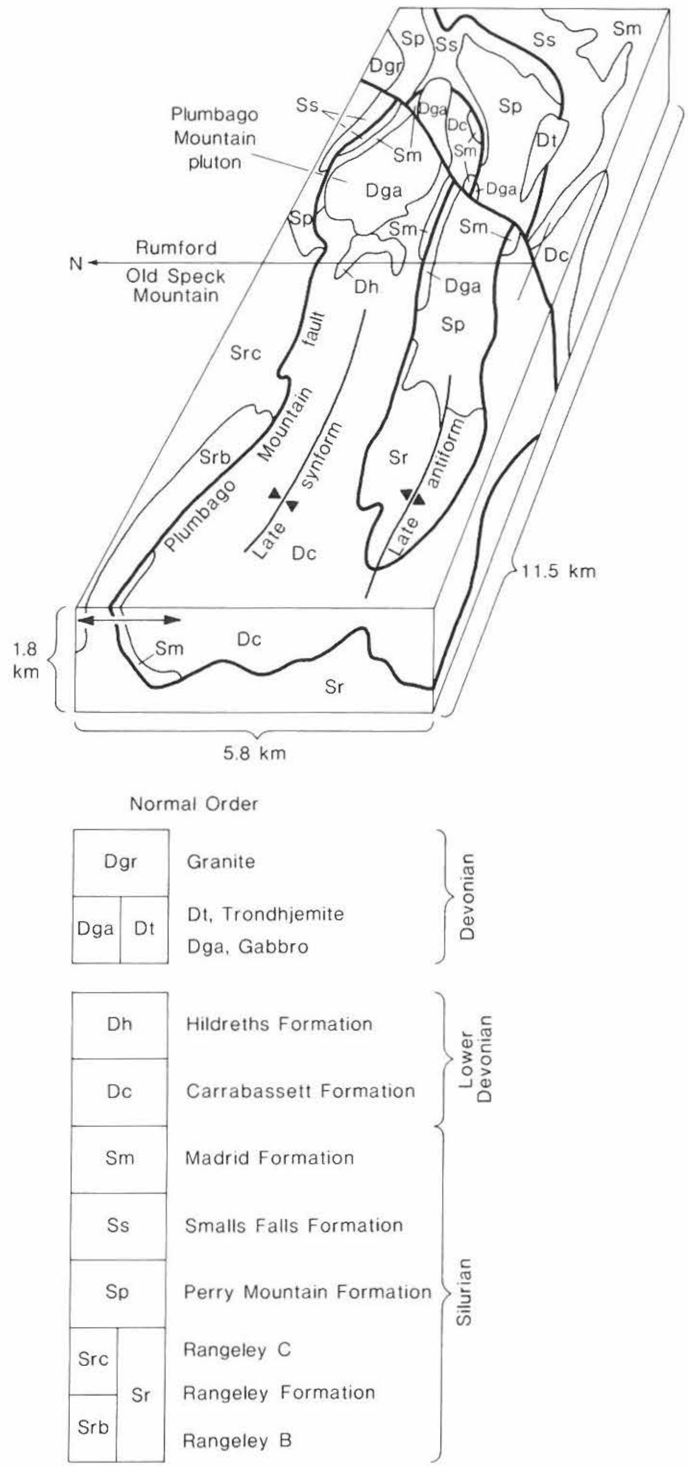


Figure 3. Block diagram showing inferred structure of the Plumbago Mountain early fault and truncation of formations in the upper and lower plates. Old Speck Mountain and Rumford quadrangles. Location shown on Fig. 1b.

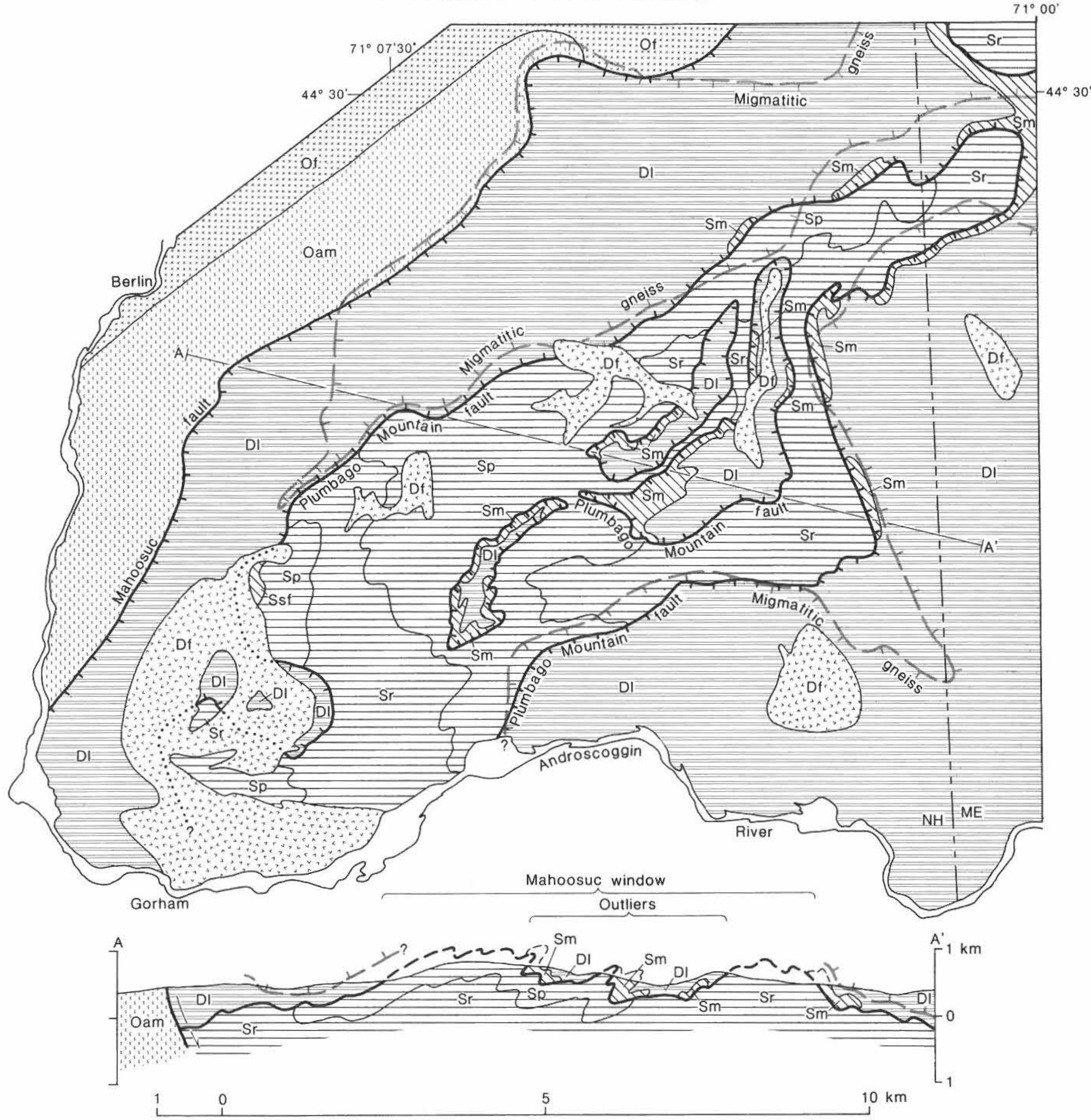


Figure 4. Geologic map and section of the Mahoosuc window, New Hampshire. Plutonic rocks: Df, felsic to intermediate (Devonian); Of, felsic (Ordovician). Stratified metamorphic rocks: Dl, Littleton Formation (Lower Devonian), includes Hildreths Formation at north edge of area, and remnants of Madrid Formation in southeastern part of area; Sm, Madrid, Sf Smalls Falls, Sp, Perry Mountain, Sr, Rangeley Formations (Silurian); Oam, Ammonoosuc Volcanics (Middle Ordovician). Revised by Moench from Billings and Fowler-Billings (1975). Location shown on Fig. 1b.

ably would connect northeastward across the synform to the same contact mapped just east of the quadrangle boundary. The strong discordance between the upper and lower plates indicates that the fault cuts early folds.

The Plumbago Mountain pluton (Fig. 3) is a layered mafic-ultramafic body inferred to have been emplaced along the Plumbago Mountain fault. A sheet of metagabbro lies along the fault on the southeast side of the synform.

In the Mahoosuc Range, northeast of Gorham, New Hampshire, (Fig. 4) is an inferred window, here called the Mahoosuc window, that exposes parts of the Rangeley and Perry Mountain Formations through structurally overlying rocks of the

Madrid and Littleton Formations. The Mahoosuc window contains three small, synclinal, inferred outliers of the Madrid and Littleton Formations. The Littleton Formation of this area is composed mainly of massive gray schist (within the outliers) or migmatitic gneiss (around the window) and is comparable to the Carrabassett Formation of Maine. The southwestern inferred outlier is enclosed in sharply interbedded quartzite and schist of the Perry Mountain Formation, but the two northeastern outliers truncate the Rangeley-Perry Mountain Formation contact. This relationship indicates that the Plumbago Mountain fault truncates earlier folds.

The inferred younger-over-older relationship and the apparent truncation of the Plumbago Mountain fault by the Mahoosuc fault are shown on section A-A′ (Fig. 4). An alternative older-over-younger interpretation, turning the structure "inside out," would change the inferred window to an outlier, and the synclinal outliers to windows. Although the alternative is not ruled out, it seems easier to explain the synclinal bodies as outliers within a window rather than as synclines accidentally exposed as windows within an outlier.

BLUEBERRY MOUNTAIN AND WINTER BROOK EARLY FAULTS AND THE RUMFORD ALLOCHTHON

The Blueberry Mountain and Winter Brook faults are inferred to mark the sole of the Rumford allochthon, on the northwest and southeast sides respectively (Fig. 1a,b). The allochthon is cut internally by the Bald Mountain fault, which has not been seen in outcrop and is inferred to exist as a resolution of structural problems within the Devonian rocks of the allochthon.

The Blueberry Mountain fault extends at least from the Lexington batholith to the Songo pluton (Fig. 1a,b; Osberg et al., 1985). It is cut by several Devonian plutons. The small body of metamorphosed gabbro on the west side of the Rumford pluton (Fig. 1a,b) may be a remnant of a mafic body that was emplaced along the fault. Near the Lexington batholith, where black sulfidic schist of the Carrabassett Formation is in contact with the Madrid Formation, displacement across the fault is small and the allochthon may not be completely detached nearby.

Southwestward, the Blueberry Mountain fault parallels the southeast limb of the Salem anticline; thence, just south of the Rangeley quadrangle, it slices across the core of the anticline to its west limb. This relationship is shown in detail in Figure 5. The oldest exposed formation in the core of the Salem anticline is the Rangeley Formation, which is truncated on the north side of the Blueberry Mountain fault. South of the fault is the Tory Hill syncline, which is mappable northeastward almost to the Lexington batholith. This early fold also is truncated by the fault (Fig. 5). As shown, the fault and the earlier folds are strongly warped by the Tumbledown antiform and related late folds.

In the southwest corner of the area of Figure 5 are two, narrow, bent, canoe-like outliers of Seboomook Formation that are surrounded by the Perry Mountain Formation. A similar outlier is exposed about 6 km farther south- southwest (Moench and Hildreth, 1976). The Seboomook-Perry Mountain Formation contact is the Blueberry Mountain fault. Because these outliers are exposed on high ridges and are apparently synclinal, and because they lie to the northwest of the main trace of the Blueberry Mountain fault, they support the inferred younger-over-older structure of the fault, and suggest further that the fault originally dipped southeast.

Along the southwestern half of the trace of the Blueberry Mountain fault, the Carrabassett, Hildreths, and Seboomook Formations on the east are juxtaposed against generally west-younging rocks of the Rangeley, Perry Mountain, Smalls Falls, Madrid, and Carrabassett Formations, as shown in the lower cross section of Figure 6. The upper section is based on outcrops along and near Philbrick Brook, where the fault is exposed (Moench, 1970, Figs. 12, 13); the brook and location of the outcrop are shown by Moench and Hildreth (1976). The rocks are in the sillimanite zone of metamorphism; east of the Bunker Pond pluton the rocks are gneissic and migmatitic, but west of the pluton the rocks are well-bedded schist. Well bedded pelitic schist of the Seboomook Formation, shown above ground in the section, occupies the southernmost of the three small outliers described above; this outlier is about 2 km north-northwest of where it is shown in the section. Its position west of the main trace of the Blueberry Mountain fault suggests that the fault originally dipped southeastward. West of the fault is a generally west-younging sequence that includes, in ascending order, the Rangeley (exposed about 1 km south of the line of section), Perry Mountain, Smalls Falls, Madrid, and Carrabassett Formations. Where the Hildreths and almost-basal Perry Mountain Formations are juxtaposed, the stratigraphic separation is about 3 km.

The Tumbledown antiform is a post-cleavage, doubly-plunging dome. We propose that it is a modified early anticline. If so, the west-younging sequence that lies between the crest of the antiform and the Blueberry Mountain fault might express back-tilting on the west limb of a rollover, now the tumbledown antiform, above an east-dipping, listric, normal fault, now the deformed Blueberry Mountain fault.

In addition to the outcrop of the Blueberry Mountain fault already described (Moench, 1970), the fault is exposed on the east limb of the southern outlier. The outcrop is 1.1 km north of the north knoll of Flathead Mountain in the Rumford quadrangle; its location is shown by Moench and Hildreth (1976). Here, well-bedded rocks of the Perry Mountain and Seboomook Formations are in knife-sharp contact; graded bedding on both sides youngs away from the contact. A peculiar feature (also seen on a branch of the Blueberry Mountain fault in the southeast corner of the Rangeley quadrangle) is a lens of quartzose metasandstone about 25×50 cm in cross section that occurs exactly on the contact. It is elongate parallel to bedding on both sides, which wraps around the lens. On the basis of its quartzose composition, the lens might have been derived from sandstone of the Perry Mountain Formation. Tentatively, this puzzling feature may be a result of disaggregation, redistribution, and subsequent "balling" of sandstone of the Perry Moun-

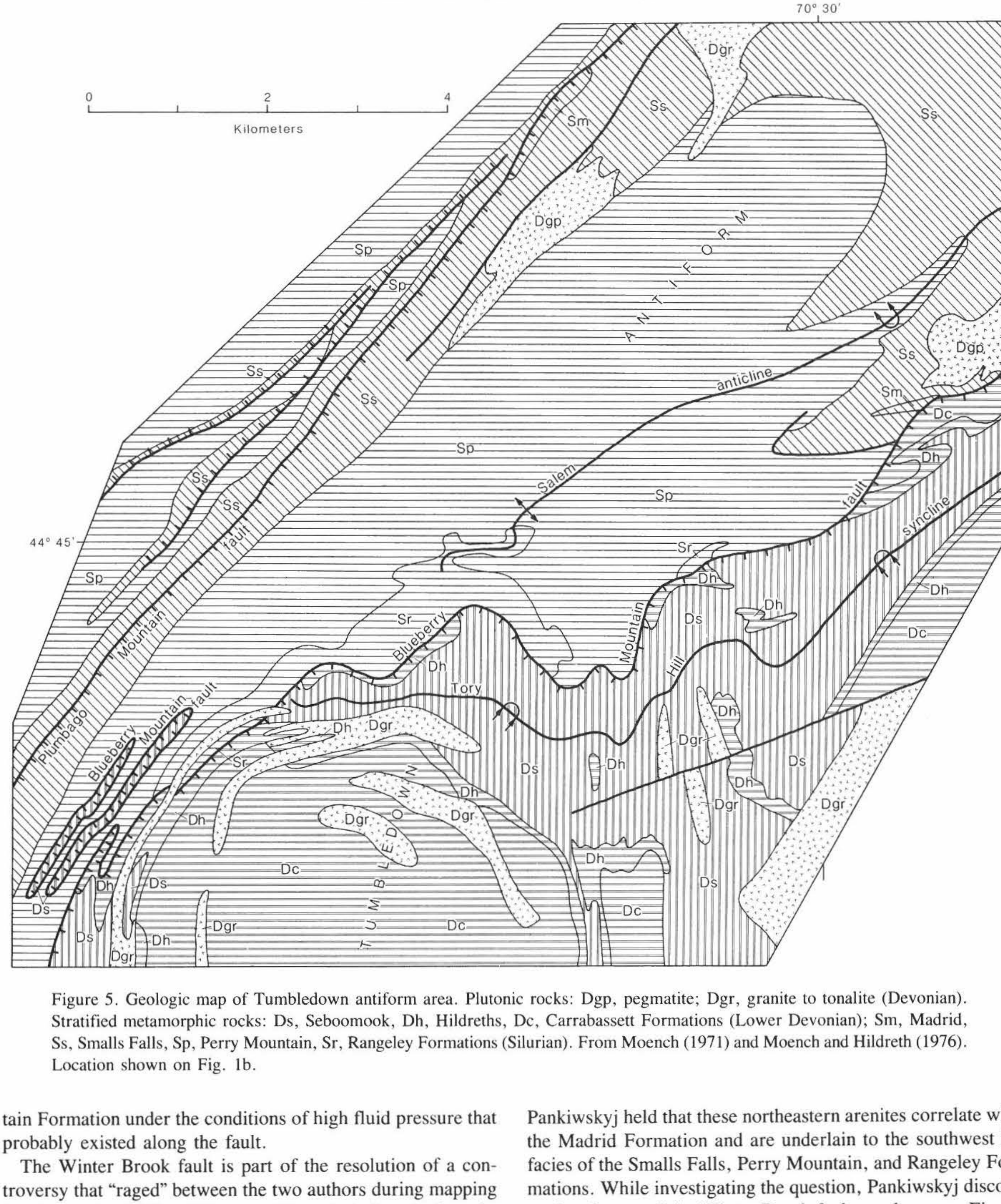


Figure 5. Geologic map of Tumbledown antiform area. Plutonic rocks: Dgp, pegmatite; Dgr, granite to tonalite (Devonian). Stratified metamorphic rocks: Ds, Seboomook, Dh, Hildreths, Dc, Carrabassett Formations (Lower Devonian); Sm, Madrid, Ss, Smalls Falls, Sp, Perry Mountain, Sr, Rangeley Formations (Silurian). From Moench (1971) and Moench and Hildreth (1976). Location shown on Fig. 1b.

tain Formation under the conditions of high fluid pressure that probably existed along the fault.

The Winter Brook fault is part of the resolution of a controversy that "raged" between the two authors during mapping in the early 1970's. Whereas Moench believed that the predominantly gray pelitic rocks and arenites surrounding the Phillips pluton are Devonian in age and are overlain to the northeast (down-plunge and up-section) by younger arenites perhaps related to the Tarratine Formation of northern Maine, Pankiwskyj held that these northeastern arenites correlate with the Madrid Formation and are underlain to the southwest by facies of the Smalls Falls, Perry Mountain, and Rangeley Formations. While investigating the question, Pankiwskyj discovered and mapped the Winter Brook fault, as shown on Figure 1a. As a result, neither author was required to yield; and, moreover, they agreed that the Blueberry Mountain and Winter Brook faults probably are the same, forming the sole of a feature they called the Rumford allochthon. Pankiwskyj mapped

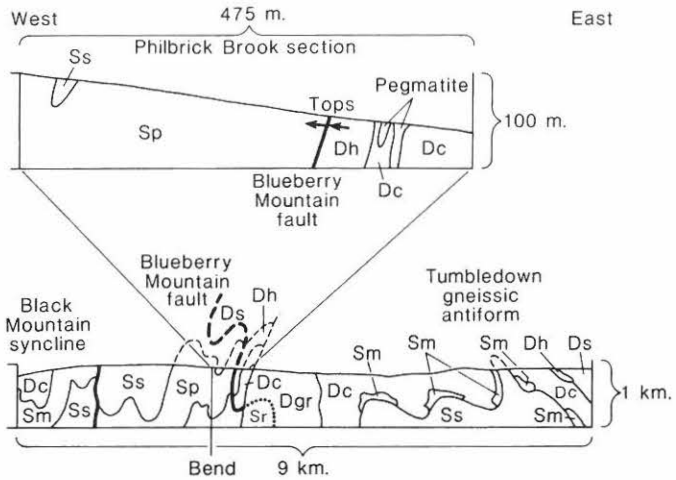


Figure 6. Sections across the Blueberry Mountain early fault. Plutonic rocks: Dgr, granite to granodiorite (Devonian). Stratified metamorphic rocks: Ds, Seboomook, Dh, Hildreths, Dc, Carrabassett Formations (Lower Devonian); Sm, Madrid, Ss, Smalls Falls, Sp, Perry Mountain, Sr, Rangeley Formations (Silurian).

several inferred elongate windows near the southeast side of the allochthon. These windows, according to our interpretation, indicate that the fault originally dipped gently to the northwest.

The Winter Brook fault takes its name from Winter Brook in the southwest corner of the Kingfield quadrangle, where the fault can be seen in outcrop (Pankiwskyj, 1979, p. 42). Here, black sulfidic schist and arenite of the Temple Stream Member of the Seboomook Formation and limy arenite of the Madrid Formation meet at a contact that cuts bedding in both formations; there are no intervening quartz veins. At two other localities, a quartz vein 1 cm thick occupies the trace of the fault; on both sides are zones in which rafts of metasandstone are enclosed in a matrix of less competent pelitic schist. These zones are interpreted as evidence of soft-sediment deformation (Pankiwskyj, 1979, p. 42).

Our interpretation of the Winter Brook fault differs from that of Osberg et al. (1985), who depict the fault as a low-angle, west-directed, older-over-younger thrust that extends at least 200 km across central Maine. Our windows become their outliers. Northeast of the Lexington batholith, the inferred thrust of Osberg et al. (1985) is a rather simple curvilinear feature that displaces the Carrabassett Formation northwestward over somewhat younger rocks of the Carrabassett or the Seboomook Formation, or the Madrid Formation over the Carrabassett Formation. For the most part, formations are not displaced out of their normal stratigraphic order, although at one place member C of the Rangeley Formation is in contact with the Carrabassett Formation. Osberg et al. (1985, section E-E′) project the fault to a depth of 7 km nearly 40 km southeast of its surface trace. Reinterpretation as a high-angle thrust, or even a normal fault, would result in much simpler structure and a displacement of 1 km or less.

Southwest of the Lexington batholith, Osberg et al. (1985) depict the same feature, the Winter Brook fault of this paper, as a complexly deformed thrust having a far greater stratigraphic separation than its correlative northeast of the batholith. Southwest of the batholith, the Silurian Sangerville, Smalls Falls, and Madrid Formations are shown by Osberg et al. (1985) to lie discordantly above various members of the Seboomook Formation. The structure depicted on their section F-F′ requires at least 20 km of thrust displacement. If outcrop characteristics can be disregarded, their interpretation is permitted by the fact there is no certain way in this area to determine younger-over-older versus older-over-younger relationships. The structure can be turned permissively "inside out" simply by placing the barbs (or teeth) on the other side of the fault.

We disagree with the thrust interpretation of the Winter Brook fault for two main reasons. First, it is a complex map pattern, and outcrop characteristics indicate that the Winter Brook fault belongs to the family of early faults mapped farther northwest. Second, present knowledge of Silurian lithofacies on both sides of the fault argues against the long-distance transport that is required by the thrust model. Moench and Pankiwskyj (1987) have divided the Madrid and Smalls Falls Formations into a western facies containing thinly-bedded calc-silicate rocks (upper part of Smalls Falls and lower part of Madrid Formations) and an eastern facies lacking these rocks. This facies change can be placed outside the allochthon of our interpretation a short distance south of the Lexington batholith (Fig. 1b), and within the allochthon about 8 km southeast of Rumford. It probably lies outside the allochthon a few kilometers northwest of the Songo pluton, but more data are needed in that area. Although the facies change is poorly controlled, as presently defined it is consistent with transport of only a few or several kilometers, but not, say, 50 km.

According to our interpretation, the Rumford allochthon originated as a body of semi-lithified, water-bearing sedimentary material a few kilometers thick that moved by gravity sliding a few or several kilometers to its present position. Folding accompanied movement and the body moved far enough over previously folded beds to produce the observed juxtapositions. The northeastern end of the body may not have been completely detached. Although we have no conclusive evidence that bears on direction of transport, we prefer a model of "crowding" toward an axial structural depression along the Kearsarge-central Maine synclinorium.

SUMMARY AND CONCLUSIONS

Northeast-trending premetamorphic early faults of western Maine and adjacent New Hampshire originated prior to Acadian compression, as steeply dipping, southeast-down normal faults along presently exposed relatively little-deformed straight segments, and as gently dipping, younger-over-older detachments along the complexly folded sinuous parts. All were probably listric.

The Hill 2808 and Mahoosuc faults, respectively northeast and southwest of the Mooselookmeguntic batholith, mark the boundary between pre-Silurian and Siluro-Devonian rocks along the Kearsarge-central Maine synclinorium and are tentatively correlated across the batholith. When considered as one feature, the fault extends at least 100 km from near Rangeley, Maine, to north-central New Hampshire. It is interpreted to be a deeply penetrating, southeast-down normal fault which, on the basis of its position near or on the Silurian tectonic hinge, marks the northwestern boundary of the ancestral basin of the synclinorium. The Hill 2808 segment, on the basis of its smaller displacement (0-3 km) and exposed features indicating that the wall rocks were poorly lithified during faulting, is inferred to represent shallower conditions than the Mahoosuc segment, which displaced Silurian and Lower Devonian strata several kilometers downward against hard Ordovician granite and older rocks. The fault cuts the limbs of early folds. However, on the basis of the relationship between the direction of increasing displacement along the Hill 2808 fault and the directions of plunge within the Mountain Pond syncline and Brimstone Mountain anticline to the southeast, early faulting and folding are shown to be coeval and dynamically related. The Hill 2808 fault may have begun to form as a growth fault during deposition of member A of the Lower Silurian Rangeley Formation. By analogy with Gulf Coast structure (see Moench, 1970), the Brimstone Mountain anticline is interpreted to have originated as an open rollover above the assumed listric Hill 2808 fault; the Mountain Pond syncline, lying between the anticline and the fault, may have formed by a combination of back-tilting and drag. During Acadian compression, these folds were greatly tightened and amplified, and slaty cleavage was overprinted across the whole fault-fold structure.

The Barnjum fault was originally mapped between its southwest end near Madrid, Maine, about 25 km northeast to the south end of the gabbroic Sugarloaf pluton, where the Lower Devonian Carrabassett Formation is down-faulted on the southeast against the Silurian Perry Mountain Formation. Between these two points, displacement increases northeastward to more than 1 km; a major syncline on the hanging-wall side plunges in the same direction. This relationship, similar to relationships already described between the Hill 2808 fault and adjacent folds, indicates a temporal and dynamic relation between early faulting and folding; and it shows that the sense of movement in the hanging wall of the Barnjum fault was scissor-like, and southeast and down.

We propose that the Barnjum fault once extended as a low-angle dislocation through a wide area to the north of its originally mapped segment. According to our interpretation, it passed northward from a rupture contained within the conformable Silurian-Devonian basin sequence to one that dislodged shoreward facies of these deposits, approximately along their basal unconformities. The fault is inferred tentatively to mark the sole of the Kennebago outlier, which centers 25 km northwest of the originally mapped segment, but this interpretation is not essential to our hypothesis. We favor a southeastward direction of transport for the whole feature on the basis of the fault-fold relationships already described along the segment of the Barnjum fault that lies between Madrid and the south end of the Sugarloaf pluton. The fault may have originated as a growth fault during deposition of the Silurian Smalls Falls Formation, but the latest movements occurred after deposition of the Lower Devonian Seboomook Formation. Interpreted as a wide-spread, southeast-transported, low-angle detachment feature that steepens southeastward across the Silurian tectonic hinge and eventually dies out and becomes attached within the basin sequence, movement along the Barnjum fault may be likened to a rug sliding down an incline, but tacked to the bottom and one side of the incline. Emplacement of gabbroic magma along the dislocation may have triggered movement.

The Plumbago Mountain early fault is contained within the synclinorium sequence and has a known strike-length of about 70 km, extending from its northeast end about 7 km southwest of Madrid, Maine, almost to Gorham, New Hampshire. It cuts the limbs of earlier folds and is truncated by the Mooselookmeguntic batholith and other plutons. For a distance of about 15 km northeast of the batholith, the fault has a straight trace, indicating a probably steep original dip. The maximum offset here is somewhat less than 1 km, southeast side down. Farther southwest, the trace of the fault becomes increasingly complex, owing to younger folding, and larger amounts of section are excised. Gentler dips and larger displacements are indicated. Structural relationships suggest, but do not prove, that younger rocks lie structurally above older rocks. The most compelling evidence that supports this interpretation is seen in the inferred Mahoosuc window, exposed along the Mahoosuc Range in New Hampshire, where three synclinal bodies of younger rocks occur that would be difficult to explain by an alternative older-over-younger interpretation.

Farther southeast are the Blueberry Mountain and Winter Brook early faults, which outline the 30 × 75 km Rumford allochthon, which lies along the axial zone of the Kearsarge-central Maine synclinorium. The allochthon is interpreted as a body of mainly Lower Devonian rocks that lies discordantly above autochthonous Silurian rocks. We infer that the Blueberry Mountain and Winter Brook faults, respectively, mark the northwest and southeast sides of the allochthon and join at depth to form the sole of the allochthon. The allochthon is ruptured internally by the Bald Mountain early fault. On the basis of the distribution of inferred windows and outliers near the margins of the allochthon, we believe that the allochthon is a younger-over-older detachment whose sole dips inwardly. It does not appear to be far-travelled, and we propose that it formed by detachment and crowding near the axis of the ancestral basin.

Probably the most difficult remaining problem is the younger-over-older versus older-over-younger controversy, which applies specifically to the Plumbago Mountain fault (and Mahoosuc window) and the Blueberry Mountain and Winter Brook faults (and Rumford allochthon). Without disrupting map patterns, the inferred windows and outliers of our interpretation can be turned "inside out," to form older-over-younger thrusts. The small syn-

clinal inferred outliers that occur within the inferred Mahoosuc window and on ridges northwest of the main trace of the Blueberry Mountain fault support our interpretation, but not conclusively. By turning windows into outliers and outliers into windows along the Winter Brook fault, but not along the Blueberry Mountain fault, Osberg et al. (1985, sections E-E′, F-F′) interpret the Blueberry Mountain fault as a low-angle normal fault and the Winter Brook fault as a low-angle thrust fault of large displacement. On the basis of their conspicuous similarity, however, we consider the Plumbago Mountain, Blueberry Mountain, and Winter Brook faults to belong to the same family of structures; as such, a common origin seems necessary.

The most compelling line of evidence for the younger-over-older model is the comparison of coeval sedimentary facies across major faults. Whereas the younger-over-older model requires only enough horizontal displacement (a few or several kilometers) to account for the structural juxtapositions, the older-over-younger thrust model requires long-distance transport across strike, say from the margin to the center of the sedimentary basin. Under the thrust model one would expect to see a juxtaposition of sedimentary facies derived from different parts of the basin. So far, no such juxtaposition has been identified.

Another problem is presented by the observations: (1) that the early faults truncate the limbs, and locally the axial traces of folds; (2) that folding and faulting are demonstrably contemporaneous near the ends of two early faults; and (3) that the faults are deformed by early cleavage and folds that predate emplacement of the 400 Ma Lexington batholith, and are further deformed by younger folds. The apparent inconsistencies between relationships (1) and (2) can be resolved by a model of simultaneous faulting accompanied by folding along two or more sub-parallel listric faults. According to this model, a syncline-anticline pair that developed contemporaneously with faulting within the rocks above one fault would tend to be "decapitated" by the next higher, or basinward, fault. Carried to the extreme, the resulting pattern of juxtapositions would be very complex. Almost everywhere the faults would cross the limbs of contemporaneous folds; but only near the identified ends of faults would evidence be preserved that faulting and folding were in fact contemporaneous. During younger compression (3), anticlines and synclines that developed earlier would only be tightened and amplified, as we infer to have happened in the area of this study.

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Geologic Relations Within the Shale-Wacke Sequence in South-Central Maine

Philip H. Osberg
Department of Geological Sciences
University of Maine
Orono, Maine 04469

ABSTRACT

The shale-wacke sequence of south-central Maine is in the east flank of the Kearsarge-central Maine synclinorium. The dominantly Silurian stratigraphy consists of a monotonous assemblage of metamorphosed wackes, gray and green shales, thin-bedded limestones, and lenticular conglomerates. Sparse, local fossil data indicate that this section accumulated in Ashgillian or earlier(?) through Pridolian(?) time. Six formations have been defined: from oldest to youngest, these are the Hutchins Corner (new name), Waterville, Sangerville, Perry Mountain, Smalls Falls, and Madrid Formations. Although these units overlie one another in the area of study, when taken over a larger geographic area, they undoubtedly in part interfinger with each other. This stratigraphic section is inferred to rest unconformably on the Nehumkeag Pond unit of the Cushing Formation.

Structural features include a pervasive schistosity, a second cleavage, various fold forms, and faults. The youngest folds (F_3) are upright and asymmetrical, and the second cleavage is parallel to their axial surfaces. An earlier group of folds (F_2) is also upright but mainly isoclinal. The pervasive foliation is parallel to the axial surfaces of these folds. A still earlier set of recumbent and nearly isoclinal folds (F_1) is postulated to exist on the basis of limited minor structural features and on observations of upward and downward-facing F_2 folds. Foliation associated with the early recumbent folds is mostly confused with the schistosity that is parallel to the axial surfaces of F_2. Late cleavage bands and boundinage are less prominent than the above features.

The Messalonskee Lake thrust cuts across large F_1 folds, but is in turn folded by F_2. Its easterly dip and the relative ages of the formations involved suggest westward transport on this thrust. Two high-angle faults are present within the area of study. These post-date F_3 and an associated metamorphic event.

Hypabyssal dikes and stocks of andesite and dacite intrude the Silurian section and postdate F_2, but predate F_3. Plutons of granodiorite and granite also intrude all elements of the stratigraphy of the synclinorium. Their radiometric ages range from approximately 395 to 360 Ma. At least the older plutons cut F_2, but predate F_3.

The sedimentary rocks have been regionally metamorphosed to at least low greenschist facies either synchronously or slightly after the development of F_2. The hypabyssal dikes and stocks, and probably the older plutons of granite and granodiorite, have been metamorphosed in a second event that is approximately coeval with F_3. This second metamorphic event ranges from chlorite zone in the north to sillimanite zone in the south and accounts for most of the observed metamorphic features. $^{40}Ar/^{39}Ar$ closure ages for metamorphic minerals suggest a Late Devonian age for the second metamorphic event. Chlorite pseudomorphs after cordierite and biotite may be due to a third metamorphic event, possibly Carboniferous, whose intensity increases to the southwest.

The Silurian shale-wacke section is interpreted to have prograded in an east-facing sedimentary prism with provenance in a volcanic terrane to the west. Early folds (F_1), thrusts, and upright isoclinal folds (F_2) formed after the deposition of Silurian and Lower Devonian (probably Emsian) rocks and before the intrusion of plutons dated at approximately 395 Ma. The duration of the F_1 to F_2 deformational sequence was probably of the order of 10 million years. The asymmetric folds (F_3) deform hypabyssal rocks and at least the earlier plutonic rocks and may be synchronous with the second metamorphic event. Intrusion of granodiorite and granite continued until the end of the Devonian and into the Carboniferous.

INTRODUCTION

The shale-wacke sequence in the east flank of the Kearsarge-central Maine synclinorium of south-central Maine (Figs. 1 and 2) consists of a monotonous succession of wackes, shales, thin shaly limestones, lenticular conglomerates, and their metamorphic equivalents. Repeated lithologic sequences that are not time-equivalent, as well as structural complexities, serve to confuse the geologic relationships. Furthermore, the rocks of south-central Maine have been intruded by plutons, and much of the terrane has been multiply metamorphosed.

The purpose of this paper is to present some of the results of studies made over a twenty-five year period by the author. The main focus is on the stratigraphic and structural relationships, but because plutonism and metamorphism bear heavily on solving the stratigraphic and structural problems, their relevant aspects are also considered.

The area under consideration is located in south-central Maine (Fig. 1) bounded approximately by latitude 44°15′N and longitude 70°W, and latitude 45°N and longitude 69°15′W. It consists of parts of eight 15-minute quadrangles.

This paper depends heavily on the work of other geologists who have done detailed work in specific areas or on selected topics within the area of study. Mapping credits for Figure 2 are given in Appendix 1. Among those who have provided valuable insights into the geology of the region through publication and discussions are Allan Ludman, John Griffin, Kost Pankiwskyj, William Berry, Arthur Hussey, Donald Newberg, Robert Moench, David Wones, John Ferry, David Dallmeyer, and David Stewart. An early draft of this paper benefited greatly from reviews by Robert Moench and David Stewart. Subsequent reviews by David Stewart, Daniel Milton, Allan Ludman, and John Lyons further improved the paper.

Early work in this region was supported by the Maine Geological Survey. More recent work has been under the auspices of the U.S. Geological Survey.

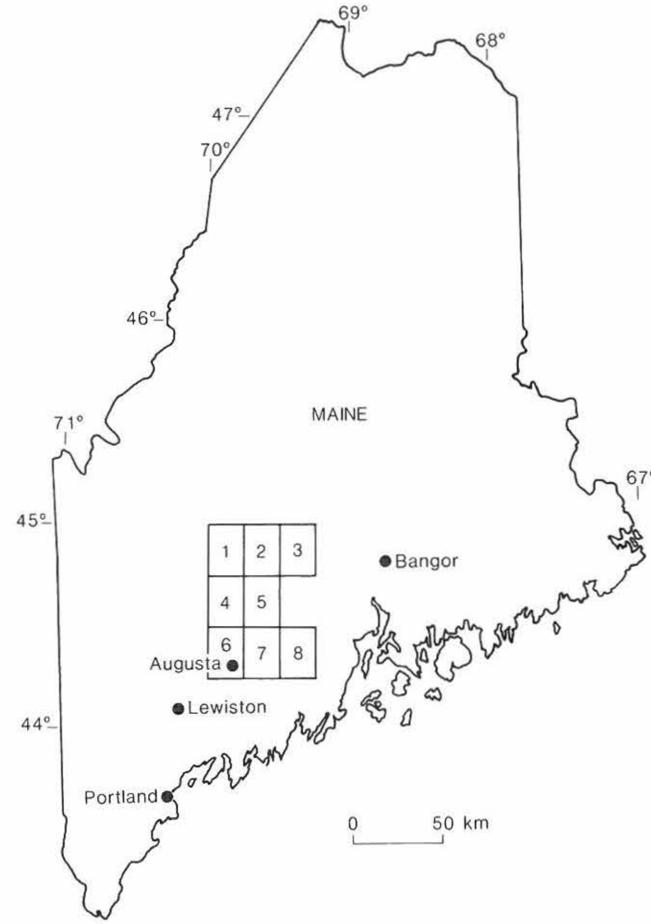

Figure 1. Index map for south-central Maine showing the 15′ quadrangles covered by this study. Quadrangles are identified by the following numbers: 1 = Anson, 2 = Skowhegan, 3 = Pittsfield, 4 = Norridgewock, 5 = Waterville, 6 = Augusta, 7 = Vassalboro, and 8 = Liberty.

STRATIGRAPHY

General Statement

The determination of the stratigraphic section has been a vexing problem throughout the history of geologic work within the region. An understanding of the stratigraphy is complicated by isoclinal upright folds that turn the beds throughout the region into a vertical position. Moreover, both normal and inverted sections have been identified. Because of the structural complications, outcrops containing critical facing directions must be sought at formational contacts; these constraints, along with the paucity of outcrop, limit the number of stratigraphically relevant observations (Fig. 2). Finally, although numerous fossil collections have been made from the rocks of the region (Osberg, 1968; Griffin, 1973; Pankiwskyj et al., 1976; Ludman, 1977), the state of preservation of the fossil forms in many cases limits their usefulness to unequivocally fix the age of the sequence.

The work of Osberg (1968), Pankiwskyj et al. (1976), and Ludman (1976) forged the basis for an integrated stratigraphic section for south-central Maine. Osberg (1980) suggested that

Figure 2. Geologic map of south-central Maine. Section X-Y is shown in Figure 13. 1 = Hartland pluton (360 ± 8 Ma), 2 = Old Point pluton, 3 = Norridgewock pluton, 4 = David Pond pluton, 5 = Lord Hill pluton, 6 = Hallowell pluton (387 ± 11 Ma), 7 = Togus pluton (394 ± 8 Ma), and 8 = Three Mile Pond pluton (381 ± 14 Ma). Locations of towns and cities indicated by open circles: HY = Harmony, AT = Athens, H = Hartland, S = Skowhegan, C = Canaan, PT = Pittsfield, BS = Benton Station, EB = East Benton, M = Mercer, W = Waterville, CH = China, HC = Hutchins Corner, NP = North Palermo, CC = Carrs Corner, P = Palermo, MA = Manchester, A = Augusta, and SW = South Windsor.

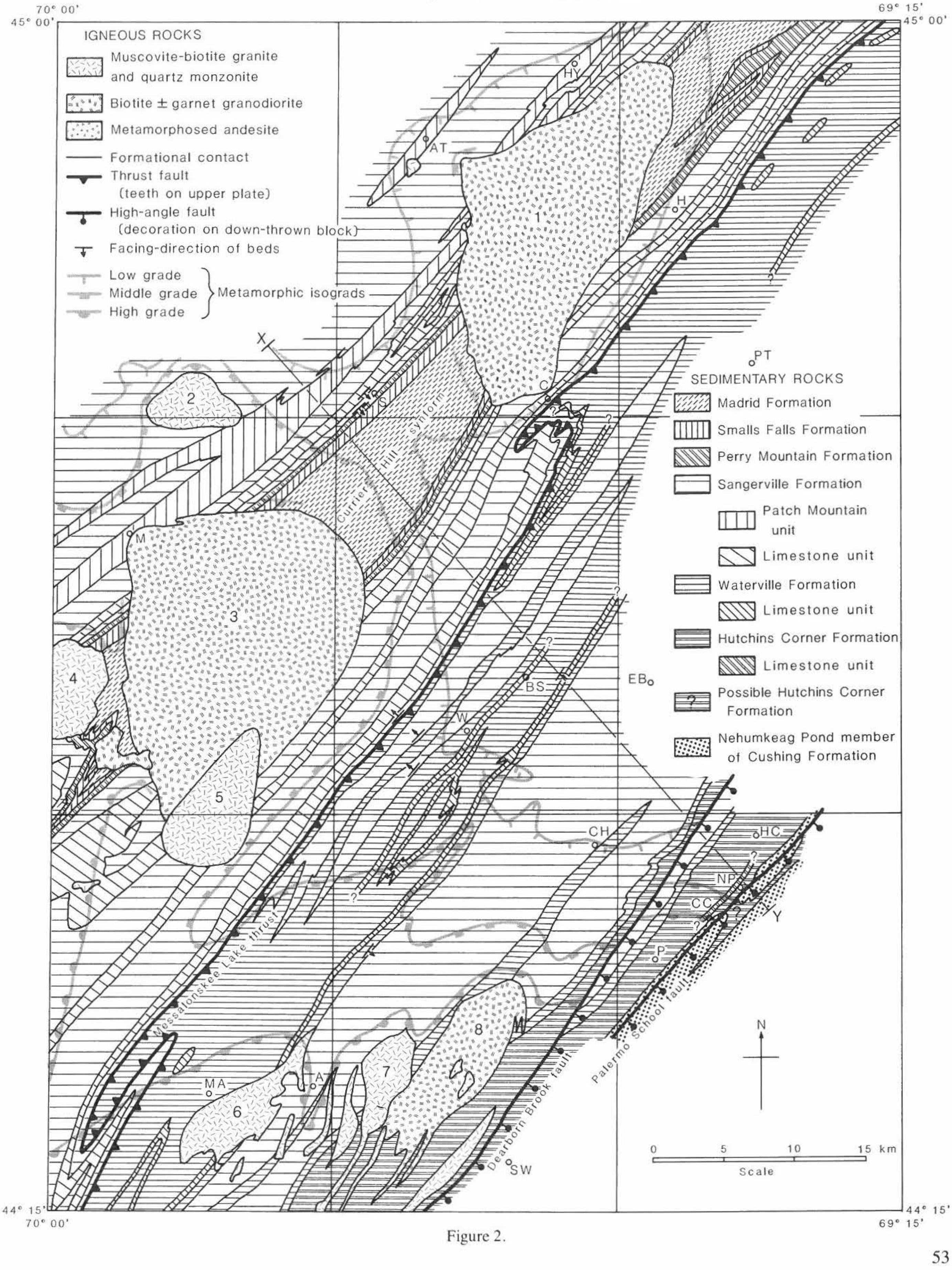
70° 00'
45° 00'
69° 15'
45° 00'
IGNEOUS ROCKS
Muscovite-biotite granite and quartz monzonite
Biotite ± garnet granodiorite
Metamorphosed andesite
Formational contact
Thrust fault (teeth on upper plate)
High-angle fault (decoration on down-thrown block)
Facing-direction of beds
Low grade
Middle grade
High grade
Metamorphic isograds
SEDIMENTARY ROCKS
Madrid Formation
Smalls Falls Formation
Perry Mountain Formation
Sangerville Formation
Patch Mountain unit
Limestone unit
Waterville Formation
Limestone unit
Hutchins Corner Formation
Limestone unit
Possible Hutchins Corner Formation
Nehumkeag Pond member of Cushing Formation
HY
AT
H
PT
C
S
X
Y
Currier Hill synform
M
EB
BS
W
CH
HC
NP
CC
P
MA
A
SW
Messalonskee Lake thrust
Dearborn Brook fault
Palermo School fault
N
0
5
10
15 km
Scale
44° 15'
70° 00'
44° 15'
69° 15'

Figure 2.

TABLE 1. STRATIGRAPHIC INTERPRETATIONS FOR SOUTH-CENTRAL MAINE. NAMES CURRENTLY IN USE ARE CAPITALIZED.

Age	A Osberg (1968)		B Ludman (1976)	C Pankiwskyj (1979)	D This paper
Early Devonian(?) or Pridolian(?)	Vassalboro Formation		Brighton Formation	Fall Brook Formation	MADRID FORMATION
Ludlovian			Eddy Formation	Parkman Hill Formation	SMALLS FALLS FORMATION
Wenlockian	Western facies of the WATERVILLE FORMATION	Eastern facies of the WATERVILLE FORMATION	Upper SANGERVILLE FORMATION	Anasagunticook Formation Upper SANGERVILLE FORMATION	PERRY MOUNTAIN and SANGERVILLE FORMATIONS (includes western facies of Waterville, Mayflower Hill, and part of Vassalboro Formations) WATERVILLE FORMATION (=Anasagunticook Formation)
Llandoverian and Ashgillian(?)	Mayflower Hill Formation		Lower SANGERVILLE FORMATION	Lower SANGERVILLE FORMATION	HUTCHINS CORNER FORMATION
Pre-Ashgillian(?)					NEHUMKEAG POND unit, CUSHING FORMATION

the sequence at Waterville is partly equivalent to and partly underlies the sequence that had been established near Skowhegan by Ludman (1977) and Pankiwskyj (1979), and this relationship was adopted for the new Bedrock Geologic Map of Maine (Osberg et al., 1985). A historical comparison of stratigraphic columns is presented in Table 1, columns A, B, and C.

These stratigraphic relations are further modified in this paper and presented as column D in Table 1. The stratigraphic position of the Vassalboro Formation and its internal relationships have never been satisfactorily studied or understood. Recent observations by Newberg (1984, 1985) and by the author suggest that the Vassalboro Formation includes rocks that can be correlated lithologically with the Sangerville and Waterville Formations, as well as rocks older than the Waterville Formation. As a consequence, the name, Vassalboro Formation, should be abandoned. In this paper the rocks that lie beneath the Waterville Formation are called Hutchins Corner Formation (new name).

In stratigraphic column D of Table 1, the Nehumkeag Pond unit of the Cushing Formation (Newberg, 1984; Hussey, this volume) is thought to be the oldest unit. It is interpreted to lie unconformably beneath the Hutchins Corner Formation, and to be pre-latest Ordovician in age. The Hutchins Corner Formation may be Llandoverian or older in age. The Waterville and Sangerville Formations are Llandoverian to Wenlockian in age. The Perry Mountain and Smalls Falls Formations are possibly latest Wenlockian to definite Ludlovian in age. The Madrid Formation may be Pridolian in age.

Critical facing evidence exposed at the "Great Eddy" in the Kennebec River 0.8 km south of Skowhegan (Fig. 2) establishes the stratigraphic relation between the Smalls Falls and the Perry Mountain Formations. Cross-beds in quartzites of the Perry Mountain Formation 1.5 m from the contact with the Smalls Falls Formation indicate that the Smalls Falls Formation is the younger unit. Because the formational contacts are fairly straight and uncomplicated in this section, this observation sets the stratigraphy for the entire section that includes the Sangerville, Perry Mountain, Smalls Falls, and Madrid Formations. This sequencing corroborates the section previously set for the same formations in western Maine by Moench (in Osberg et al., 1968).

Facing directions between the Sangerville and the Waterville Formations located on Kennedy Drive, 1.1 km west of Messalonskee Stream in Waterville (Fig. 2), determine the relationship between the Sangerville and the Waterville Formations. In this section black sulfidic phyllite a few tens of meters thick, considered to be part of the Sangerville Formation, separates wackes of the Sangerville Formation from the phyllites of the Waterville Formation. Graded beds in this outcrop occur within a meter of the contact between the wacke and black sulfidic phyllite, indicating that the wackes of the Sangerville Formation are younger than the black sulfidic phyllite and the Waterville Formation. Similar observations may be made on Route 95, 1.5 km north of the Waterville exit (#34).

The stratigraphic relations within the Vassalboro Formation (as formerly used) are as follows. A reexamination of facing

relationships at the contact between the Waterville Formation and former Vassalboro Formation (Fig. 2) has reaffirmed that the rocks of the Vassalboro Formation are younger than the Waterville Formation as was originally indicated (Osberg, 1968). A thin black sulfidic phyllite similar to that at the base of the Sangerville Formation separates the former Vassalboro Formation from the Waterville Formation, so that stratigraphic symmetry exists east and west of the Waterville Formation. These relationships indicate that the former Vassalboro Formation, including its type locality, are to be equated with the Sangerville Formation. An eastern antiformal inlier of the Waterville Formation is found east of China (Fig. 2), and the Waterville Formation reemerges to the east in the east limb of a synform. The rocks lying east of the Waterville Formation have been placed in the Hutchins Corner Formation.

The age assignments are based on scattered fossil localities that have been described from the area (Osberg, 1968; Griffin, 1973; Ludman, 1976; Pankiwskyj et al., 1976). The fossil collections contain few genera, are fragmental and badly strained, so that age determinations are difficult to make; therefore, many of the age assignments are tenuous. Monograptids collected from the Waterville Formation within the Stetson quadrangle have a reported age of late Llandoverian to Wenlockian (Pankiwskyj et al., 1976), and dendroid graptolites from Benton Station suggest a Wenlockian age (Osberg, 1968). The Sangerville Formation contains graptolites that indicate late Llandoverian through Wenlockian ages (Perkins, 1924; Osberg, 1968; Pankiwskyj et al., 1976), but since the distinction between Llandoverian and Wenlockian graptolites is not always easy to make, particularly on the basis of such fragmented fossils, and because the underlying Waterville is thought to be Wenlockian, the Sangerville is here assigned to the Wenlockian as well. The Smalls Falls Formation of south-central Maine contains Ludlovian graptolites (Pankiwskyj et al., 1976). The Perry Mountain, Madrid, and Hutchins Corner Formations, and the Nehumkeag Pond unit of the Cushing Formation are unfossiliferous.

The age of the Hutchins Corner Formation is uncertain. It forms the basal part of an apparently conformable sequence containing Silurian fossils, and it rests with sharp contact on rocks that had a different metamorphic history and that may have an Ordovician or older age. Regional relations suggest that the Hutchins Corner Formation is latest Ordovician to Llandoverian in age.

The age of the Nehumkeag Pond unit is Middle Ordovician or older. This conclusion is based on the observation that the Nehumkeag Pond unit contains a relict kyanite metamorphism overprinted by an andalusite event, whereas the Hutchins Corner Formation displays only andalusite metamorphism. In addition, W. J. Olszewski (pers. commun., 1984) has obtained an Ordovician Rb-Sr whole rock age from interlayered quartzofeldspathic gneisses and amphibolites located just south of the area of this study. This age is similar to a Rb-Sr whole rock age reported by Brookins and Hussey (1978) from the Cushing Formation near Portland. However, these ages may not reflect the true age of these rocks, but rather may define the age of metamorphic recrystallization, so their real age may be considerably older, possibly as old as Late Proterozoic.

The sequence described above, when considered over a larger area, may involve facies relationships. Permissible relations are depicted in Figure 3. Time-lines are shown horizontally, and the position of fossil collections are shown by special symbol. Both the Smalls Falls and Perry Mountain Formations are interpreted to thin toward the east, and these formations are essentially parallel to time-lines. The relationships between the Rangeley-Sangerville-Greenvale Cove-Waterville Formations are not well constrained. In Figure 3a the Rangeley and Sangerville Formations are contiguous, and the Greenvale Cove and Waterville Formations are equated. The Rangeley-Sangerville Formations contact with the Greenvale Cove-Waterville Formations transgresses time. These relationships suggest the possibility that the Waterville Formation may climb upward in time toward the east and coalesce with the Perry Mountain Formation at the expense of the Sangerville Formation. In Figure 3b the Greenvale Cove and Waterville Formations are not equated; the Greenvale Cove Formation pinches

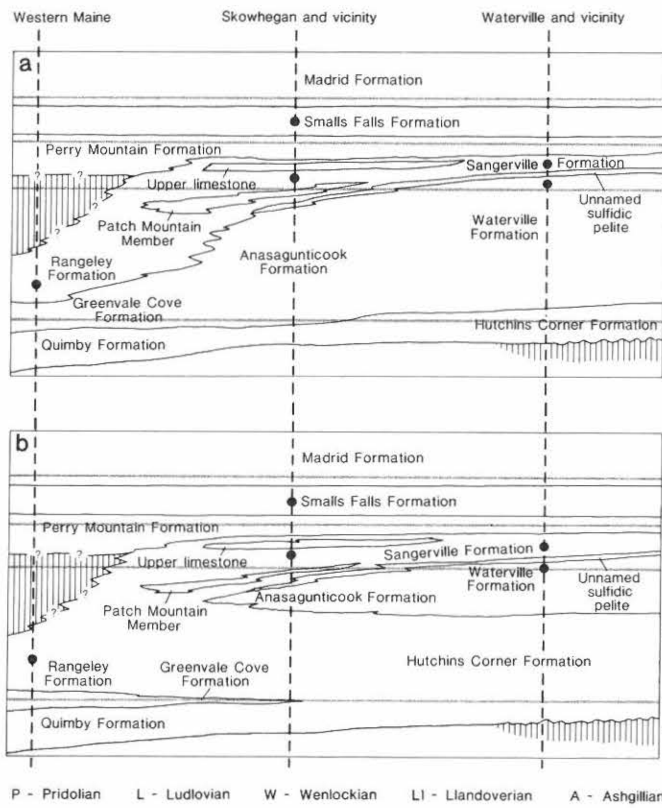

Figure 3. Stratigraphic diagrams showing permissible facies relations in south-central Maine. (a) Waterville Formation shown connected through the Anasagunticook Formation to the Greenvale Cove Formation; Sangerville and Rangeley Formations are equated, and the Quimby and Hutchins Corner Formations are equated. (b) Waterville and the Anasagunticook Formations shown as a tongue, pinching out toward the west, and the Greenvale Cove Formation as pinching out toward the east. The Quimby, Hutchins Corner, Rangeley, and Sangerville Formations are continuous. Dots show fossil control.

out toward the east, and the Waterville Formation thins to a feather edge toward the west. The Rangeley and Quimby Formations are proximal and the Sangerville and Hutchins Corner Formations are more distal parts of a graywacke apron.

Nehumkeag Pond unit of the Cushing Formation

Katz (1917) designated rocks in the Casco Bay quadrangle that he interpreted as having a plutonic origin as the Cushing granodiorite. Hussey (1971), in mapping the Portland quadrangle, found that rocks which were in part continuous with the Cushing granodiorite were interbedded with amphibolitic gneiss, and he interpreted the sequence of feldspathic and amphibolitic gneisses as having a volcaniclastic origin. Hussey (1971) changed the name to Cushing Formation because the term granodiorite no longer seemed appropriate and because of the heterogeneous character of the unit. Similar rocks have been traced into south-central Maine, and therefore, the name Cushing Formation may be applicable in this section as well (see Hussey, this volume). Perkins and Smith (1925) used the name Knox gneiss for some of these same rocks, but clearly the Cushing Formation has precedence. More recently, Newberg (1984) has proposed that the Cushing Formation in south-central Maine be divided into several units, among which is the Nehumkeag Pond unit.

The Nehumkeag Pond unit is exposed along the southeast margin of the area of this study (Fig. 2). Its west contact may be traced from the vicinity of South Windsor northeastward to Palermo. The area of the map does not extend to its eastern contact.

Little can be inferred about the thickness of the Nehumkeag Pond unit. Its west contact is thought in part to be an unconformity and its east contact does not lie within the area of this study. Newberg (1985) does not give a thickness.

No attempt has been made to decipher the internal stratigraphy of the Nehumkeag Pond unit in this report. Common lithologies include tan-weathered, quartzo-feldspathic gneissic schist containing staurolite, garnet, and relict kyanite, rusty black sulfidic schist, white, lineated quartzite, calc-silicate granofels and marble, and interlayered biotite-bearing amphibolite and plagioclase-quartz-biotite gneiss and pegmatitic gneiss. All of the sub-units within the Nehumkeag Pond unit have been metamorphosed at middle grade.

Hutchins Corner Formation (new name)

The name Hutchins Corner Formation is used for rocks lying beneath the Waterville Formation and above the Nehumkeag Pond unit and replaces the name Vassalboro Formation. The type-locality of the Hutchins Corner Formation is in the vicinity of Hutchins Corner (Fig. 2) in the Palermo quadrangle. Representative outcrops of the upper part of the Hutchins Corner Formation may be observed for a distance of 1.6 km south of Hutchins Corner along the road leading to North Palermo and for a distance of 1.3 km along the road west of Hutchins Corner. The lower part of the Hutchins Corner Formation may be observed in North Palermo and in the vicinity of Carrs Corner (Fig. 2) in the Palermo quadrangle. A distinctive unit of marble that separates the upper and lower parts of the formation outcrops on a woods road 0.5 km west-southwest of Carrs Corner.

The Hutchins Corner Formation underlies a tract that extends northeasterly from the vicinity of South Windsor (Fig. 2). The width of this tract is approximately 1.5 km at its south end and 5 km at its north end. In addition, rocks possibly assigned to the Hutchins Corner Formation form a complex outlier (Newberg, 1985) within the general outcrop of the Nehumkeag Pond unit of the Cushing Formation.

The minimum breadth of outcrop of the Hutchins Corner Formation is approximately 1500 m; therefore, an appropriate thickness could be on the order of 540 m after taking into consideration the nature of folding.

The Hutchins Corner Formation is represented by slightly rusty, gray to dark gray, variably bedded, calcareous and noncalcareous wacke interbedded with less abundant packets of thin-bedded phyllite and quartzite. A thin marble unit has been discontinuously mapped within the Hutchins Corner Formation in the vicinity of North Palermo, and slabby sandstones east of the marble are considered to be the basal beds of the formation.

Wackes and siltstone make up the bulk of the Hutchins Corner Formation. These beds have thicknesses that range from 7 cm to over 1 m. Most beds are ungraded and consist of fine sand to coarse silt exhibiting a fine sedimentary lamination. The wackes and sandstone at low grade contain clasts of subrounded to subangular quartz, plagioclase, and muscovite. At higher metamorphic grades biotite, actinolite, garnet, zoisite, Ca-amphibole, and diopside may be present.

Packets of thin-bedded phyllite and quartzite contain beds of slightly micaceous quartzite 3 mm to 5 cm thick alternating with quartz-muscovite-chlorite phyllite beds of approximately equal thickness. Primary features are generally absent. Where metamorphosed, muscovite, biotite, garnet, andalusite, and sillimanite may appear.

The thin discontinuously mapped marble unit (Fig. 2) consists of light gray calcite marble and blue-gray, biotite- and amphibole-bearing calcite marble. The unit is not more than a few tens of meters thick. North of North Palermo it was once quarried for agricultural lime.

The slabby sandstones consist of quartz-plagioclase-biotite-muscovite granulite and biotite schist. The beds of granulite are devoid of calcite or calc-silicate minerals in marked contrast with the wackes higher in the formation. Beds of granulite range from 2 cm to as much as 8 cm in thickness; the biotite schist forms selvages rarely greater than a few millimeters thick against adjacent granulite.

The Hutchins Corner Formation is interpreted to lie unconformably on the Nehumkeag Pond unit, although it is to be noted that Newberg (1985) considered these relations differently. This difference of interpretation hinges on the assigned stratigraphic position of the basal beds of the Hutchins Corner Formation.

Newberg (1985) interpreted these beds to be part of the Nehumkeag Pond unit, whereas in this paper they are considered to be basal Hutchins Corner Formation. The contact between the rocks in question and the Nehumkeag Pond unit is beautifully exposed in a field 0.1 km south of Carrs Corner. Here the contact between the two units is knife-sharp, and the Nehumkeag Pond unit contains relict kyanite metamorphism overprinted by andalusite metamorphism, whereas the lower part of the Hutchins Corner Formation contains only minerals consistent with andalusite metamorphism.

The Hutchins Corner Formation has uncertain correlation with rocks in northern Maine. To the south it may correlate with at least part of the Berwick Formation.

Waterville Formation

Perkins and Smith (1925) named the rocks west of the Kennebec River in the vicinity of Waterville (Fig. 2) as the Waterville shale. Osberg (1968) renamed this unit the Waterville Formation, partly because of the diversity of included lithologies, but also because the formation includes different mineral assemblages at different metamorphic grades. Osberg (1968) designated the thinly bedded quartz-mica schist and interbedded quartzite as the eastern facies of the Waterville Formation and certain wackes, phyllite, and limestone as the western facies. The western facies of the Waterville Formation now is included in the Sangerville Formation (Osberg, 1980).

The Waterville Formation is exposed in a northeast-trending belt that extends from the vicinity of Manchester to the vicinity of Pittsfield (Fig. 2). The Waterville Formation is also exposed in inliers south of Augusta and south of China. Eastern belts of the Waterville Formation trend northeasterly west of Palermo. Its maximum breadth of outcrop is about 14 km at the latitude of Pittsfield, and its minimum breadth of outcrop is south of China, where it measures about 600 m.

The Waterville Formation is nearly homoclinal in the vicinity of South China and near Palermo, so its thickness must be on the order of 500 to 600 m as measured from its breadth of outcrop. There are few constraints on its thickness to the west because its lower contact is not exposed.

The dominant lithology of the Waterville Formation is thinly bedded, quartz-mica phyllite and quartzite. Phyllitic beds alternate with quartzitic beds, the thickness of the beds ranging from 1 to 7 cm, with the phyllitic beds being slightly thicker. The quartzite beds locally exhibit small, ill-defined cross-beds, and in some outcrops the quartzite beds have a graded relationship to the phyllitic beds. The phyllitic beds everywhere exhibit an excellent schistosity which over most of the area is essentially parallel to bedding. Chlorite, muscovite, biotite, garnet, staurolite, cordierite, andalusite, and sillimanite occur at appropriate metamorphic grade.

A minor variant within the thin-bedded phyllite and quartzite unit is rusty-weathered, sulfidic quartz-mica phyllite. This consists of quartz with simple boundaries, muscovite with good shape alignment, and abundant pyrrhotite. This lithology occurs at various stratigraphic positions, and attempts to map some of the larger units were unsuccessful.

A distinctive gray limestone member at about the middle of the formation is from 30 to 100 m thick (Fig. 2). It characteristically is well bedded with 2 to 8 cm beds of limestone interlayered with less common beds of quartz-mica phyllite and quartz phyllite, 5 mm to 5 cm thick. Generally, the limestone beds are devoid of primary structure, but locally at low metamorphic grade they display cross-beds. Mg-biotite, Ca-amphibole, garnet, and diopside appear at higher metamorphic grades.

The thin beds of quartz-mica phyllite and quartz phyllite that are interbedded with the limestone consist of fine-grained quartz, muscovite, and/or chlorite. Many of these beds carry enough pyrrhotite so they weather to a rusty brown. Little happens to these beds with increase in metamorphic grade. Biotite becomes abundant and sparse garnet is present.

The Waterville Formation is lithologically similar and probably equivalent to part of the Greenvale Cove Formation in western Maine. To the north it is similar to the rocks of Island Falls (Ekren and Frischknecht, 1967), but to the south possible lithologic equivalents have not been identified. It may be in part a time equivalent of the Sangerville Formation.

Sangerville Formation

Ludman and Griffin (1974) referred to wackes, conglomerates, slates, and limestone in the Guilford and Kingsbury quadrangles immediately north of the area of this report as Sangerville Formation. Rocks of the Sangerville Formation extend southwestward into the Skowhegan and Anson quadrangles (Ludman, 1977; Pankiwskyj, 1979). They also extend through the Pittsfield quadrangle (Pankiwskyj et al., 1976) into rocks that this author originally mapped as the western facies of the Waterville Formation (Osberg, 1968). A reinterpretation of geologic relationships within the Sangerville Formation and rocks previously mapped as the Mayflower Hill and Vassalboro Formations of Osberg (1968) suggests that these units may be equated, and the older names dropped. It must be noted, however, that rocks assigned to the Sangerville Formation east of the Messalonskee Lake thrust (i.e., former Mayflower Hill and former Vassalboro Formations of Osberg, 1968) do not have stratigraphic continuity with rocks of the Sangerville Formation located to the west of the thrust in south-central Maine. Therefore, this assignment is somewhat tenuous. Both sections contain similar gross lithologies and have similar ages, but the eastern section does not contain limestone units and conglomerate is less abundant than in the western section. The eastern section of the Sangerville Formation is here interpreted as being more distal than the western section.

The belt of Sangerville Formation exposed in the Anson, Skowhegan, and Norridgewock quadrangles (Fig. 2) is 11 km wide at its northern end and about 14 km wide at its southern end. A second belt, parallel to the first, trends southwesterly through the central parts of the Pittsfield and Waterville quad-

rangles and underlies the northwestern half of the Augusta quadrangle. The breadth of outcrop of this belt ranges from 2.5 km in the Pittsfield quadrangle to 16 km in the Augusta quadrangle. Two outliers of the Sangerville Formation occur in the Waterville quadrangle, and an eastern belt of Sangerville Formation extends from Augusta northeasterly to East Benton. This eastern belt of Sangerville Formation has a breadth of outcrop that varies from 5.5 to 7.5 km.

The thickness of the Sangerville Formation cannot be measured within the area of study, because a complete section through the formation is not available. Ludman (1977) and Pankiwskyj (1979) estimated thicknesses in excess of 2000 m in the Skowhegan and Anson quadrangles, respectively.

The Sangerville Formation is represented mainly by wacke intercalated with slate or phyllite. Polymictic conglomerate forms several poorly defined lenses within the wacke sequence; these are more abundant in the northwestern belt of the outcrop than in the southeastern belt. Limestones are tentatively placed in two units on the basis of map patterns and structural interpretations of mesoscopic structures. One, near the upper part of the formation, forms a useful mapping horizon within the Currier Hill synform and to its east, and a second, the Patch Mountain unit (Guidotti, 1965), is thought to be within the middle third of the Sangerville Formation. Rusty-weathered, black sulfidic quartz-mica slate and phyllite occur discontinuously at the base of the Sangerville Formation. Sparse, poorly defined units of rusty-weathered, black quartz-mica phyllite also occur locally throughout the formation, but none of these units has been satisfactorily mapped.

Wacke and phyllite commonly occur as white or light gray, slightly calcareous feldspathic wacke or quartz wacke and gray quartz-mica phyllite or slate in beds 4 cm to 1.5 m thick. The wacke contains subangular clasts of quartz, highly twinned sodic plagioclase, sericitized alkali feldspar, broken and ragged flakes of muscovite, and grains of slate and volcanic material. The wacke generally displays fine sedimentary lamination. In many outcrops the wacke and phyllite occur as distinct beds without gradation, but at some exposures graded couplets are common with proportions of wacke to phyllite ranging from 10:1 to 1:3. Cross-bedding, climbing ripple sets, scour-and-fill features, convolute bedding, and load casts are present, particularly in the low-grade part of the outcrop. Ellipsoidal calcareous concretions occur at several horizons within the formation. Actinolite, calcic plagioclase, garnet, and zoisite appear at middle grades of metamorphism, and diopside occurs at high grades.

The phyllite or slate that forms the upper part of the graded couplets, or which is interbedded with the wacke, is gray or greenish gray and contains fine-grained anhedral quartz, muscovite, and ragged chlorite. Rhombs of ferroan carbonate occur sporadically. Porphyroblastic andalusite, staurolite, and garnet appear in the middle grade of metamorphism, and sillimanite is present at high grade.

Conglomerate occurs both as large unbedded lenses and as layers 1-2 m thick interbedded with the wacke-phyllite assemblage. The clasts within polymictic conglomerates have a continuous range of sizes from 3 cm to 2 mm. Ludman (1977) and Pankiwskyj (1979) indicate that both monomineralic and polymineralic fragments are present, including quartz, feldspar, wacke, quartzite, chert, slate, felsic and mafic volcanics, granite, granodiorite, and hypabyssal plutonic clasts. Dacitic fragments are reported to be the most common (Pankiwskyj, 1979). The matrix consists of quartz, muscovite, and chlorite.

The Patch Mountain unit (Guidotti, 1965) is present at about the middle of the Sangerville Formation. It consists of fine-grained, blue-gray limestone and sandy limestone in beds 2 to 10 cm thick. Cross-bedding is common in the sandy limestone, and load casts develop where the sandy limestone rests on slaty units. Beds of graywacke, gray siltstone, and gray slate, 5 mm to 4 cm thick, are interbedded with the limestone. With increase in metamorphic grade the limestone recrystallizes to marble and calc-silicate minerals develop at the expense of quartz, chlorite, and calcite. The more siliceous interbeds become biotite-rich schist.

The upper limestone unit near the stratigraphic top of the Sangerville Formation is somewhat similar to the Patch Mountain unit and, therefore, will not be described further. Subtle differences include a more shaly character and perhaps thicker bedding style.

The basal unit of black phyllite and other occurrences of sulfidic phyllite occur in outcrops that are heavily coated with limonite and jarosite. The rock is well foliated and is commonly crumpled by minor folds. Quartz, muscovite, chlorite, pyrite, and graphite are the common constituents. At higher metamorphic grade the black sulfidic units contain Mg-rich biotite, pyrrhotite, and locally andalusite or sillimanite.

The Sangerville Formation may be in part a facies equivalent of the Waterville Formation (Osberg, 1968; Ludman et al., 1972; Pankiwskyj et al., 1976). In outlying areas it is correlated with part of the Rangeley Formation of western Maine, the Allsbury Formation in northern Maine, and possibly with part of the Berwick Formation in southern Maine.

Perry Mountain Formation

The name Perry Mountain Formation is used in this report to designate a thin unit that previous workers (Ludman, 1977; Pankiwskyj, 1979) have either included in the upper part of the Sangerville Formation or called Waterville Formation. This unit has lithic and stratigraphic similarity to the Perry Mountain Formation as defined by Moench (in Osberg et al., 1968), although it has not yet and possibly can never be traced into the type Perry Mountain Formation of western Maine. The usage of stratigraphic names adopted here is that of Osberg et al. (1985).

The Perry Mountain Formation is exposed in a thin band in the flanks of the Currier Hill synform (Fig. 2). The Perry Mountain Formation can be traced in a general way (ignoring interruptions by plutons) from the northwest corner of the Pittsfield quadrangle southwest through Skowhegan and into the southwestern corner of the Norridgewock quadrangle, where it forms a complicated pattern in the hinge of the Currier Hill synform

(Fig. 2). From this hinge area it can be traced northeastward, with interruptions by plutons, through the northwest part of the Waterville quadrangle into the central part of the Pittsfield quadrangle.

The thickness of the Perry Mountain Formation is 50 to 100 m based on the minimum breadth of outcrop.

The Perry Mountain Formation contains lithic variants that are similar to parts of the Perry Mountain Formation of Moench (in Osberg et al., 1968). It consists of beds of quartzite from 1 to 8 cm thick alternating with beds of quartz-mica phyllite from 1 to 3 cm thick. The quartzite beds are thickest in the northwest flank of the Currier Hill synform and toward the top of the formation. Cross-lamination and convoluted features are common in the thicker beds of quartzite. Sparse, lenticular carbonate-rich pods, measuring up to 7 cm thick and 17 cm long, occur in the upper part of the formation. The pelites of the Perry Mountain Formation are sufficiently aluminous to produce garnet, staurolite, andalusite, and sillimanite at higher metamorphic grades.

That part of the Perry Mountain Formation in the southeastern limb of the Currier Hill synform has been correlated previously with the Waterville Formation (Pankiwskyj et al., 1976) on the assumption that the two formations coalesce in the east limb of the Currier Hill synform just north of this area of study. Ludman (1976) and Pankiwskyj et al. (1976) interpreted this relationship as a sedimentary facies, with the Sangerville Formation thinning to a feather edge between the coalescing Waterville and Perry Mountain Formations. The cutting out of lithic units along the western boundary of the Waterville Formation, however, argues for this contact being a thrust instead of a facies contact. This thrust, the Messalonskee Lake thrust, cuts out the Sangerville Formation and juxtaposes the Waterville and Perry Mountain Formations just north of this area of study. These local relationships, however, do not necessarily invalidate the concept of a regional facies as envisaged by Ludman (1976) and Pankiwskyj et al. (1976) between the Waterville, Sangerville, and Perry Mountain Formations (see Fig. 3).

The Perry Mountain Formation to the south is possibly correlative with unit 2 of Eusden et al. (1984), formerly mapped as upper Rindgemere Formation (Hussey, 1968; Eusden et al., 1984) in southern Maine and adjacent New Hampshire.

Smalls Falls Formation

The Smalls Falls Formation includes generally rusty-weathered, sulfidic, quartz-rich sandstones, pelites and uncommon conglomerates overlying the Perry Mountain Formation (Osberg et al., 1968). Pankiwskyj et al. (1976) used the name Parkman Hill Formation for the same rocks, but in this paper the usage of Osberg et al. (1985) is followed.

The Smalls Falls Formation forms a narrow outcrop, parallel to that of the Perry Mountain Formation, outlining the Currier Hill synform (Fig. 2) in the Pittsfield, Skowhegan, Waterville, and Norridgewock quadrangles.

The thickness of the Smalls Falls Formation is approximately 900 m at its type locality (Osberg et al., 1968). Ludman (1976) estimated its thickness in the west limb of the Currier Hill synform to be from 300 to 900 m. Estimates of the thickness of the Smalls Falls Formation in the east limb of the Currier Hill synform range between 100 and 200 m. These estimates suggest that the Smalls Falls Formation thins from western Maine toward the southeast.

In the west limb of the Currier Hill synform (Fig. 2) the Smalls Falls Formation consists dominantly of rusty-weathered, sulfidic quartz arenite interbedded with rusty, black sulfidic mica phyllite, and in the east limb the dominant rock is rusty, black sulfidic phyllite.

Dark gray to light gray sulfidic quartz arenite, quartz wacke and quartz-rich siltstone occur as beds 10 cm to 2 m thick. All are to some extent rusty-weathered. Most have been sufficiently recrystallized so that the detrital character of the grains is not obvious, although Pankiwskyj (1979) noted sparse, rusty, quartz-bearing conglomerates.

The rusty-weathered, black, sulfidic mica phyllite is the dominant lithology of the Smalls Falls Formation. Most outcrops are heavily coated with limonite and jarosite, so that primary features are difficult to detect, but in some well-washed stream outcrops lamination can be observed. Andalusite and sillimanite in the phyllites appear at middle and high grades, respectively. A crumpled schistosity is common in these phyllites.

Pankiwskyj et al. (1976) correlated the Smalls Falls Formation with the discontinuous sulfidic phyllite at the bottom of the Sangerville Formation. This correlation was undoubtedly based on the assumption that the Perry Mountain and Waterville Formations are equivalent, an assumption now thought to be in error (see previous discussion).

Regionally, correlatives of the Smalls Falls Formation in northern Maine have not been separately mapped. To the south it may be equivalent to unit 3 of Eusden et al. (1984), and it is presumed to correlate with the Francestown Formation in New Hampshire (Robinson, 1981; Hatch et al., 1983).

Madrid Formation

The name Madrid Formation was established by Moench (in Osberg et al., 1968) for generally massively bedded wackes and minor pelite in western Maine. The type section is in the village of Madrid in the Phillips quadrangle. Rocks assigned to the Madrid Formation in south-central Maine cannot be traced into the type section because of intervening faults, but their lithic and stratigraphic similarity warrants the use of the name. Pankiwskyj et al. (1976) used the name Fall Brook Formation for these rocks, and Ludman (1976) used the names Brighton Quartzite and Brighton Formation for the same rocks.

The Madrid Formation is exposed in the core of the Currier Hill synform (Fig. 2). This outcrop extends with interruptions by plutons from the north-central part of the Pittsfield quadrangle southwestward through the Skowhegan and the Waterville quadrangles into the Norridgewock quadrangle.

No thickness can be estimated for the Madrid Formation as its upper contact is not exposed in the area of this study. Pankiwskyj (1979) states that its thickness is about 1000 m in the Anson quadrangle, whereas Moench (in Osberg et al., 1968) estimated a thickness of about 300 m in western Maine.

The Madrid Formation generally consists of thickly bedded, slightly calcareous quartz wacke interbedded with substantially less abundant gray phyllite. Beds of quartz wacke have thicknesses that range from 15 cm to 4 m, and beds of phyllite are commonly 3 to 15 cm thick. Beds are commonly ungraded but contain thin sedimentary laminations. Locally, wacke grades into phyllite through thin interlaminations of siltstone and phyllite with the proportion of phyllite increasing in each lamination toward the top of the bed. Cross-lamination is conspicuous in some beds and ellipsoidal, calcareous concretions are locally common. With increase in metamorphic grade biotite, plagioclase, clinozoisite, garnet, actinolite, and diopside may be present.

Gray phyllite that is interbedded with the wacke consists of quartz, plagioclase, muscovite, chlorite and a trace of ferroan carbonate. Grain size is uniformly fine, ranging from 0.01 to 0.03 mm. The muscovite and chlorite have good alignment and define a generally excellent schistosity. This rock is fairly insensitive to metamorphic reconstitution.

The thickly bedded character of the Madrid Formation is locally broken by packets of thinly bedded quartz-rich siltstone and greenish gray phyllite, up to a few meters thick. Thicknesses of individual beds within these packets range from 1 to perhaps 10 cm. The beds within these packets at some places are graded from siltstone into phyllite, but more commonly the beds have sharp contacts at both top and bottom of the siltstone beds with adjacent phyllite. Locally, the siltstones exhibit sedimentary lamination and cross-bedding. At higher grades biotite and local garnet may be present.

The Madrid Formation may correlate to the north with the Jemtland Formation in the Presque Isle area (Roy and Mencher, 1976). It is possibly equivalent to the Warner Formation in New Hampshire (Robinson, 1981).

HYPABYSSAL AND PLUTONIC ROCKS

Hypabyssal Rocks

The oldest hypabyssal rocks recognized in south-central Maine are metamorphosed andesites and possibly dacites. Dikes of these hypabyssal rocks have been observed in the Waterville, Burnham, Pittsfield, and Stetson quadrangles, and Ludman (1977) has described small stocks containing metamorphosed andesite in the Skowhegan quadrangle (Fig. 2). The dikes are less than 1 to 3.5 m thick and of unknown length. Most are nearly parallel to bedding in the limbs of isoclinal folds, although at a few localities they cut bedding at a high angle. All contacts, where examined in detail, have local crosscutting relationships with the enclosing stratified rocks. These dikes cut across isoclinal folds (F_2) and are folded by small asymmetric folds (F_3). They also display cleavage that parallels the axial surfaces of F_3 and are locally boudinaged.

The hypabyssal dikes and stocks are metamorphosed to plagioclase granofels, the texture and mineralogy of which varies with the grade of metamorphism. Ludman (1977) described a metamorphosed andesite from the central part of the Skowhegan quadrangle. It consists of plagioclase, antigorite, chlorite, and calcite. Plagioclase (An_{35}) is the major constituent and occurs both as relict phenocrysts and in the groundmass. Pseudomorphs of amphibole by calcite and antigorite have well defined shapes. The matrix consists of calcite and chlorite in addition to plagioclase. Osberg (1968) has described somewhat similar rocks, but with a more extensive metamorphic fabric from the vicinity of Waterville (Fig. 2).

The age of the plagioclase-rich granofels dikes and stocks is probably Early Devonian. In south-central Maine they intrude rocks as young as Wenlockian and postdate isoclinal folds. The isoclinal folds deform rocks possibly as young as Siegenian north of this area of study. Thus, these dikes and small stocks must be younger than Siegenian. They are metamorphosed by an event that Dallmeyer (1979) suggests is coeval with the emplacement of the Hartland pluton, which is placed at 360 ± 8 Ma (Dallmeyer et al., 1982).

A few basaltic dikes have been observed in the area of study. These truncate all of the rocks of sedimentary origin, but their relationship to the plutons is unknown. They have widths up to a meter and lengths measured in several tens of meters. They have chilled margins and most have well developed cross joints. Although fine-grained, they exhibit ophitic texture with laths of calcic plagioclase enclosed in augite. Chloritic patches may represent pseudomorphs after olivine (Ludman, 1977). The intrusion of these rocks postdates the metamorphic aureole of the Hartland pluton, and, therefore, their age is younger than 360 ± 8 Ma. They could be as young as Mesozoic.

Plutonic Rocks

Slightly foliated granodiorite is exposed in the Hartland pluton in the Skowhegan quadrangle, in the Norridgewock pluton in the Norridgewock quadrangle, and in the Three Mile Pond pluton in the Vassalboro quadrangle (Fig. 2). Although these plutons are not homogeneous texturally or mineralogically, they are essentially composed of medium- to coarse-grained, inequigranular-seriate granodiorite. The margins of the plutons tend to be better foliated than the more central parts. Dioritic phases are associated with the Three Mile Pond pluton, and Ludman (1977) reports that the Hartland pluton locally carries muscovite-bearing phases at its margin.

The granodiorite contains plagioclase, quartz, microcline, biotite, and locally hornblende and/or garnet. Chlorite appears as a common alteration product of biotite. The preferred orientation of biotite produces a weak to strong foliation in these rocks. Plagioclase, showing polysynthetic twinning, and microcline, showing microcline twinning and perthitic intergrowths, occur as grains 0.6 to 1.6 mm in size. Myrmekitic

intergrowths are present where plagioclase and microcline are in contact. The abundance of hornblende varies from 0 to 3%.

The texture in the Three Mile Pond pluton differs subtly from the Norridgewock and Hartland plutons. It contains microcline with a greater extent of perthitic intergrowth, particularly near its margins, and the development of mortar texture.

The granodioritic plutons intrude all of the stratigraphic section exposed in south-central Maine. All cut F_2. Dallmeyer and Van Breeman (1981) have defined a whole rock Rb/Sr isochron age of 381 ± 14 Ma for the Three Mile Pond pluton, and Dallmeyer et al. (1982) have defined a whole rock Rb/Sr isochron age of 360 ± 8 Ma for the Hartland pluton. The other plutons of granodiorite have not been dated.

Muscovite-bearing granite occurs in the Old Point pluton (Pankiwskyj, 1979) at the south margin of the Anson quadrangle, in the David Pond and Lord Hill plutons in the southwestern corner of the Norridgewock quadrangle, in the Hallowell pluton and associated bodies (Barker, 1961), and in the Togus pluton and associated stocks (Fig. 2). Equigranular binary granite is common to the Old Point, Lord Hill, Hallowell, and Togus plutons, whereas the David Pond pluton contains binary granite in which the muscovite occurs in large diamond-shaped books.

The binary granite is equigranular to inequigranular-seriate, with a grain size that ranges from 0.3 to 1.4 mm. Oligoclase, microcline, and quartz form interlocking grains. The plagioclase tends to exhibit subhedral form and to occur as the largest crystals in the rocks. Microcline forms equant grains with irregular margins; it contains patch perthite intergrowths. The feldspars have a strong mortar fabric in the Hallowell and Togus plutons. Quartz occurs as anhedral grains. Muscovite and biotite are subhedral and have a faintly developed preferred orientation. Chlorite is a common alteration product of biotite, and many of the plagioclase crystals are saussuritized. Garnet is a common accessory.

The binary granite near Togus is younger than the granodiorite, as the Togus pluton contains inclusions of granodiorite. Plutons of binary granite intrude the Silurian sedimentary section, transect F_2, and contain inclusions of country rock that are cut by thin dikes of binary granite that are folded and cleaved by F_3. Dallmeyer and Van Breeman (1981) have published whole rock Rb/Sr isochron dates for the Hallowell and Togus plutons of 387 ± 11 Ma and 394 ± 8 Ma, respectively.

STRUCTURAL GEOLOGY

General Statement

Prominent structural features in south-central Maine include schistosity and cleavage, various groups of folds, and faults. These features occur at various scales, from a few millimeters in size, seen in thin sections as crumples and off-sets in foliation, through outcrop-sized features, to those measured in tens of kilometers. These features have a diversity of orientations.

Three ages of structural features have been recognized: late asymmetrical folds (F_3) and associated features; upright, mostly isoclinal folds (F_2) and associated features; and early recumbent, isoclinal folds (F_1) and associated structures. Late cleavage bands and boudinage are less prominent, and the limited observations of them preclude any valid analysis of these features.

In addition, a pre-F_2 thrust, the Messalonskee Lake thrust, has been identified. Two high-angle faults postdate major ductile deformation.

F_3 and Associated Structural Features

More-or-less open, asymmetrical folds are widely distributed in south-central Maine. These folds deform bedding, schistosity, and early cleavages. In general, their sizes are measured in the range of 10 to 20 cm and they are sufficiently small that they do not control the distribution of lithic units on the map. Most of these folds have a right-handed sense and have amplitude to wavelength ratios of 0.3 to 0.5, although in the southern part of the area they are slightly more appressed. Figure 4 shows the character of these folds.

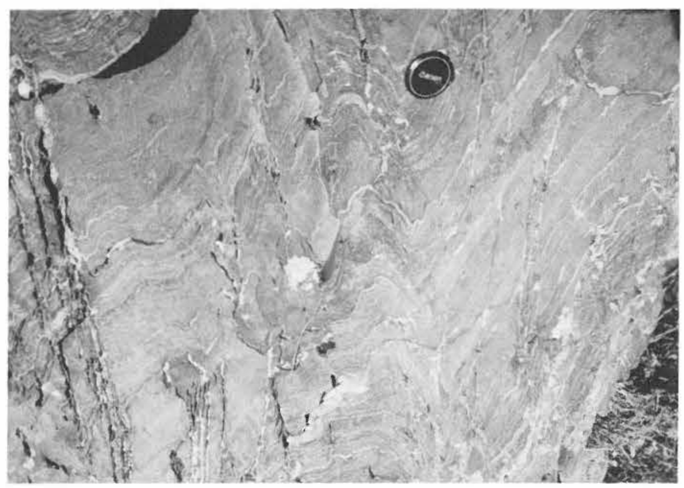

Figure 4. Asymmetrical folds (F_3), Winslow.

As shown in Figure 5a, the axial surfaces of these asymmetrical folds strike consistently about N10°E and dip steeply toward the east. The common orientation of these features indicates that they are a late feature in the structural history and are not folded in any pervasive way by other folds.

The axes of F_3 typically plunge steeply, but they plot on the great circle that marks the intersection of F_3 axial surfaces with bedding and preexisting schistosity (Fig. 5b). The steep inclination of these axes indicates that they were superimposed on surfaces that had steep dips, and their right-handed character indicates that the steeply dipping surfaces had rather uniform orientations before this folding event.

A spaced cleavage (S_3), approximately parallel to the axial surfaces of F_3, is well developed throughout the region. Poles to these cleavage surfaces are indicated in Figure 5c. The distance between cleavage surfaces is 1 to 8 mm. At low meta-

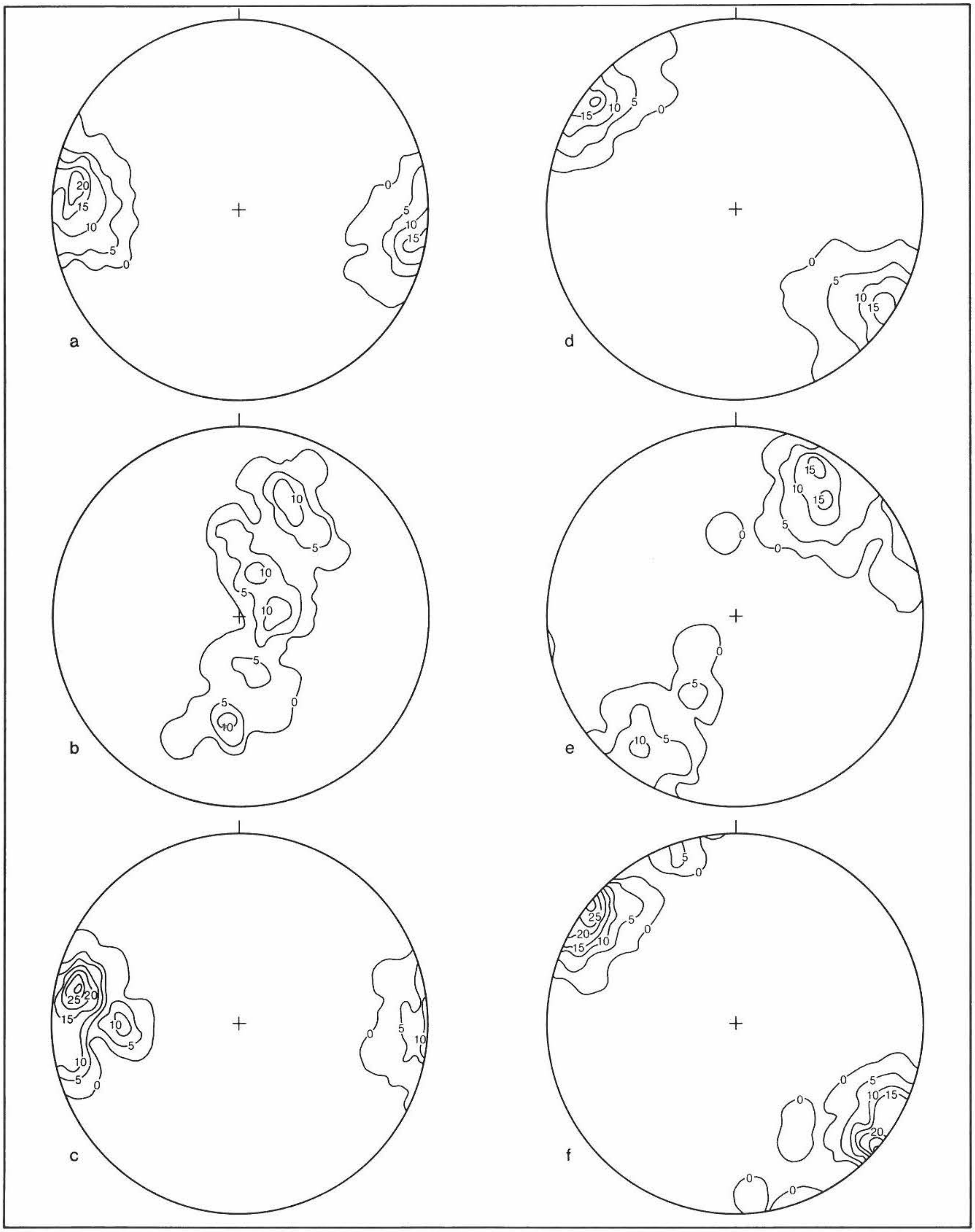

Figure 5. Orientation of structural elements. Diagrams are lower hemisphere, equal-area projections. (a) Poles to axial surfaces of F_3; 50 points. (b) Axes of F_3; 50 points. (c) Poles to S_3; 140 points. (d) Poles to axial surfaces of F_2; 65 points. (e) Axes of F_2; 65 points. (f) Poles to S_2; 490 points.

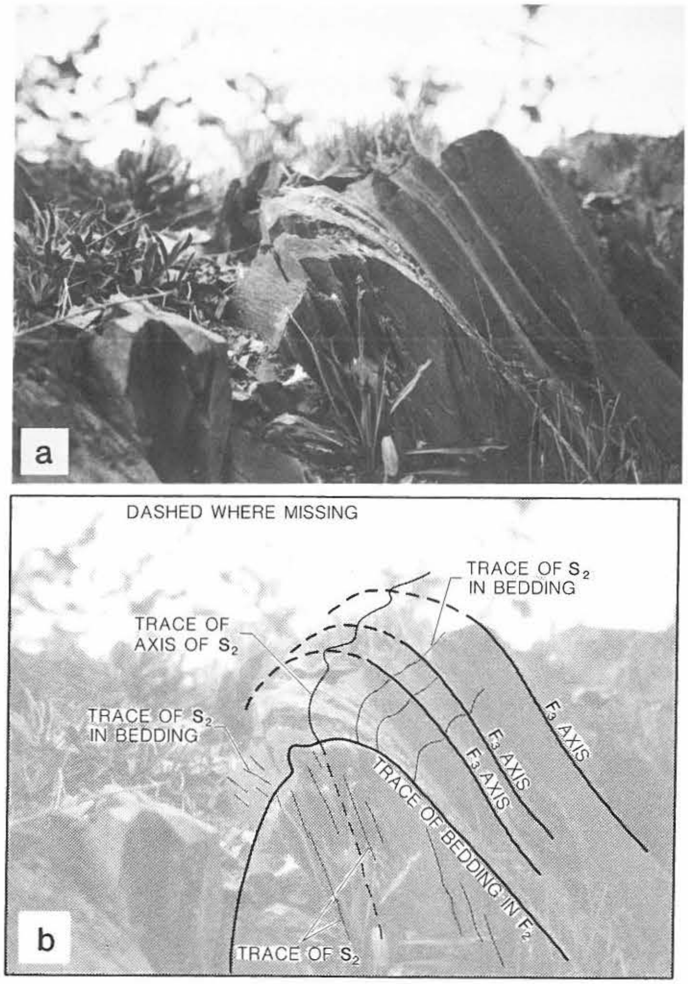

Figure 6. (a) Upright isoclinal fold (F_2) deformed by F_3, Route 137, 4.2 km east of Winslow. (b) Line sketch of photo in (a) which identifies various structural features.

morphic grade some micaceous minerals grow along these surfaces and at higher grades sillimanite grows along them.

F_3 deforms isoclinal F_2 (Fig. 6) and deforms schistosity that is related to F_2. F_3 is also younger than the metamorphosed andesitic or dacitic dikes. In addition, F_3 is found in inclusions in peraluminous granite of the Hallowell pluton. A small outcrop located on Route 95, 0.3 km north of the Augusta interchange, contains quartz-plagioclase-biotite-calc-silicate granulite that is cut by a thin dike of leucogranite (Fig. 10a). The dike is folded and cleaved by F_3, indicating that F_3 is younger than the Hallowell pluton dated at 387 ± 11 Ma (Dallmeyer and Van Breeman, 1981).

F_2 and Associated Structural Features

Upright isoclinal folds are the most important fold element in the region. F_2 folds, because they are significantly larger than F_3 folds, control the lithic distribution of formations on the geologic map (Fig. 2), and where mapping horizons are well defined, the axial traces of F_2 folds can be confidently drawn. The abundance of mesoscopic F_2 folds is much higher in the axial regions of map-sized F_2 folds than it is in their limbs. These folds deform bedding and possibly a faint schistosity. They have amplitude-wavelength ratios that range from 0.5 to 1.0. Bedding thicknesses measured in sections perpendicular to their fold axes are plotted on Figure 7 against the dips of bedding (Ramsay, 1967) and suggest that these are flattened folds. The square root of the ratio of the intermediate to largest principal extensions ranges from 0.6 to 0.15. Figure 8 shows the style of these folds.

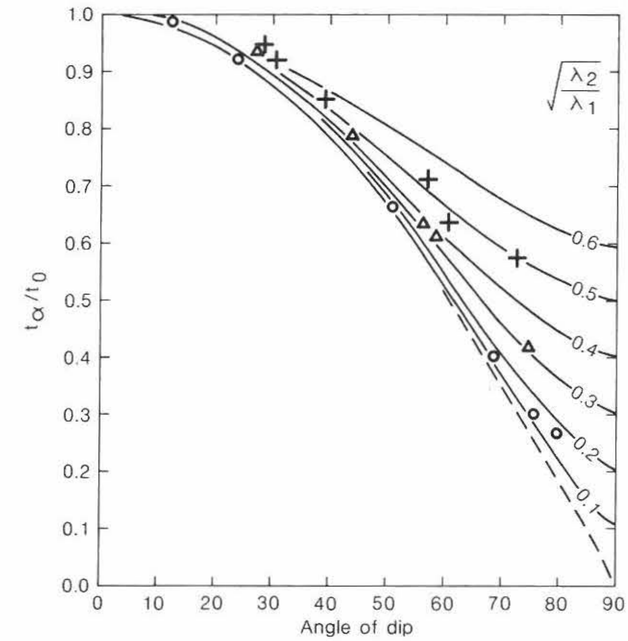

Figure 7. Flattening of F_2 (after Ramsay, 1967). + = thin bedded quartzite and pelite; Δ = wacke; o = limestone. λ_1 = largest principal extension, λ_2 = intermediate principal extension, t_0 = normal thickness at hinge, t_α = normal thickness at dip α, α = dip of bed.

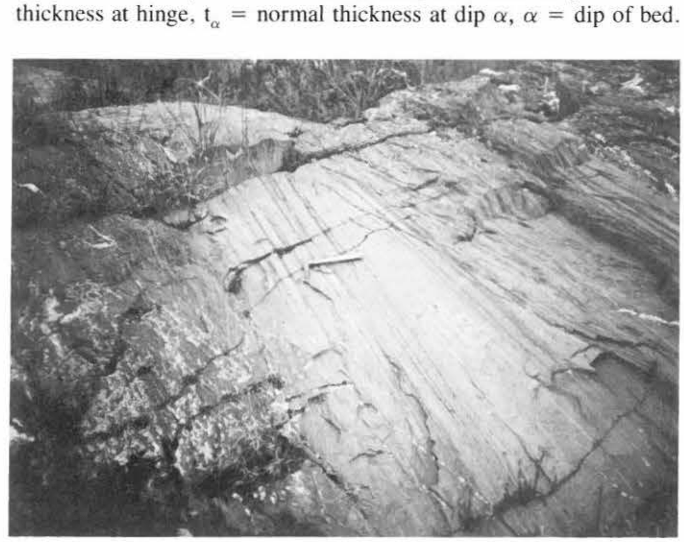

Figure 8. Style of upright isoclinal folds (F_2), Winslow.

The axial surfaces of F_2 folds have constant orientations over large areas (Fig. 5d). Their dips are consistently a few degrees from vertical.

The axes of the upright folds have plunges generally less than 35°. Most plunge to the northeast, but some plunge to the southwest. Representative plunges for these folds are shown on Figures 5e and 12.

A penetrative schistosity in pelitic rocks is parallel to the axial surfaces of F_2 folds. Small plates of muscovite and chlorite and the ellipsoidal shapes of non-micaceous crystals are aligned in this schistosity. Over much of the area the schistosity is sensibly parallel to bedding, however in the hinges of F_2 folds this schistosity cuts across bedding.

In more massive beds of wackes and limestone, pressure solution cleavage is common (Fig. 9). This cleavage occurs as zones, 3 to 7 mm thick, in which biotite and accessory minerals are concentrated. The micaceous minerals within these zones have an excellent preferred orientation parallel, or nearly parallel, to the margins of the zones. The cleavage zones separate microlithons, 1 to 1.5 cm thick, of normal rock. Cleavage surfaces have a slightly anastomosing arrangement, but all are aligned nearly parallel to the axial surfaces of F_2 folds. Poles to schistosity and pressure solution cleavage are plotted in Figure 5f.

Figure 9. Pressure solution cleavage (S_2), Route 137, 4.2 km east of Winslow.

The schistosity and pressure solution cleavage of S_2 may deviate locally from parallelism with the axial surfaces of F_2 folds. Where beds of differing competency are interbedded, the schistosity may exhibit differences in strike of 10°-15° and differences in dip of as much as 20°. In addition, the orientation of S_2 may have a fan-like arrangement in the hinges of F_2 (Fig. 10b). In one outcrop (Fig. 10c) the schistosity adjacent to a quartzite bed displays orientations predicted for zones of contact strain (Ramsay, 1967, p. 417). The schistosity throughout the outcrop is essentially parallel to the axial surface of F_2, but in a zone a few millimeters thick at the convex side of the bed of quartzite, the schistosity is parallel to bedding. A finite neutral point must exist in the rock between the zone in which schistosity parallels bedding and where schistosity parallels the axial surface. At the concave side of the quartzite bed, the schistosity is everywhere parallel to the axial surface.

F_2 and its associated structural features are older than F_3 and its related features. Cleavage associated with F_3 commonly cuts across both limbs of F_2 without deflection, and at several localities the axial surfaces and limbs of F_2 folds are folded by F_3 (Fig. 6). F_2 is older than all intrusive bodies in the region, including the metamorphosed andesitic and dacitic dikes, the granodiorite, and the muscovite-bearing granites. F_2 is also either synchronous with or slightly predates the earliest metamorphic event in the region.

Outcrop-sized F_2 folds that deform bedding and show primary facing directions do not everywhere have stratigraphic directions consistent with anticlines and synclines. At some places, antiforms contain beds whose facing directions indicate younger rocks are in the core of the fold, and at other places synforms contain beds whose facing directions indicate that older rocks form its core. These relationships suggest that an earlier fold event predated F_2.

F_1 Folds

Evidence for an earlier set of folds (F_1) is mostly indirect, consisting of a few minor structural features and inverted stratigraphic sequences. Minor structural features indicating an early folding event are uncommon. A single outcrop, located on the east bank of the Kennebec River 0.2 km south of the Waterville-Winslow bridge, contains an isoclinally refolded, isoclinal fold. Figure 10d shows the relationship. The younger isoclinal fold has a style and orientation that identifies it with F_2. The earlier isoclinal fold has a "hooked" shape, and both of its limbs are folded by F_2. A faint pressure-solution cleavage is parallel to the axial surface of the early fold and this cleavage is also folded by F_2. The early fold form is thought to be F_1 and to represent an early isoclinal fold. The plunge of the early fold is uncertain.

In addition, a few isoclinal folds with abnormally steep plunges have been observed throughout the area of this study. The style of these folds is much like that of F_2. It is only their plunge that distinguishes them. They have axial surfaces that are appropriate for F_2, but their plunges vary from 65° to 80°. The style and orientation of these folds is consistent with their being F_1 folds preserved in the limbs of F_2 folds.

Another indication of earlier folds (F_1) is the presence of downward-facing F_2 folds (Fig. 12). Downward- and upward-facing F_2 folds have been recognized by examining primary sedimentary features, e.g., cross-bedding and graded bedding in their hinges. If the facing directions indicate that younger beds are met progressively upward along the axial surface (Fig. 11a and b), the fold is upward-facing and the section is normal. But if the facing directions indicate that younger beds are met progressively downward along the axial surface (Fig. 11c and d), the fold is downward-facing and the section is invert-

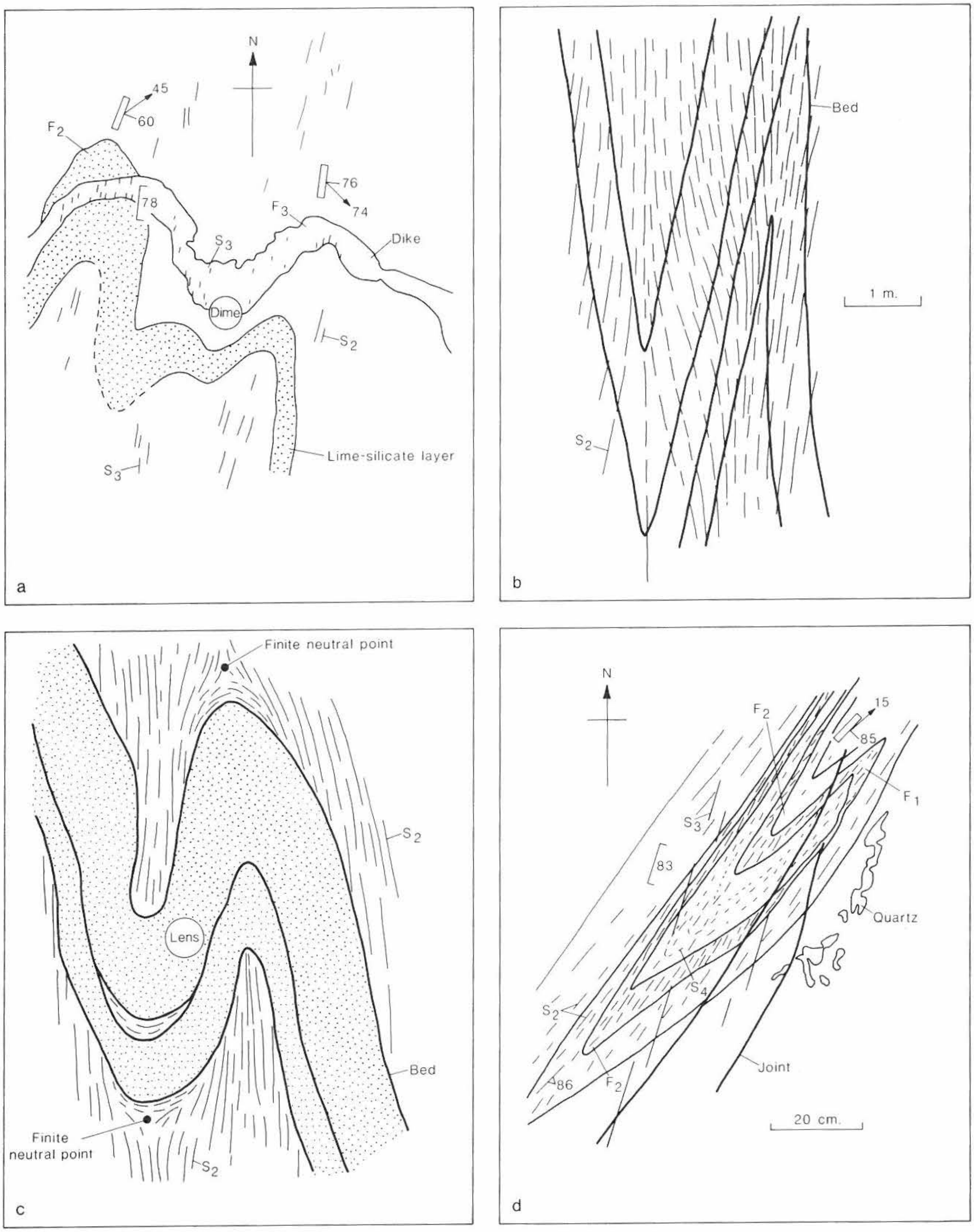

Figure 10. Sketches of mesoscopic structural features. (a) Folded relationships of granitic dike and bedded calc-silicate-bearing granulite. Intersection of Routes 95 and 11, Augusta. (b) Fanned cleavage associated with F_2 in limestone member of Sangerville Formation. Intersection of Messalonskee Stream and Routes 137 and 11, Oakland. (c) Contact strain relations in F_2. Intersection of Routes 2 and 152, West Palmyra. (d) Isoclinal F_1 folded by isoclinal F_2. East bank of Kennebec River, Winslow.

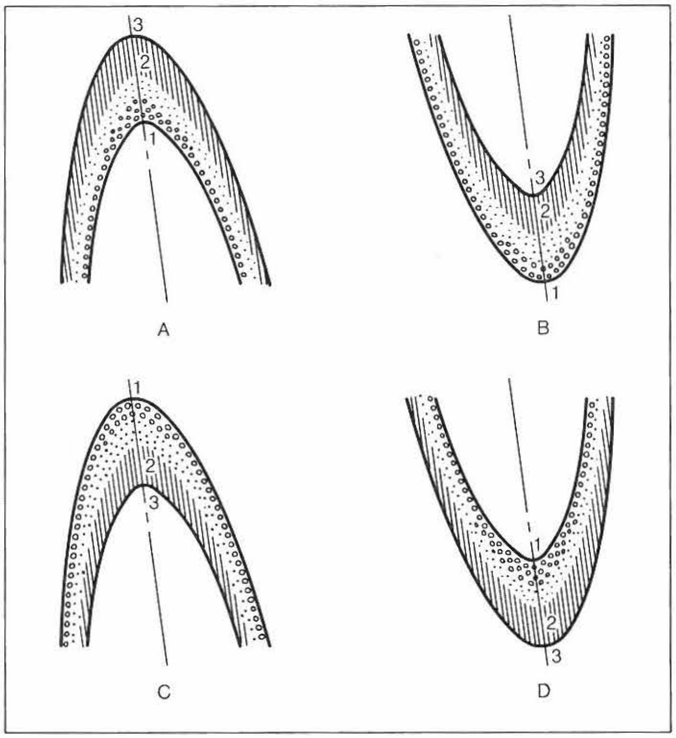

Figure 11. Facing directions of folds. (a) Upward-facing anticline. (b) Upward-facing syncline. (c) Downward-facing antiform. (d) Downward-facing synform. Coarse patterning is at bottom of bed. Age of beds, 1 > 2 > 3.

ed. Because schistosity in F_2 is parallel to the axial surface of these folds, observations of facing directions of sedimentary features on schistosity provides information on the facing directions of F_2 as well (see Shackleton, 1958; Borradaille, 1976). Of course in terranes of isoclinal folding the bedding-schistosity relations are generally meaningful only near the hinge regions of folds because of the parallelism of bedding and schistosity in the limbs.

A good example of a downward-facing antiform may be observed on the east side of the Kennebec River 200 m south of the Waterville-Winslow bridge, and the downward-facing traces of graded beds are well preserved on cleavage related to F_2 1.2 km northeast of Benton Station (Fig. 2).

In south-central Maine observation of downward-facing F_2 and therefore, inverted sections have been made at several localities. These localities are indicated in Figure 12. The distribution of these observations is taken to mean that a part of the Waterville Formation is inverted and that some structural feature existed prior to F_2 to invert the section. Axial traces of early folds (F_1) have been drawn in Figure 12 as lines separating regions of upward-facing F_2 from those of downward-facing F_2. Regions underlain by the inverted limbs of F_1 folds are patterned while regions underlain by normal limbs of F_1 folds are unpatterned, as shown in the inset of Figure 12.

No local information is known about the facing directions of F_1.

Cleavage associated with F_1 is difficult to identify. However the outcrop in Winslow, showing F_1 folded by F_2 (Fig. 10d), does exhibit a cleavage that was formed as an ancillary structure to F_1. Moreover, the observation that in the hinges of some F_2 folds a weak foliation is folded with the bedding indicates the presence of an earlier cleavage. It is concluded that any relict cleavage connected to the F_1 event would approximately parallel bedding because of the isoclinal character of F_1, and this cleavage would tend to be misidentified as S_2 after the F_2 event because F_2 is also isoclinal.

Map-scale Folds

Mapped folds are principally F_2 folds, although a consideration of upward- and downward-facing sequences leads to the delineation of F_1 folds as well. The axial traces of these structural features are shown in Figure 12. F_2 features will be described first, followed by a description of F_1 features.

The large tract of Sangerville Formation east of Waterville and extending northeasterly from Augusta contains F_2 and associated cleavage which when combined with facing directions of beds indicate that F_2 structures face upwards. These relationships indicate that this tract of the Sangerville Formation is in a normal stratigraphic sequence and occupies the core of an F_2 synform, the Vassalboro synform.

The outcrop of Waterville Formation to the west is antiformal based on the plunges of mesoscopic F_2 and the symmetry of stratigraphic units. Its axial trace is shown in Figure 12, and this structure is designated the Waterville-Newport antiform. Facing directions within this antiform are complex and will be discussed under the F_1 folds.

The two "canoe"-shaped outcrops of Sangerville Formation west of Waterville define F_2 synforms (Fig. 12). The axes of mesoscopic F_2 folds near their northern terminations plunge gently toward the southwest, and near their southern extremities they plunge northeasterly. Moreover, the facing directions of beds in F_2 structural features indicate that these features face upward. The large areal extent of the Waterville Formation north and south of the these synforms (Figs. 2 and 12) occupies culminations in the F_2 folds.

The tract of Sangerville Formation between the Messalonskee Lake thrust and the Currier Hill synform (Fig. 12) contains mesoscopic F_2 folds that generally plunge toward the northeast. These folds are left-handed in the eastern part of the tract and right-handed in its western part, indicating a large F_2 antiform. The trace of this map-scale F_2 fold is drawn to separate the left- and right-handed mesoscopic folds and to separate the two belts of limestone (Fig. 12). The major fold is called the Canaan-Readfield antiform. The geometry of this structure requires the limestone unit exposed east of the Currier Hill synform to belong to the upper limestone unit of the Sangerville Formation.

The Currier Hill synform (Fig. 12) is an F_2 structure. Mesoscopic F_2 axes plunge mostly toward the northeast, and the facing directions of beds in these folds indicate that they

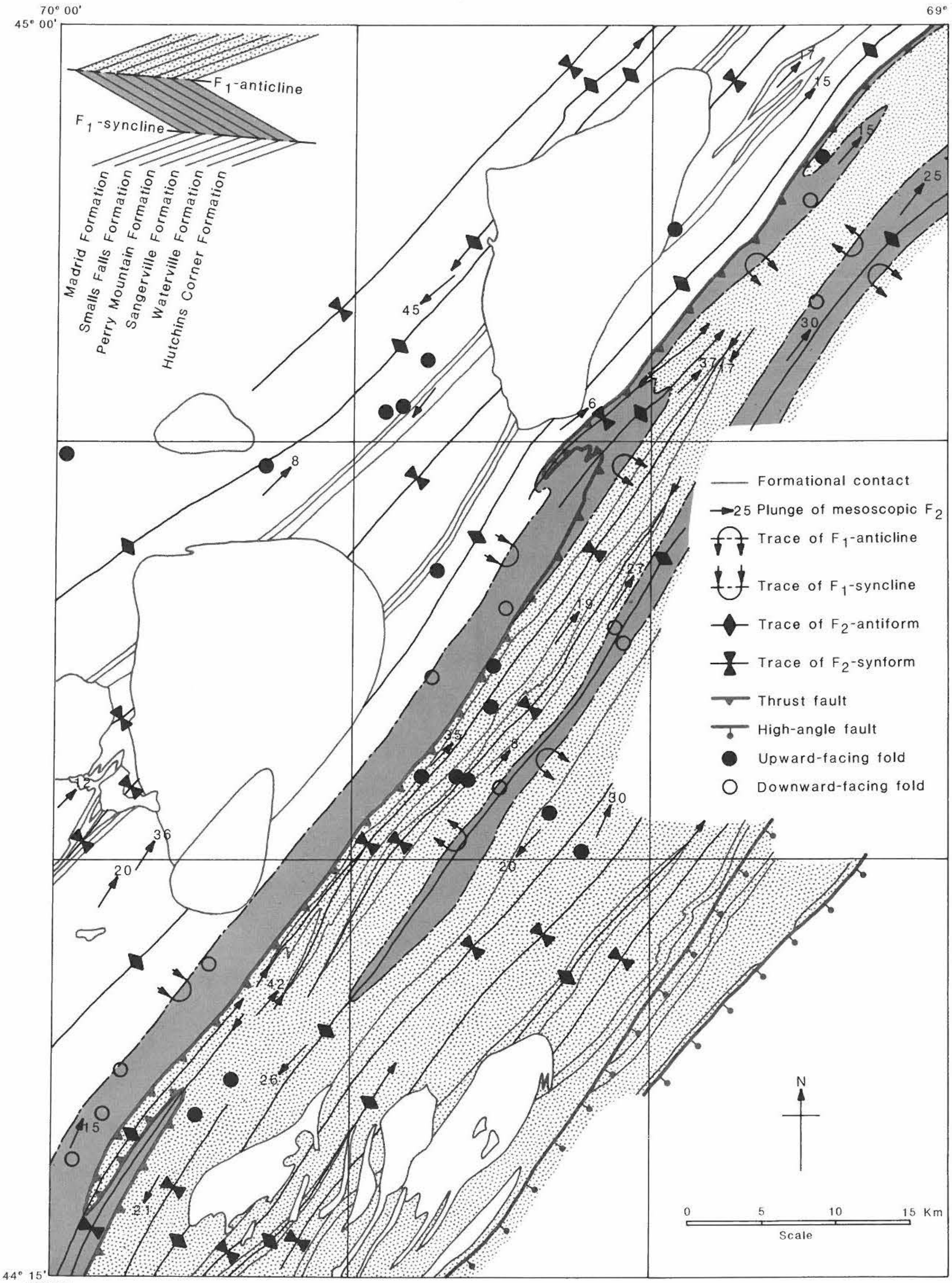

Figure 12. Map showing axial traces of F_1 and F_2 folds. Locations of upward-facing and downward-facing folds are shown by open and closed circles. Plunges of F_2 are shown by arrows. Inset shows patterning scheme.

face upward. The relationships that hold for the mesoscopic folds are taken to be true for the map-scale fold as well.

The Cornville antiform (Fig. 12) to the west was first recognized by Pankiwskyj et al. (1976). It is outlined by the Patch Mountain unit of the Sangerville Formation. This structure is upright and nearly isoclinal, both characteristics of F_2. Mesoscopic F_2 axes plunge to the northeast, and the facing directions of beds within these mesoscopic structures indicate that they face upward. As a consequence, the map-scale fold is interpreted to face upward and within the region of study to plunge northeastward.

The large area of Sangerville Formation west of the Cornville antiform must lie in a synformal structure. This structure contains many small and large folds, and the isolated outcrops of limestone in the vicinity of Athens and Harmony (Fig. 2) probably lie in the cores of F_2 antiforms. Mesoscopic folds at Athens face upward, so that the larger fold is assumed to do so also. In contradistinction, Mulry (in press) indicates that at its southern extremity this synformal tract faces downward, suggesting that an earlier fold axial trace separates the two areas.

At least three large F_1 recumbent folds are interpreted to exist within the map area. The most westerly of them is exposed in the east limb of the Canaan-Readfield antiform (Fig. 12). Here the facing directions of F_2 folds at several outcrops face downward and indicate that the section is inverted. The axial trace of an F_1 fold is drawn within the eastern two belts of limestone and just to the west of those localities exhibiting downward-facing folds. Because the geometry of the F_2 Canaan-Readfield antiform requires the limestone to be the upper limestone unit of the Sangerville Formation, the F_1 fold must be synclinal and face west. It is truncated by the Messalonskee Lake thrust southeast of Canaan.

The most eastern of the F_1 folds is exposed in the Waterville-Newport antiform (Figs. 2 and 12). Within the core of the Waterville-Newport antiform, F_2 structural features face downward, indicating that the stratigraphy is inverted, but elsewhere in this antiform, F_2 structural features face upward, indicating normal stratigraphy. A line in Figure 12 separating normal and inverted stratigraphic sections delineates the folded trace of the F_1 axial surface. This recumbent fold is thought to pass beneath the easternmost of the "canoe"-shaped F_2 synforms west of Waterville and possibly to re-emerge in the thin band of Waterville Formation that lies immediately to its west (Fig. 12), although evidence for the position of its axial trace there has not been observed. The facing direction for this fold is uncertain, but because the F_1 fold to the west verges west, this fold is also assumed to do so.

The axial trace of a third F_1 recumbent fold is thought to exist in a narrow belt of the Waterville Formation along the east side of the Messalonskee Lake thrust (Figs. 2 and 12). This F_1 fold is defined by the facing directions of F_2 folds northeast of Canaan, but to the south it is undefined, and it could be cut out by the Messalonskee Lake thrust. The data set for this F_1 fold gives no information on its vergence, but because the F_1 fold to the west of the Messalonskee Lake thrust faces west, this fold is also drawn so as to face west.

Beds at the southeastern extremity of the Currier Hill synform have abnormally shallow dips, suggesting that an F_1 feature is folded into this part of the F_2 structure, but their geometry cannot be reconstructed given the current observations.

The vergence for the F_1 folds described in this report does not conform to the vergence of early folds described in outlying areas (Hussey et al., 1984; Eusden et al., 1984; Mulry, in press). Of course, all of the facing directions for these folds, except those decribed by Hussey et al. (1984), are based on somewhat intricate and tenuous arguments, but assuming that they have been correctly identified, the differences in facing directions can be explained by two different folding events (Osberg et al., in press). The first event involved recumbent folding and thrusting with westward transport, and the second involved back-folding with eastward vergence.

Faults

Two high-angle, post-F_3 faults and an early thrust, the Messalonskee Lake thrust, are recognized in the area of study (Figs. 2 and 12).

The two high-angle faults have been mapped in the eastern part of the area. The Dearborn Brook fault (Pankiwskyj, 1976) is located along a pronounced topographic lineament that extends northeasterly from the vicinity of South Windsor (Fig. 2). There may be several splays, and as a result several different positions have been taken for this fault by different geologists (see Pankiwskyj, 1976; Newberg, 1985). As interpreted in this paper, it cuts and offsets the Waterville Formation, and according to Newberg (1985) it displaces metamorphic isograds. The Dearborn Brook fault is thought to be steeply dipping, and its displacement is not known. Newberg (1985) thought its slip direction was horizontal, but the inferred dips of the Waterville Formation and the isograds suggest that it had a large component of dip-slip. It postdates F_2, and because it cuts metamorphic isograds, it probably postdates F_3 as well.

A second high-angle fault, the Palermo School fault, has been mapped along the west side of the Nehumkeag Pond unit of the Cushing Formation (Newberg, 1985). In this paper the Palermo School fault is interpreted to lie within the outcrop of the Nehumkeag Pond unit, although it truncates one of the inliers containing possible Hutchins Corner Formation. Displacement on this fault is unknown. It is thought to have similar displacement to and to have the same age as the Dearborn Brook fault.

The Messalonskee Lake thrust is placed along the boundary between the Waterville and Sangerville Formations. This fault juxtaposes different stratigraphic units along its trace, and to the north of the area of study, it cuts out the entire outcrop of the Sangerville Formation, bringing into contact the Waterville and Perry Mountain Formations. F_2 folds that deform this thrust (Fig. 2) plunge toward the northeast, indicating that the thrust and the Waterville Formation structurally overlie the Sangerville Formation. This geometry, coupled with the relative

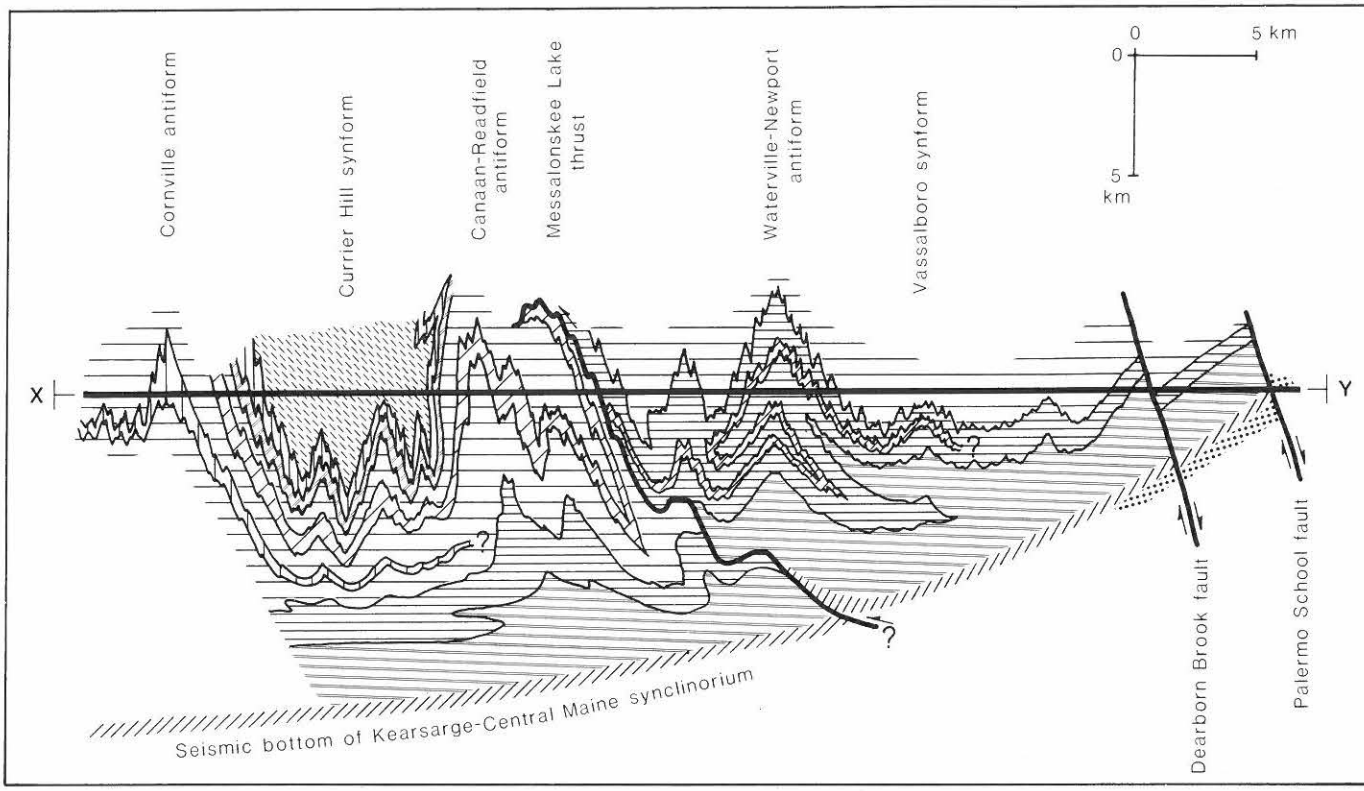

Figure 13. Structure section for south-central Maine. Line of section is shown in Figure 2. Patterns are the same as in Figure 2.

ages of the Waterville and Sangerville Formations, suggests that the thrust dips toward the east. This dip is corroborated by the U.S. Geological Survey's Maine seismic reflection profile which shows this thrust as a secondary reflector (D. B. Stewart, pers. commun., 1985). These relationships suggest that transport on it was toward the west. It is folded by F_2 and postdates F_1, but the thrust cuts the axial surfaces of F_1 folds at a sufficiently low angle as to suggest that they may belong to the same deformational event.

Structure Section

Figure 13 shows a structure section produced by projecting the surface geology northeastward parallel to the axes of F_2 onto a plane oriented perpendicular to the axes. The section is located by X-Y on Figure 2.

F_2 folds dominate the structure section. They are upright and steep sided, but these large F_2 folds must die out at deeper levels, and at a depth of from 12-15 km the structures must be essentially flat.

These subsurface relations are constrained by a prominent reflector that delineates the seismic bottom of the Kearsarge-central Maine synclinorium on a deep reflection seismic profile (Stewart et al., 1986; Unger et al., 1987). This reflector is west-dipping and comes to the surface near the contact between the Hutchins Corner Formation and the Nehumkeag Pond unit. It is here interpreted to be a normal sedimentary contact separating rocks of high velocity contrast. However, other interpretations are possible; it has been interpreted as a detachment between a cover sequence and stiffer basement (Stewart et al., 1986), or it might represent a thin transition zone where vertical structures above are collapsed into more-or-less flat structures below.

F_1 folds are shown in the central part of the section. These folds must have been extremely tight to isoclinal in order to have their observed interaction with F_2. In the structure section (Fig. 13), F_1 folds are shown to verge toward the west. In this interpretation the transport directions of both the F_1 folds and the Messalonskee Lake thrust are toward the west and give rise to the possibility that the two features may be related.

The high-angle faults in the eastern part of the sections postdate the major ductile deformation in the region.

METAMORPHISM

The rocks of southeastern Maine have been recrystallized in widespread metamorphic events. Isograds are shown in Figure 2. Detailed studies of the metamorphic rocks have been made by Osberg (1971), Ferry (1976a, 1976b, 1978), and Novak and Holdaway (1981). Well-mapped contact metamorphic aureoles exist around the Hartland pluton (Ludman, 1977) and the Old Point pluton (Pankiwskyj, 1979), but the contact aureole around the Norridgewock pluton merges to the southwest with a

widespread regional metamorphism. The Hallowell, Togus, and Three Mile Pond plutons are embedded in the field of the regional metamorphism, and the isograds bear only an approximate spatial relationship to those plutons.

All rocks in the northern part of the area of study have been recrystallized to chlorite-bearing assemblages. The first appearance of biotite is shown by the low-grade isograd in Figure 2. The middle-grade zone is based on the first appearance of andalusite, cordierite, or staurolite in pelitic rocks and on the first appearance of amphibole plus calcic plagioclase (An_{60}-An_{95}) in calcareous wackes. The high-grade zone is based on the first appearance of sillimanite in pelitic rocks and on the first appearance of diopside in calcareous wackes. The zones are locally telescoped around plutons and, consequently, not all zones can be shown.

Textural relationships in the metamorphic rocks suggest that at least three metamorphic events (M_1, M_2, and M_3) have affected the rocks. The dominant schistosity in the rocks is controlled by the parallelism of fine muscovite and chlorite. It is parallel to the axial surfaces of F_2 and, because these folds are isoclinal, the schistosity for the most part is also parallel to bedding. The early metamorphic event that establishes the parallel arrangement of muscovite and chlorite is regarded as M_1. It is not known whether any of the chlorite-grade rocks in south-central Maine are affected by only M_1. Certainly a large part of the chlorite-grade rocks and all rocks of higher grade have been recrystallized by later thermal events.

Andesitic and dacitic dikes cut across the schistosity associated with M_1, but are themselves metamorphosed to mineral assemblages consistent with the regional metamorphism. This observation, along with overprinting of the M_1 schistosity by porphyroblastic minerals with a static growth pattern, indicates the existence of a second metamorphic event (M_2). This event accounts for most of the regional metamorphic assemblages in the area. It is a Buchan-type metamorphism that produced andalusite, staurolite, and cordierite in middle grades in the pelitic rocks of appropriate compositions.

Examination of the chemistry of the associated minerals produced in M_2 indicates that those mineral assemblages were frozen in near the peak of M_2 (Osberg, 1971; Ferry, 1976a). Estimates of the conditions of this metamorphic event (Osberg, 1974; Ferry, 1976b; Novak and Holdaway, 1981) from coexisting minerals indicate that the temperature ranged from 380°C at the biotite isograd to about 550-600°C at the sillimanite isograd. Pressures have been estimated to have been in the range 3000-3800 bars. Partial pressure of fluid was certainly less than total pressure, perhaps considerably less.

Muscovite and chlorite of the M_2 event have parallel alignment in S_3 cleavage that cuts the metamorphosed andesitic and dacitic dikes, and within the middle grades of metamorphism muscovite has two orientations: relict alignment (M_1) parallel to schistosity (S_2) and a second alignment of small plates parallel to S_3. At high grades of metamorphism, mats of fibrolite locally occur in S_3. Because many of the M_2 minerals lie in S_3, the M_2 event might be coeval with or slightly younger than the development of F_3 and its related structures.

The relationship of M_2 to the Hallowell and Togus plutons is important. Barker (1961) regarded the minerals of the Hallowell pluton as primary, and Ferry (1978) thought that the metamorphism and plutonism were synchronous. This author interprets the relationship to indicate that M_2 metamorphism has affected the plutons; that the plutons were emplaced, possibly with metamorphic aureoles, and were subsequently engulfed by the thermal pulse that produced M_2. The author feels that the mortar textures in the granitoids, the metamorphic temperatures calculated within the granites by Ferry (1978), and the tracing of isograds through the granites on the basis of the mineral assemblages of zenoliths support this model. Furthermore, the relationships at Augusta (Fig. 10a), where F_3 and associated cleavage deform thin granite dikes associated with the Hallowell pluton, support the young age of M_2 metamorphism.

Chlorite in subhedral plates 0.03 mm long cuts across the M_1 schistosity in some middle grade rocks. Chlorite of identical composition partly to completely replaces cordierite (regarded as an M_2 mineral) in some of these rocks as well. Thus, the growth of chlorite seems to postdate M_1 and is an alteration product of an M_2 mineral. It could be regarded as a prograde mineral produced by a reaction involving cordierite, as a post-climax M_2 mineral produced during the waning of metamorphic conditions, or as a separate and distinct retrogressive event (M_3). Novak and Holdaway (1981) have suggested that the upper sillimanite zone in the western part of the Augusta quadrangle was produced by an M_3 event, and possibly the chlorite in the eastern part of the Augusta quadrangle was produced by the same event.

The contact metamorphic aureoles surrounding the Hartland and Old Point plutons and the northern part of the Norridgewock pluton are superimposed on the schistosity that is related to M_1. Observations relating them to M_2 or M_3 have not been made, but the age of the Hartland pluton (Dallmeyer et al., 1982) is approximately coeval with M_2 to the south.

The ages of metamorphism have been determined from relationships to stratigraphic units, plutons, structural features, overprinting relations of metamorphic minerals, and radiometrically dated rocks. The earliest metamorphic event (M_1) observed in this area of study has textures that are related to F_2 and, therefore, is roughly coeval with those folds. These folds deform Siegenian and possibly Emsian rocks in northern Maine and are intruded by plutons that have been dated at approximately 394 Ma (Dallmeyer and Van Breeman, 1981). The second metamorphic event (M_2) has textures that are related to F_3. Both F_3 and M_2 overprint the Hallowell and Togus plutons (394-387 Ma, Dallmeyer and Van Breeman, 1981), and $^{40}Ar/^{39}Ar$ spectral ages (Dallmeyer, 1979) on amphibolites recrystallized in this event from the vicinity of Augusta give apparent ages of approximately 360 Ma. The youngest metamorphic event (M_3), represented by the growth of chlorite in middle-grade rocks of M_2, is superimposed on the fabric of M_2, but it has no younger limit. It may be coeval with the intrusion of the Se-

bago pluton and its attendant metamorphism, dated as Carboniferous (Lux and Guidotti, 1985).

TECTONIC CONSIDERATIONS

Stratigraphic Relationships

The grain size within the protolith of the Sangerville and Smalls Falls Formations in south-central Maine decreases from northwest to southeast. This decrease in grain size is even more apparent when the grain size within these formations is compared with that of the equivalent Rangeley Formation in western Maine (Osberg et al., 1968). Conglomeratic lenses with coarse clasts are common in western Maine and decrease in abundance and grain size toward the southeast in the Rangeley, Sangerville, and Smalls Falls Formations. Assuming that the stratigraphic relations of the Perry Mountain Formation and the Waterville Formation discussed in the section on stratigraphy are valid, the distribution, thickness, lithic character, and ages of the Silurian formations can be interpreted as a sedimentary prism having a proximal edge that lies to the west, and a more distal section that lies in south-central Maine.

Conglomerates within the Sangerville Formation and to a lesser extent within the Smalls Falls Formation contain mafic and felsic volcanics as well as hypabyssal and plutonic clasts. The occurrence of these clasts indicates that erosion in the source area had stripped deeply into a volcanic-pluton terrane.

The thick, Silurian and Gedinnian sedimentary prism of the Kearsarge-central Maine synclinorium converges in western Maine where it undergoes a remarkably fast facies change into a thin shelf sequence of equivalent age. A similar facies change occurs in the Miramichi anticlinorium, where a thin Silurian and Gedinnian section is juxtaposed east and west with much thicker shale-wacke sections of similar age. Coastal Silurian sedimentary sections are also thin. These thick-and-thin Silurian sections suggest a horst and graben terrane, indicative of extension (see Berry and Osberg, in press).

In the context of crustal extension, the Llandoverian through Wenlockian volcanic sections of the coastal belt and the Pridolian and Gedinnian volcanics of northern Maine could be rift-generated as suggested by Gates and Moench (1981; also see Pinette and Osberg, in press; for other interpretations of the northern Maine volcanics see Hon and Roy, 1981 and Laurent and Belanger, 1984). The Pridolian through Siegenian volcanics in coastal Maine may be arc-generated as suggested for the correlative Newbury Volcanic Complex in Massachusetts by Hon and Thirlwall (1985).

Siegenian flysch spreads across both the thick and thin Gedinnian and older deposits and is interbedded with volcanic rocks in northwestern Maine. This flysch may represent deposits produced by an advancing tectonic terrane.

Johnson (1979) and Berdan (1983) indicate that the brachiopods and ostracodes of Pridolian and Gedinnian age in the coastal sections of Maine are provincially different from fossils of the same age in rocks to the west. This provinciality suggests that rifting led to a sea sufficently wide to prohibit migration of spat across it.

Fold-thrust Relationships

The volcanic and sedimentary components of the Silurian and Lower Devonian sections have been variably transported on thrusts from their original positions and multiply folded, indicating a period of crustal compression. Early recumbent folds and thrusts in south-central Maine imply a large amount of transport toward the west. Later back-folding (not preserved within the area of study) produced folds with eastward vergence. Further strain of the rocks in response to continuing northwest-southeast compression was by upright folding at high structural levels and possibly by continued recumbent folding and thrusting at deeper levels. Later a new accommodation of the rocks to new boundary conditions produced the small, north-trending, asymmetric folds.

The recumbent fold, thrust features, and early upright folds recorded in south-central Maine occurred within a geologically short span of time; sufficiently short to be within the error bars of the dating methods. These features deform rocks as young as Siegenian and possibly Emsian (401-387 Ma; Palmer, 1983) and are cut by granites dated at 402 to 394 Ma (Dallmeyer and Van Breeman, 1981; Hubacher and Lux, 1987). Only the late north-trending folds postdate the 394 ± 8 Ma Hallowell pluton, and these features are best dated at about 360 Ma.

Plutonism

The plutons of south-central Maine belong to an ill-defined group of plutons that were intruded within the interval 400-360 Ma. Possibly, these plutons belong to two overlapping groups with a high frequency of intrusion at about 395 Ma and a second at 370 Ma, although this apparent bimodal character could be due to sampling biases. A third group of plutons has a maximum frequency of intrusion at approximately 320 Ma.

Plutons belonging to the 400-360 Ma group include gabbro, diorite, granodiorite, and granite, and their chemistry suggests that some have contributions from mantle sources (Hon et al., 1981), others have igneous sources (Hon et al., 1981; Ayuso, 1984), and still others have strong contributions from sedimentary rocks. The plutons that have ages around 320 Ma are mostly granites, and nearly all of these have sedimentary sources.

The heat necessary for the production of magmas can be generated by crustal thinning, by subduction of oceanic crust, and by thickening of the crust in continent-to-continent collisions. Heat generated by these processes moves through the crust by convection (carried by magmas) and conduction with a long time constant so that magmas may not be emplaced for 20 to 50 million years after the initiation of the process leading to enhanced heat flow. As a result, plutons produced by entirely different processes may overlap in time.

The early plutons (~ 395 Ma) could have been generated by crustal thinning during a period of extension manifested by the

horst and graben features displayed by the Silurian-Gedinnian stratigraphy, by the Pridolian and Siegenian volcanics of northern Maine, and by the Llandoverian and Wenlockian bimodal volcanics of coastal Maine. This extension produced an ocean between the coastal terrane and the regions now lying to the west. The Pridolian and Siegenian volcanics of coastal Maine could have been initiated by subduction related to the closure of this ocean, leading to a continent-to-continent collision at about 400 Ma. This collision thickened the crust sufficiently to allow partial melting of crustal rocks, producing plutons that cluster around 375 Ma; further tightening of the mountain belt at the end of the Devonian produced a new round of plutons at about 320 Ma (James De Yoreo, pers. commun., 1987).

Metamorphism

The combination of conduction enhanced by the upward movement of magma (James De Yoreo, pers. commun., 1987) resulted in a large part of the crust being heated above the normal crustal distribution of heat. Sedimentary rocks in the deeper parts of the crust recrystallized in a Barrovian-type facies series and those in the higher parts in a Buchan-type facies series. Thus, a concentric relationship between the two metamorphic facies-types was developed. South-central Maine lies at the north edge of the high-level metamorphic series, and metamorphism toward the south is at a deeper level.

Summary

The Silurian-Gedinnian sedimentary rocks of New England are interpreted to have been localized by horsts and grabens in a back-arc region that lay west of an Andean-type magmatic arc. Rifting of the back-arc region produced a small ocean that separated coastal belts from more westerly parts of the terrane. In Pridolian time, westward motion of the magmatic arc began to close out the small ocean to its west, probably along an east-dipping subduction zone located along the west edge of the coastal belts. The magmatic arc ultimately collided with continental crustal terranes to its west in latest Early Devonian time producing a plethora of thrusts and folds, greatly thickening the crust. The magmatic and metamorphic features of these terranes are a consequence of these processes.

Appendix 1: Map Credits

1. Anson quadrangle – Pankiwskyj (1979); Osberg (1980), notes.
2. Skowhegan quadrangle – Ludman (1977); Osberg (1980), notes.
3. Pittsfield quadrangle – Griffin (1973); Osberg (1980), notes.
4. Norridgewock quadrangle – Pankiwskyj (1978); Osberg (1980), notes.
5. Waterville quadrangle – Osberg (1968); Osberg (1980), notes.
6. Augusta quadrangle – Barker (1961); Osberg (1968); Osberg (1980), notes.
7. Vassalboro quadrangle – Osberg (1968); Osberg (1980), notes.
8. Liberty quadrangle – Pankiwskyj (1976); Osberg (1980), notes; Newberg (1985).

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A Silurian (?) Unconformity at Flanders Bay, Maine

Richard A. Gilman
Gary G. Lash
Department of Geosciences
State University College at Fredonia (SUNY)
Fredonia, New York 14063

ABSTRACT

Reexamination of exposures of conglomerate and greenstone at the head of Flanders Bay has resulted in the recognition of an unconformity believed similar to that exposed on Spectacle Island on the east side of Frenchman Bay. It is proposed that the conglomerate is part of the Bar Harbor Formation of Siluro-Devonian age; the underlying greenstone is probably of Silurian age.

The rocks above the unconformity consist of pebbly siltstone and polymict, matrix-supported conglomerate in which clasts are predominantly of felsic volcanic rocks and quartz. A depositional model is proposed in which gravels derived from a volcanic source terrane, and transported in a high energy fluvial system, were mixed with basin-margin muds to produce debris flows.

Chemical analyses of the underlying greenstone show it to be andesitic, but to have substantial differences in composition from Silurian and Lower Devonian basalts and basaltic andesites from the Machias-Eastport area.

INTRODUCTION

In his study of the stratigraphy of the Bar Harbor Formation in the Frenchman Bay area, Maine, Metzger (1979) briefly mentions exposures of pebble conglomerates that contain clasts of volcanic rocks and quartz at several isolated outcrops at the head of Flanders Bay, in the vicinity of Jones Cove (Fig. 1). The conglomerates are associated with a weakly foliated greenstone. He concluded that the conglomerates are intraformational within a sequence of greenstones rather than belonging to the conglomerates of the Bar Harbor Formation that he describes from Spectacle, Turtle, Flat and Heron Islands on the east side of Frenchman Bay (see also Gates, in press). There, polymict basal conglomerate of the Bar Harbor Formation containing clasts of quartzite, epidote clots, greenstone, felsites, basalt, and granite, unconformably overlies a greenstone having numerous veins and "amygdules" of white quartz, the latter being flattened in the plane of a prominent spaced cleavage that strikes N70°E and dips 60 degrees to the south. Metzger correlated the greenstone of Spectacle Island with the greenstone at Flanders Bay.

Reexamination of the area at the head of Flanders Bay (by Gilman) has disclosed what is believed to be an unconformity between the greenstone and the conglomerates, and suggests an alternative interpretation to Metzger's; namely that the conglomerates are in fact part of the Bar Harbor Formation, and that the unconformity at Flanders Bay is perhaps the same as that observed on Spectacle Island.

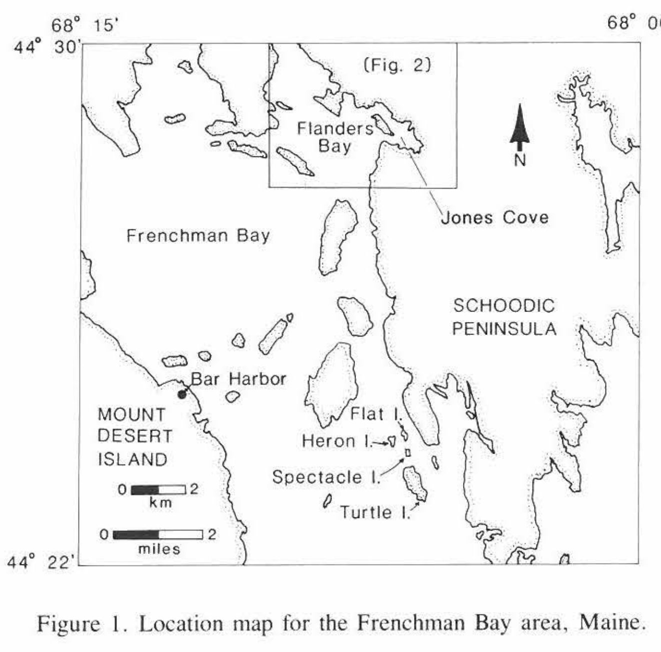

Figure 1. Location map for the Frenchman Bay area, Maine.

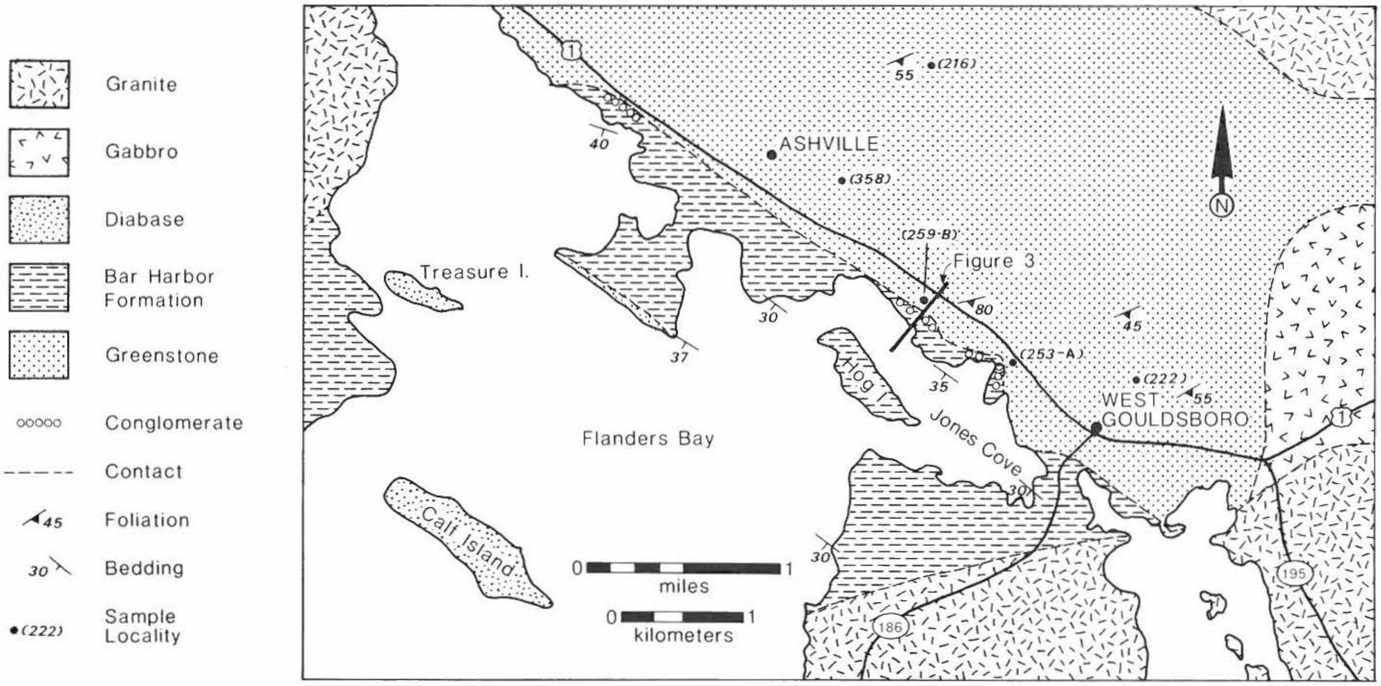

Figure 2. Geologic map of the Flanders Bay area; in part modified from Metzger, 1979. Numbers in parentheses are analyzed samples in Table 1.

DESCRIPTION OF THE UNCONFORMITY AT FLANDERS BAY

The unconformity is exposed at several locations between the shore and U.S. Route 1 west of West Gouldsboro (Fig. 2). It is usually marked by an abrupt change in slope a few hundred meters south of the highway where limited moss-covered exposures can be found in the woods (Fig. 3). It appears to be a planar but irregular surface that strikes northwestward and dips toward the southwest.

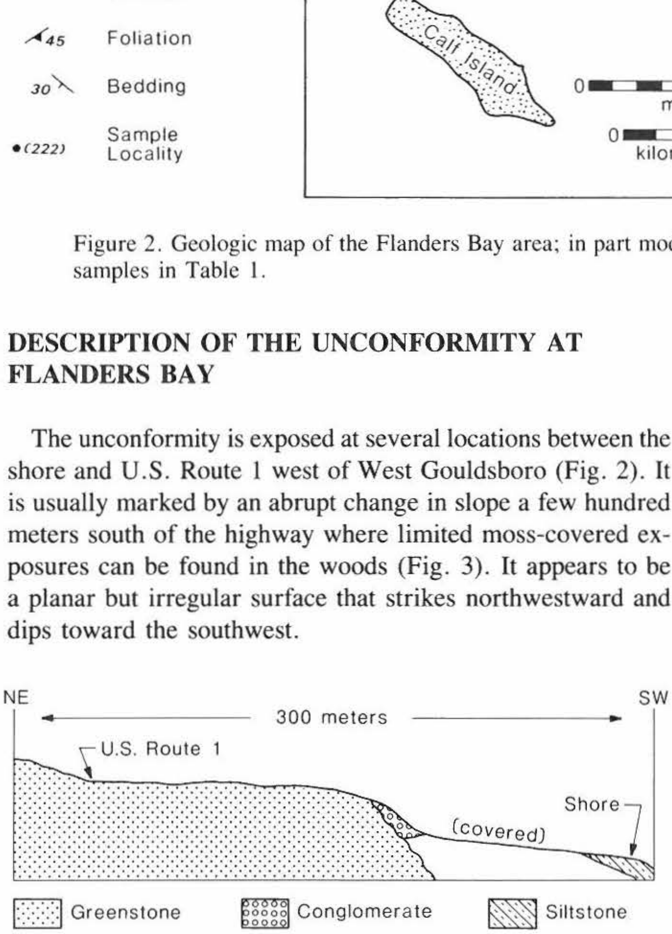

Figure 3. Cross section of the unconformity at Flanders Bay (not drawn to scale; see Fig. 2 for location).

Greenstone below the unconformity is characterized by abundant epidote veins and a faint but persistent foliation that strikes N70°E and dips 45-70 degrees to the south. Similar rocks crop out extensively north of the highway before they are truncated by granite and gabbro. In thin section the greenstone consists of a fine-grained assemblage of plagioclase and actinolite in which mats and "feathers" of actinolite usually obscure the plagioclase grain boundaries. A few sections show plagioclase phenocrysts. Primary mafic minerals have been completely replaced by actinolite and minor amounts of epidote. Some samples show a weak compositional layering, but none show a clear metamorphic foliation. The rock appears to have undergone a weak thermal metamorphism from the intrusion of granite and gabbro that lie to the north and east (Fig. 2).

Epidote veins and the foliation of the greenstone are abruptly truncated by the overlying sedimentary rocks which, in some instances, is a pebbly siltstone rather than a conglomerate. Bedding in these associated siltstones parallels the attitude of the contact, but the conglomerate itself appears to be non-bedded. The conglomerate consists of well-rounded clasts, usually less than 6 cm in diameter but locally as large as 30 cm, that typically "float" in a matrix of dark gray mudstone and siltstone. Clasts include light- and dark-colored volcanic rocks and quartz. Seven oversized thin sections were made from samples of the conglomerate. Of about fifty pebbles over 1.0 cm in diameter, the vast majority were felsites, a few of which showed porphyritic, pilotaxitic, or flow-banded textures; the remainder were polycrystalline quartz. No pebbles of plutonic igneous rocks or greenstone were noted in thin section.

Two exposures, one about a mile west of West Gouldsboro, the other just south of U.S. Route 1 west of Ashville (Fig. 2), exhibit gently dipping siltstone and quartz-pebble conglomerate, respectively, assumed to overlie the greenstone. These exposures emphasize the lithologic and structural diversity of the sedimentary rocks lying directly above the greenstone.

Exposures of the conglomerate are only a few meters thick and adjacent siltstone outcrops of the Bar Harbor Formation are commonly found within 100 meters to the south. Therefore, the conglomerate is much thinner than the conglomerates on and near Spectacle Island. This, plus the lack of greenstone clasts led Metzger (1979) to conclude that the conglomerate is not a basal unit of the Bar Harbor Formation, but rather an intraformational conglomerate in the greenstone. However, additional mapping has not disclosed greenstone exposures to the south between the conglomerate and siltstones of the Bar Harbor Formation along the shore, nor conglomerate exposures north of U.S. Route 1. Consequently, Gilman (1984) considered the conglomerate to be within the Bar Harbor Formation.

DISCUSSION

One of Metzger's (1979) arguments for not including the Flanders Bay conglomerates within the Bar Harbor Formation was the lack of clasts of the underlying greenstone such as are found on Spectacle Island and neighboring islands (see also Gates, in press). In this sense it is not the "typical" basal conglomerate in which clasts of the underlying lithologies are abundant. Moreover, the Flanders Bay conglomerates do not have massive boulder beds as described by Metzger on Turtle, Flat, and Heron Islands, nor are the Flanders Bay conglomerates as thick. But these differences need not imply that the conglomerates at Flanders Bay are not Bar Harbor Formation equivalents. If the Flanders Bay conglomerates are related to other polymict conglomerates along the coast as described by Gates (in press), his proposed origin of the conglomerates, which for some involves a debris flow process, would account for the local variability in both thickness and lithology observed at different locations, and also for the lack of locally derived clasts in a conglomerate such as at Flanders Bay.

Clast lithologies of the conglomerate clearly reflect a lithologically diverse source area. The texturally mature nature of clasts is best explained by abrasive rounding of detritus during transport in a high-gradient fluvial system prior to final deposition from debris flows. Additionally, rapid transport would enhance the preservation potential of unstable volcanic detritus. The origin of the mud matrix of the conglomerates is problematic when considered in the context of transport of the clasts. It is certainly conceivable that the well-rounded clasts mixed with offshore or slope mud during downslope gravity transport to form debris flows. According to Metzger (1979) the conglomerates on and near Spectacle Island are interlayered with sandstones and siltstones that accumulated from turbidity currents.

Sedimentologic characteristics of rocks of the Bar Harbor Formation (Metzger, 1979) suggest that the paleogeography of the source terrane was characterized by areas of moderate to high relief surrounded by flat fluvial plains and shallow basins in which sediment derived from the complex volcanic-plutonic terrane was variably reworked (Fig. 4). Periodically, downslope

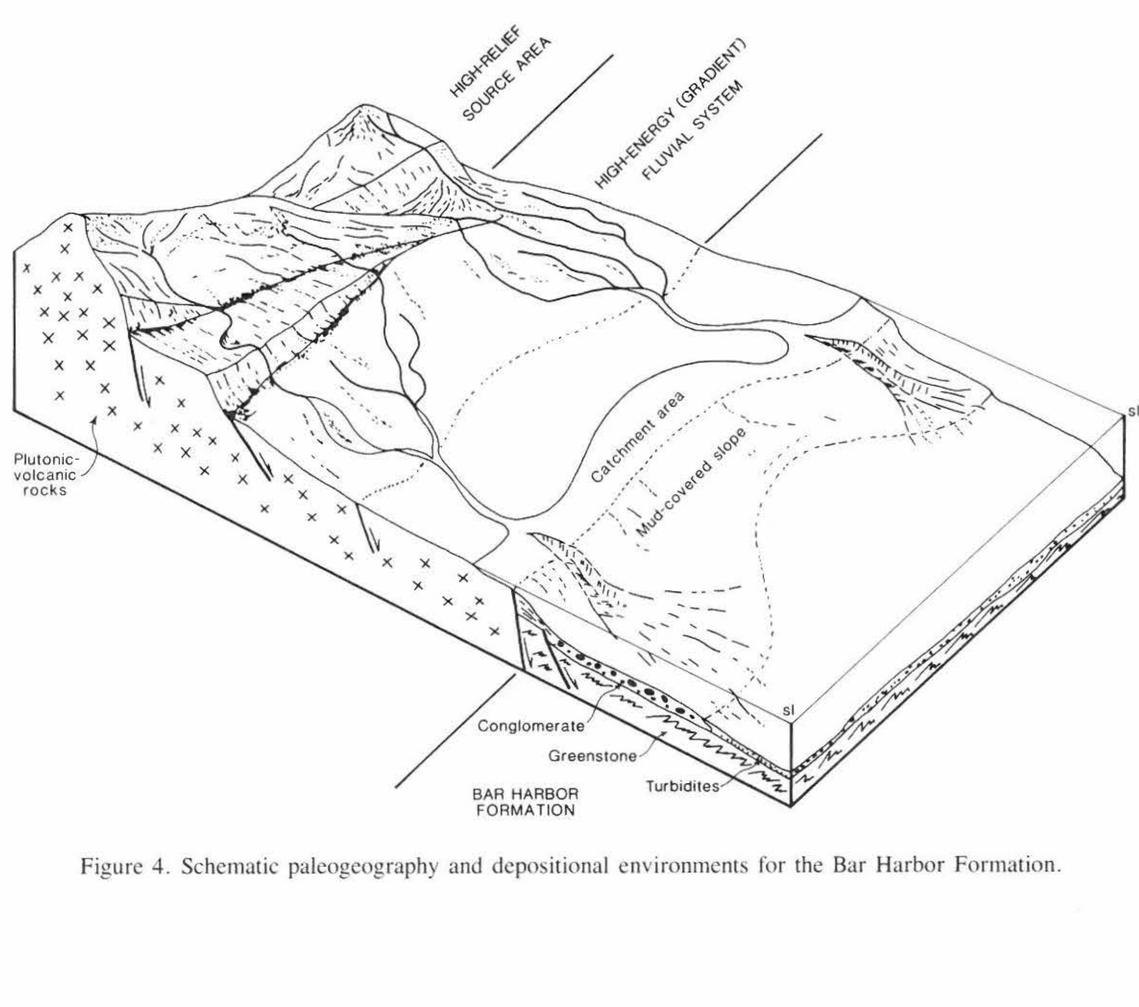

Figure 4. Schematic paleogeography and depositional environments for the Bar Harbor Formation.

gravity flows comprised of a mixture of well-rounded volcanic and/or plutonic clasts were generated. This sediment mixed with slope mud deposited farther offshore to form the debris flows (Fig. 4). Fine-grained sandstone and siltstone turbidites accumulated between periods of conglomerate deposition (Metzger, 1979). Accumulation of these deposits may reflect sediment bypassing of shallow water catchment basins in which gravel-sized detritus preferentially accumulated only to be released at a later time to form the conglomerate beds.

The interpretation that the Flanders Bay conglomerate represents Bar Harbor Formation resting unconformably on the greenstone also eliminates the need for an inferred fault postulated by Metzger (1979) with which he separates the conglomerate from the Bar Harbor Formation siltstones along the shore. The fact that there is laminated siltstone associated with the pebble conglomerate that looks much like the siltstone of the Bar Harbor Formation along the shore, and that there are no known exposures of greenstone between the conglomerate and the exposed siltstones of the Bar Harbor Formation adds support to this alternative interpretation.

GREENSTONE CORRELATION

Another question raised in the Flanders Bay area concerns the correlation of the greenstone. Chapman (1957, 1970, 1974, and pers. commun., 1985) considers the greenstone to be atypical Cranberry Island Series. This is a group of mostly rhyolites and tuffs with minor basaltic members found on the Cranberry Islands and on the southern end of Mount Desert Island on the west side of Frenchman Bay. However, Brookins et al. (1973) report a Rb-Sr age of 387 ± 9 Ma (converts to 375 Ma using $\lambda = 1.42 \times 10^{-11}$/yr) for the Cranberry Island Series and Metzger et al. (1982) obtained a Rb-Sr age of 408 ± 27 Ma for tuffs within the Bar Harbor Formation at Ireson Hill on Mount Desert Island. Thus while the Bar Harbor Formation is clearly younger than the greenstone at Spectacle Island and Flanders Bay, the correlation of the greenstone with the Cranberry Island Series as proposed by Chapman needs further study.

Metzger (1979) proposed that the greenstone is a hitherto unrecognized unit in the regional stratigraphy. The unit is older than the Bar Harbor Formation, and hence he believes older than the Cranberry Island Series, but younger than the highly tectonized Ellsworth Schist (Precambrian-Ordovician) that is exposed at the north end of Frenchman Bay and on Mount Desert Island. Given the wide range in ages between the units of the coastal volcanic belt as shown by Gates (in press), it may be that the greenstone is correlative with one of the older suites in the coastal volcanic belt. Chemical analyses of five samples of the greenstone from Flanders Bay were obtained in order to better describe the unit and to compare it to volcanic rocks from the Machias-Eastport area (Gates and Moench, 1981). Tables 1 and 2 show that the Flanders Bay greenstone is consider ably higher in SiO_2 than the Silurian basic lavas of the Machias-Eastport area, but is closer to those of the "older Devonian" Eastport Formation; both being basaltic andesites on the basis of SiO_2 content. However, other major oxides, notably CaO and Na_2O, are substantially different between these two suites.

Figure 5 compares elemental abundances and ratios of the least mobile elements (Ti, Zr, Y; Smith and Smith, 1976; Pearce and Cann, 1971) of the Flanders Bay greenstone to basic volcanics of the Machias-Eastport area. Although immobile element discriminant diagrams are most commonly used to determine the tectonic environment of eruption of orogenic basalts (e.g., Pearce, 1980, among others) they are employed here simply as a comparative rather than interpretive tool. The discriminant plots show that the Flanders Bay greenstone plots closer to the Silurian volcanics than to the "older Devonian" ones.

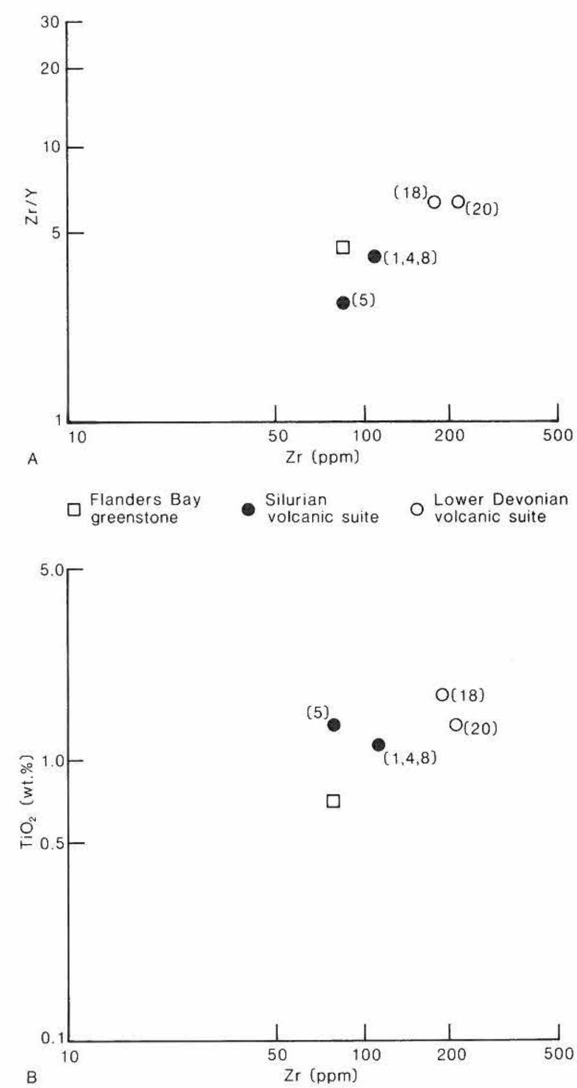

Figure 5 a,b. Abundances of Ti, Zr, and Y for greenstone from Flanders Bay compared to Silurian and Lower Devonian volcanic suites from the Machias-Eastport area (numbers in parentheses are samples from Gates and Moench, 1981). (a) Zr/Y vs. Zr (b) TiO_2 vs. Zr.

CONCLUSIONS

1. In light of new field data, the conglomerates exposed at the head of Flanders Bay are reinterpreted to be part of the Bar Harbor Formation that lies with angular unconformity above the weakly foliated greenstone, rather than as an intraformational conglomerate within the greenstone as proposed earlier by Metzger (1979).

2. The well rounded clasts, mostly of volcanic rocks and quartz supported by a fine siltstone matrix, along with the association of laminated siltstones, suggest a debris flow process of sedimentation in which coarse gravels were mixed with silt and mud prior to final deposition. Such a process has been proposed for other conglomerates along the coast (Gilman, 1966; Gates, in press).

3. Chemical analyses of the greenstone from Flanders Bay do not match closely any of the analyses for the basalts and basaltic andesites from the Machias-Eastport area.

TABLE 1. CHEMICAL ANALYSES OF FLANDERS BAY GREENSTONES RECALCULATED WITHOUT VOLATILES

Sample #	358	259-B	216	222	253-A
SiO_2 (wt.%)	58.1	53.8	55.3	54.2	54.5
Al_2O_3	14.8	15.5	14.9	15.2	16.2
Fe_2O_3 (total Fe)	9.2	11.6	10.7	9.4	10.7
MgO	4.7	4.6	5.7	6.8	5.4
CaO	9.6	10.4	9.1	9.9	9.7
Na_2O	1.6	2.6	2.5	2.5	2.0
K_2O	0.8	0.2	0.6	0.9	0.4
TiO_2	0.7	0.8	0.8	0.6	0.7
P_2O_5	0.10	0.12	0.10	0.08	0.11
MnO	0.17	0.25	0.18	0.16	0.20
Cr (ppm)	30	70	60	180	110
Sr	110	120	80	130	250
Y	20	20	20	20	20
Zr	80	100	100	60	80
Nb	20	30	20	30	20

XRF analyses by X-RAY ASSAY LTD., Don Mills, Ontario, Canada

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TABLE 2. MAJOR OXIDE ANALYSES IN WEIGHT PERCENT OF VOLCANIC ROCKS IN THE MACHIAS-EASTPORT AREA, MAINE RECALCULATED WITHOUT VOLATILES

Assemblage	Silurian						Older Devonian					
Formation	Dennys		Edmunds		Leighton		Eastport					
Sample No.	1	2	3	4	5	6	16	17	18	19	20	21
Laboratory No.	D160 372W	W176 692	W176 691	W176 689	W176 686	W178 301	W176 671	D160 375W	W176 674	W176 673	W176 672	W176 679
Rock Class	Flow	Flow	Flow	Flow	Flow	Flow	Tuff	Flow	Flow	Flow	Flow	Flow
Phenocrysts	N	N	P	C,O	N	P,C	N	P	P	P,C	P,C	P
SiO_2	51.4	49.7	52.8	51.0	48.3	53.4	56.6	49.8	52.0	53.1	54.9	56.6
Al_2O_3	17.2	17.4	16.2	17.3	17.5	16.8	16.0	19.7	17.7	15.2	16.8	15.7
FeO (FeO+0.9 Fe_2O_3)	8.8	10.7	9.4	9.7	10.6	8.1	9.2	11.1	10.0	11.6	9.5	11.9
MgO	8.0	8.5	6.5	6.9	8.3	6.2	7.0	3.1	5.3	4.3	3.8	2.9
CaO	9.5	6.7	7.7	9.5	10.9	9.8	4.7	8.9	7.4	6.3	8.0	5.2
Na_2O	2.2	3.2	4.6	3.2	2.2	3.0	3.1	3.7	3.4	4.9	3.2	4.2
K_2O	1.0	1.3	.86	.58	.16	.48	1.2	.67	1.6	.34	1.5	.43
TiO_2	1.3	1.7	1.3	1.3	1.5	1.4	1.5	1.9	1.8	2.8	1.4	2.0
P_2O_5	.26	.23	.16	.20	.16	.39	.24	.30	.29	.48	.36	.77
MnO	.17	.22	.16	.23	.16	.10	.17	.26	.21	.24	.15	.25

(modified from Gates and Moench, 1981)

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Lower Devonian Deltaic Sedimentary Environments and Ecology: Examples from the Matagamon Sandstone, Northern Maine

Stephen G. Pollock
Department of Geosciences
University of Southern Maine
Gorham, Maine 04038

Arthur J. Boucot
Department of Zoology
Oregon State University
Corvallis, Oregon 97331

Bradford A. Hall
Department of Geological Sciences
University of Maine
Orono, Maine 04469

ABSTRACT

Sedimentary environments including prodelta-marine slope, lower and upper delta front, and distributary and interdistributary delta plain environments existed side by side during Early Devonian (Siegenian) time in northern Maine. These environments are represented in the stratigraphic record by the Seboomook Group and the Matagamon Sandstone. Prodelta-marine slope and basin environments are represented by the Seboomook Group. Deltaic sedimentation in the Matagamon Sandstone is recognized by an integrated framework of 1) facies analysis, 2) vertical and horizontal arrangement of facies in stratigraphic section, 3) paleocurrent patterns, and 4) faunal communities. Utilizing these criteria, a northwesterly prograding delta front and delta plain are recognized.

The lower delta front of the Matagamon Sandstone conformably and gradationally overlies the prodelta-marine slope of the Seboomook Group. It contains three facies: 1) mudstone-siltstone facies with rare parallel laminations; 2) sharp-based, thinly bedded sandstone with vertical sequences of sedimentary structures generally ascribed to turbidites; and 3) laminated sandstone which fines upward, contains multiple sets of trough cross-lamination, and rare vertical burrows. Macerated plant debris is locally abundant in the lower delta front. Paleocurrent pattern is to the east in the laminated sandstone and northwest in the sharp-based sandstone. Associated faunas are assigned to the *Beachia* and newly introduced *Coelospira-Leptocoelia* Communities.

Conformable on the lower delta front is the upper delta front. This environment is characterized by thick beds of the laminated sandstone which contain parallel laminae and solitary and multiple sets of gently sweeping trough cross-beds. The paleocurrent pattern is complex, with an overall trend to the east. Associated faunal communities change up-section from *Coelospira-Leptocoelia* through *Tentaculites* and *Leptostrophia* to Homalonotid-*Plectonotoides*. This is interpreted as a transition from a more offshore environment that is influenced to a lesser extent by delta plain distributaries, to a nearer shore environment that is influenced to a greater extent by delta plain distributaries. Delta front deposits thicken from a minimum of 1065 m in the southeastern outcrop area to 1525 m in the northwestern outcrop area.

The delta plain environment conformably overlies and is a partial stratigraphic equivalent of the upper delta front. The rocks of this environment are approximately 520 m thick. As a whole, the environment is laterally

persistent, but the facies which comprise it are locally restricted. The delta plain environment consists of distributary channels characterized by massive, thin-bedded, ripple-laminated sandstone and shale-pebble, channel-lag conglomerates, and interdistributary bays characterized by mudstone, siltstone, and sharp-based sandstone. The latter are interpreted as crevasse-splay deposits. The paleocurrent pattern is dominantly to the northwest. Associated faunas belong to the *Cloudella* and Homalonotid-*Plectonotoides* Communities.

INTRODUCTION

Lower Devonian strata in northern Maine represent prodelta and marine slope and basin environments of the Seboomook Group, (Hall et al., 1976; Pollock, 1987), deltaic environments of the Matagamon Sandstone and presumed deltaic environments of the Tarratine Formation (Pollock, unpublished) and equivalents, including Tarratine-like localities in the Greenlaw quadrangle, north of the study area (Fig. 1). This paper interprets the depositional environments and paleoecology of the Matagamon Sandstone. Interpretation of the depositional environments and benthic paleoecology is based on detailed sedimentologic and stratigraphic analysis and comparisons with documented modern and ancient depositional environments. Interpretation of regional depositional patterns, paleoecology, and paleogeography during the Siegenian is beyond the scope of this paper.

Previous geologic investigations of the Matagamon Sandstone and similar lithologies within northern and western Maine have been primarily concerned with mapping, finding paleontologically useful fossils, and correlation. No previous efforts have been made to analyze depositional environments, nor to relate previously recognized benthic communities to depositional environments. Benthic communities have received general discussion where relative water depth or proximity to shoreline was assumed to be the primary ecologic control (Boucot, 1982; Boucot and Heath, 1969). Investigations into the Seboomook Group (Pollock, 1987) have also been concerned with mapping, finding paleontologically useful fossils, and correlation. However, portions of the Seboomook Group, which are partial time-stratigraphic equivalents of the Matagamon Sandstone, represent an extensive flysch basin which contains sediments of turbidity current (submarine fan, fan channel, or canyon) and slump and slide (slope and base-of-slope) origin, interbedded with sediments of pelagic or hemipelagic (basin plain) origin. Portions of the Seboomook Group are interpreted to have been deposited in the prodelta and marine slope environment (Hall et al., 1976; Hall and Stanley, 1972, 1973; and Hall, 1973, unpublished).

The Matagamon Sandstone has been recognized and studied for over a century. Hitchcock (1861) correctly recognized Early Devonian fossils on the east branch of the Penobscot River and Matagamon Lake, and mapped the Matagamon Sandstone as rocks of Devonian age. Clarke (1909) designated the Matagamon Sandstone as part of the Moose River Sandstone and showed it as part of a discontinuous band of Lower Devonian sandstone (Fig. 1), trending from Moosehead Lake northeast to Hay Lake and Mud Pond in the Shin Pond quadrangle. Boucot (1961) replaced the term Moose River Sandstone with the term Moose River Group. He specifically excluded all other units of Siegenian age from the Moose River Group. Rankin (1961) mapped the Traveler Mountain quadrangle, and proposed formation status for the Matagamon Sandstone in 1965. Neuman (1967) mapped the Matagamon Sandstone where it crops out in the Shin Pond quadrangle.

GEOLOGIC SETTING

Figure 1 illustrates the general distribution of Siegenian-age rocks in north-central Maine. The Matagamon Sandstone crops out on the southeastern edge of an extensive belt of interbedded slate and sandstone (Seboomook Group). The slate and sandstone belt extends from the Gaspé Peninsula of Quebec south into southern New England (Boucot, 1971). Within Maine, the Matagamon Sandstone represents the northern end of a northeast-trending belt of discontinuous sandstone bodies which underlie felsic volcanic rocks of the Piscataquis volcanic belt (Rankin, 1968). These sandstone bodies overlie and are the partial lateral equivalents of the Seboomook Group. Contact between the sandstones and the Seboomook Group is gradational (Boucot, 1961; Boucot and Heath, 1969; Rankin, 1961). The sandstones are presently isolated units that may or may not have formed a continuous stratigraphic unit. There is an almost one-to-one correspondence between the sandstone units and the overlying felsic volcanics within this belt.

The major structural element that has been superimposed on the belt of Siegenian age rocks of this study is the Traveler Mountain synclinorium (Rodgers, 1970). This synclinorium is on strike with the Moose River synclinorium to the southwest, and lies between the Lunksoos anticlinorium to the southeast and the Munsungun anticlinorium to the northwest. The Matagamon has been deformed into one large, open syncline and smaller complementary anticlines and synclines. Maximum dip of the fold limbs is 45 degrees. Fold axes plunge less than 10 degrees to the northeast.

FACIES ANALYSIS

General

Six facies are identified in the field by differences in grain size, sedimentary structures, unit geometry, bioturbation and fossils. Four facies are lithic wackes with average grain size ranging from very fine to medium sand, and matrix contents ranging from 18% to 29%. The two remaining facies, the mud-

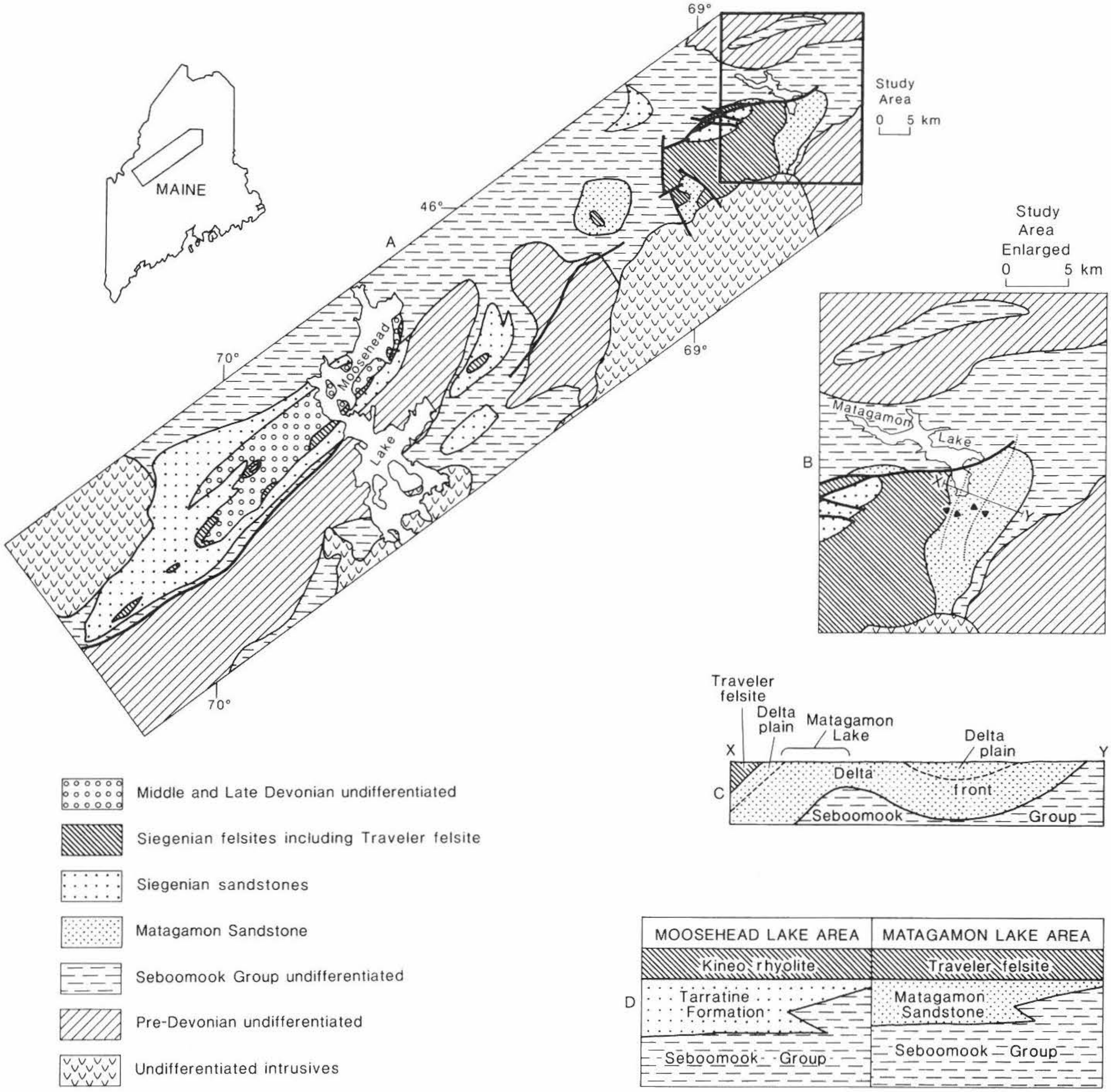

Figure 1. (a) Sketch map showing the location and outcrop distribution of the Matagamon Sandstone. It also illustrates the distribution of interpreted Lower Devonian delta and prodelta-slope-basin environments. The delta front and delta plain environments consist of the Matagamon Sandstone and Moose River Group, while the prodelta-slope-basin environments consist of the Seboomook Group. (b) Detail of study area. (c) Cross section through Matagamon Sandstone showing distribution of environments. (d) Correlation diagram.

stone and siltstone, and shale-pebble conglomerate, are texturally distinct. The four sandstone facies are differentiated primarily on the basis of bed and cross-bed characteristics. Bioturbation and fossils are generally uncommon, but their occurrence is important to environmental interpretations. Table 1 summarizes the important aspects of the six facies.

Facies 1. Mudstone and Siltstone

This is a volumetrically minor facies totalling an estimated 10% of the Matagamon Sandstone. Grain size ranges from clay to coarse silt. Bed thickness ranges from 4 cm to 3 m. Sedimentary structures are uncommon, however, texturally distinct

TABLE 1. SEDIMENTARY STRUCTURES

Number	Facies	Color	Grain size	Grain shape	Bedding thickness	Common structures	Dispersal pattern	Interpreted mechanism of deposition
1	Mudstone and siltstone	black (N1) or brownish black (5 YR 2/1)	clay to fine silt		4 cm to 3 m	flow rolls and flaser bedding in silts		suspension
2	Sharp-based sandstone	medium bluish gray (5B 5/1)	coarse silt to very fine sand	angular to subangular	1 cm to 60 cm	massive or sequences of parallel, cross, and ripple lamination. Rare sole mark and flute casts	unidirectional to northeast	turbidity currents
3	Laminated sandstone							
3-1	Thick-bedded, laminated sandstone	medium bluish gray (5B 5/1) to greenish gray (5G 6/1)	fine to medium sand	angular to subrounded	2 m to 15 m	parallel and cross lamination	multi-directional	storm wave surge, long shore or tidal currents
3-2	Thin-bedded, laminated sandstone	medium bluish gray (5B 5/1)	fine to medium sand	angular to subangular	15 cm to 1.5 m	parallel and cross lamination, ripple laminations uncommon	multi-directional with south-easterly average	storm or wave surge
3-3	Laminated-shelly sandstone	medium bluish gray (5B 5/1)	coarse silt to fine sand	subangular to sub-rounded	1 m to 2 m	massive or parallel lamination, cross-beds uncommon		?
3-4	Ripple-laminated sandstone	medium bluish gray (5B 5/1)	coarse silt to fine sand	angular to subangular	less than 1 m	asymmetric ripples and ripple drift		channelized flow in lower flow regime
4	Massive-bedded sandstone	medium bluish gray (5B 5/1)	fine to medium sand	angular to subrounded	1 m to 15 m	none, lamination rare		channelized flow, upper flow regime
5	Thin-bedded sandstone	medium bluish gray	medium sand	subangular to rounded	4 cm to 16 cm	horizontal or trough cross-beds	unidirectional to northwest	channelized flow, lower flow regime or lower upper flow regime
6	Shale-pebble conglomerate	variable browns and gray blacks	clasts are pebble size; matrix from coarse sand to silt	pebbles rounded; matrix sub-angular to subrounded	30 cm to 45 cm	none		upper flow regime

parallel and ripple cross-lamination, flaser bedding, silt lenses, and small-scale ball and pillow structures are present. These structures are composed of quartz silt in most outcrops and tuffaceous lithic grains near the top of the Matagamon Sandstone. Small macerated plant fragments are the only fossil material recovered from this facies.

Facies 2. Sharp-Based Sandstone

This facies is generally finer grained than the other sandstone facies and consists of poorly sorted coarse-grained siltstone to very fine-grained sandstone. Bed thicknesses range from 1.0 to 60 cm, with an average thickness less than 30 cm. Beds are

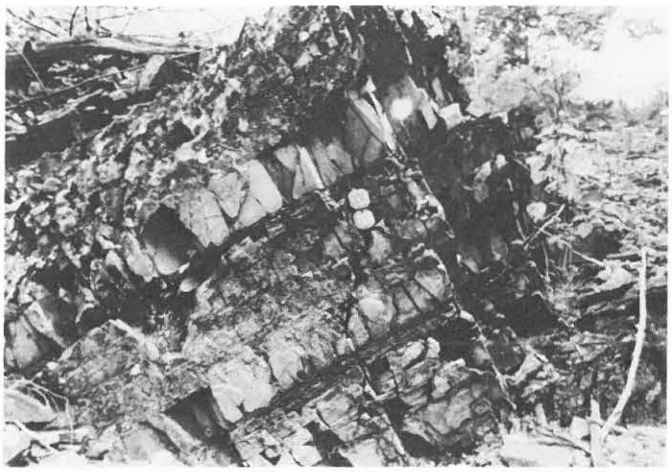

Figure 2. Sharp-based sandstone interbedded with mudstone-siltstone facies. The lighter colored sandstone beds are parallel laminated throughout with no textural gradation. The mudstones are weakly parallel laminated throughout.

Figure 3. Thick-bedded, laminated sandstone facies. Bedding and lamination style illustrated in this photograph is typical of the majority of the facies. This facies is volumetrically the most important part of the Matagamon Sandstone.

characterized by sharp bases (Fig. 2).

Textural grading from very fine-grained sandstone to siltstone or mudstone is locally common. Bed tops are undulatory or irregular, and locally the beds pinch and swell. The upper portions of these beds are commonly burrowed. The burrows are oriented sub-parallel to bedding and have diameters less than 1.0 cm.

Primary sedimentary structures include parallel and ripple lamination, slump structures, and flute and groove casts. Locally, beds are massive, whereas others consist wholly of parallel or ripple lamination. Sedimentary structures commonly exhibit a sequential arrangement, from massive through parallel into ripple laminated zones, which corresponds to the three lowest intervals of the ideal turbidite sequence (Bouma, 1962). Flute and groove casts are small, usually less than 3 cm, and occur on the bases of the thicker beds. Ball and pillow structures, rarely exceeding 1 m are locally common. The beds with sequentially arranged sedimentary structures are most common in this facies near the base of the Matagamon Sandstone in the transition zone with the Seboomook Group. Faunas have not been collected from this facies at this stratigraphic level.

At higher stratigraphic levels this facies is generally graded, massive or parallel laminated, and intensely burrowed. Burrows, approximately 1 cm in diameter, are oriented at all angles to bedding and partially obliterate bedding and sedimentary structures locally. Faunas which occur at higher stratigraphic levels are most commonly gastropods, and rarely brachiopods, and are assigned to the Homalonotid-*Plectonotoides* and *Cloudella* Communities.

Facies 3. Laminated Sandstone Facies

Four lithologically similar subfacies comprise the laminated sandstone facies. These are characterized by distinct differences in bedding, sedimentary structures, and faunal content. The rocks are poorly to moderately well sorted, angular to subangular, very fine to fine grained sandstone.

3-1. Thick-Bedded, Laminated Sandstone. This subfacies (Fig. 3) makes up most of the type section as described by Rankin (1965). It is laterally and vertically the most persistent and volumetrically the most important part of the Matagamon Sandstone. Beds range from approximately 2 to 15 m in thickness. Upper bedding surfaces are delineated by one or more of the following: 1) fossil debris, primarily tentaculitids; 2) discoidally shaped mud chips; and 3) most commonly, a slight textural grading in the upper 30 cm of the beds. Parallel and cross-lamination are the dominant sedimentary structures. Cross-lamination sets, both solitary and grouped, appear as shallow, gently sweeping troughs. The lower boundary of the sets is normally erosional, and the bases commonly truncate lower laminae. Set thickness is less than 75 cm, and more commonly less than 30 cm. It is stressed that the sets and co-sets are internally laminated. The maximum observed trough width is approximately 45 m. However, these normally are not traceable over distances of more than 4 or 5 m. Where a trough geometry is not apparent, the bases of the cross-laminae are tangential. Sets of this type are generally solitary, but rarely may be grouped. Low relief circular domes typical of hummocky cross-stratification (Harms et al., 1975) have not been recognized. The cross-stratification of this facies more closely resembles the superimposed, concave-upward shallow scours typical of swaley cross-stratification (Leckie and Walker, 1982).

Fossils occur in lenses at the base of cross-lamination sets. Representatives from five of the six benthic faunal communities occur in this facies; only the *Beachia* Community has not been recognized.

3-2. Thin-Bedded, Laminated Sandstone. Beds range in thickness from 0.15 to 1.5 m, averaging 0.45 m. Argillaceous laminae less than 2 mm thick are common. The beds have distinct tops and bottoms and commonly are coarser grained in

the lower portion. Sedimentary structures are gently sweeping trough cross-lamination, parallel lamination, ripple lamination, and small-scale slump features. These occur in the coarser basal portion of the beds and gradually die out in the upper finer-grained part of the bed. Cross-laminated beds are either solitary or grouped. Groups may contain four or five sets. Maximum set thickness is 40 cm. Width of the troughs may be as great as 8 m. The bases of the troughs are erosional and rest on other troughs or parallel laminae. Hummocky cross-stratification (Harms et al., 1975) has not been recognized. Cross-stratification within this subfacies is more typically swaley, resembling the cross-stratification in the thick-bedded, laminated subfacies. Burrows oriented at high angles to bedding, and up to 35 cm in length, are common. Burrow-filling material is finer grained than that of the surrounding bed. Fossils are uncommon but where found are primarily brachiopods assigned to the *Beachia* Community.

This subfacies differs as follows from the sharp-based sandstone: 1) coarser grain size, 2) larger scale cross-lamination, 3) lack of sub-horizontal burrows in the upper finer-grained portions of the beds, 4) lack of sole marks, 5) lack of sequentially arranged sedimentary structures, 6) lack of pinch and swell as in beds of the sharp-based facies, 7) paleocurrent pattern, and 8) faunal content.

3-3. Laminated-Shelly Sandstone. The laminated-shelly sandstone consists of one laterally extensive unit or two separate units at approximately equivalent stratigraphic horizons. Grain size of this subfacies is generally coarser than the other three, ranging from fine to medium sand size. Exposed bed thickness is 2 m, and thickness of the subfacies is estimated to be 4 to 5 m. Parallel lamination, medium-scale planar cross bedding, and massive or thin parallel bedding is locally present, but uncommon. This subfacies contains abundant small lenses of recrystallized fossil material. Fossils consist mostly of articulated and disarticulated brachiopods. Disarticulated valves are oriented concave down. Faunas from this subfacies are assigned to the *Coelospira-Leptocoelia* Community.

3-4. Ripple-Laminated Sandstone. This subfacies is comprised of beds up to 1 m thick which are ripple-laminated throughout. Two basic ripple types are present: 1) out-of-phase asymmetrical ripples and 2) types 1 and 2 ripple drift of Walker (1963) or type B of Jopling and Walker (1968). Ripple drift is the least common of the two and grades laterally into asymmetric ripples. Erosion of ripple crests is a relatively common feature. Flaser structures, in the form of mud drapes, occur in ripple troughs and on ripple foresets. Fossils have not been recovered from this facies.

Facies 4. Massive-Bedded Sandstone

Bed thickness of this facies ranges from approximately 1 to 15 m. Discoidal mud flakes locally occur near the top and bottom of beds and along internal erosion surfaces. Most beds are internally structureless (Fig. 4), although parallel and small-scale ripple and cross-laminations are rarely present. Where present, these laminae occur in the upper few centimeters of the bed. Medium-scale planar cross-beds (to 30 cm) are uncommon at the base of some beds. This facies exhibits a channel morphology. Maximum observed width of channel fill is approximately 0.5 to slightly less than 1.0 km. Maximum thickness is 15 m.

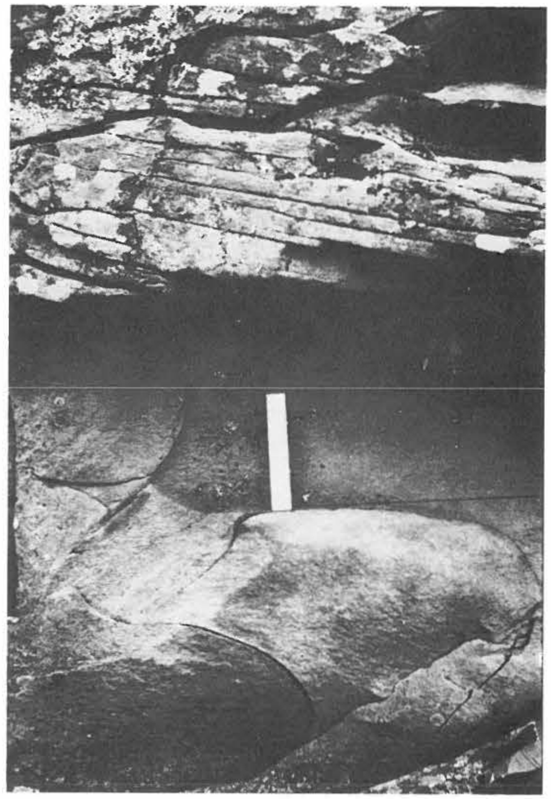

Figure 4. Massive bedded sandstone overlain by thin tabular cross beds of the thin-bedded sandstone facies.

Fossils are rare in the massive-bedded sandstone, but disarticulated brachiopods are found on erosion surfaces at the bases of some channels. Carbonaceous remains of fragmented plant stems may be dispersed along laminae within the upper few centimeters of the beds. Faunas collected from this facies are assigned to the *Cloudella* Community.

Facies 5. Thin-Bedded Sandstone

Bedding of this facies is characteristically thin and platy (Fig. 5). Bed thickness ranges from 4 to 16 cm with an average thickness of approximately 10 cm. Parting parallel to bedding is moderately well developed locally. Preconsolidation deformation (slump) structures are present locally. Small-scale cross or ripple lamination may be common in horizontal beds greater than 15 cm thick. Common structures include medium-scale trough cross-beds and medium-scale planar cross-beds. Trough sets may be solitary or grouped and range in thickness to 2 m. Erosional bases of the troughs rest on siltstone, massive-bedded sandstone, or other troughs. Trough widths reach 20 to 25 m.

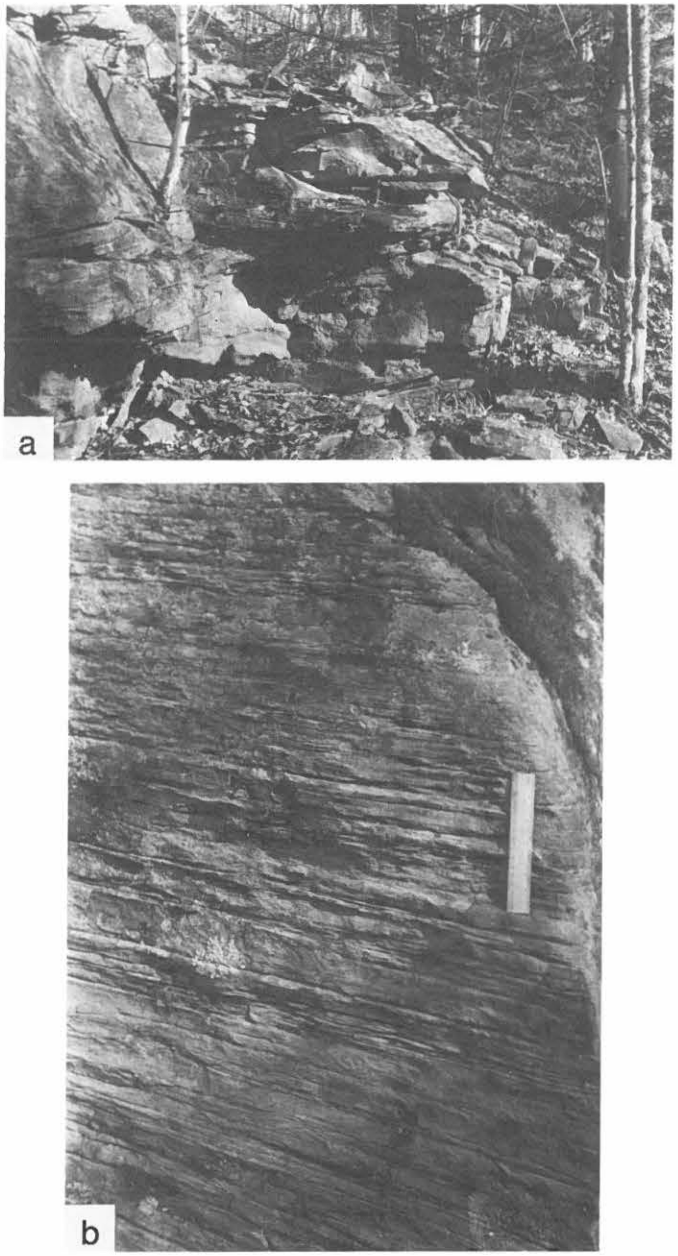

Figure 5. (a) Small channel-shaped, thin cross beds. Troughs truncate earlier deposited sets. Width of the troughs is approximately 3 m in this photograph. (b) Horizontal, thin beds of the thin-bedded sandstone.

Planar sets are commonly solitary with a maximum set thickness of 40 cm. Bases are sharp or slightly tangential to the underlying strata. Both types of tabular cross-bedding can be traced over distances of several meters. Faunas collected from this facies are assigned to the *Cloudella* Community.

Facies 6. Shale-Pebble Conglomerate

The shale-pebble conglomerate is of little volumetric importance, occurring only at the base of the massive-bedded sandstone, or at the base of trough cross-beds of the thin-bedded sandstone. The shale pebbles are predominantly of the mudstone and siltstone facies. They are ellipsoidal in shape, and are less than 3 cm in maximum diameter. The pebbles are suspended in a matrix of clay and coarse silt, fine to medium grained sand, or shell material. The sand matrix is the most common. Beds are irregular in thickness and are laterally discontinuous. Maximum thickness is approximately 50 cm. Faunas collected from this facies are assigned to the *Cloudella* Community.

BENTHIC FAUNAL COMMUNITIES

General

Fossils are uncommon to rare in the Matagamon Sandstone and Seboomook Group, and where found, are mechanically concentrated death assemblages dominated by brachiopods. Analyses of 21 bulk assemblages (Table 2) by cluster method was performed following the method described by Sneath and Sokal (1969). Results of the analysis are displayed in the dendrogram (Fig. 6). This figure illustrates the recurrence of faunal assemblages within facies, together with the interpreted depositional environments. The recurrence of faunal assemblages within similar sedimentary environments has been recognized by Walker and Laporte (1970), Anderson (1974), and Thayer (1974). Recurring faunal assemblages historically have been considered communities.

The community concept has been applied to Paleozoic brachiopods by a variety of workers; see for example Ziegler (1965), Boucot and Johnson (1967), Boucot et al. (1969), Boucot (1975, 1982), and Lesperance and Sheehan (1975). Faunal assemblages from the Matagamon Sandstone and Seboomook Group are assigned to three previously recognized Paleozoic brachiopod communities and one newly defined brachiopod community. Two non-brachiopod communities are also recognized. The brachiopod communities are: 1) the *Beachia* Community based on the presence of *B. thunii* (Boucot and Johnson, 1967; Boucot, 1982); 2) the *Coelospira-Leptocoelia* Community based on the presence of *Coelospira* sp. and/or *L. flabellites* (Boucot, 1975); 3) the *Leptostrophia* Community based on the presence of *L. cf. magnifica* or unidentified species of that genus; and 4) the *Cloudella* Community based on the presence of *C. matagamoni*. The fifth community is based on the presence of small tentaculitids of the genus *Tentaculites*. The sixth community is based on the presence of the gastropod *Plectonotoides* and homalonotid trilobites.

The collections often contain a mix of the dominant elements of more than one community (Table 2 and Fig. 6). Assignment to a particular community is based on the dominant or relative abundance of the identifying genera for which the community is named, or the matching coefficient to which the collections correlate.

TABLE 2. FAUNAS COLLECTED FROM MATAGAMON SANDSTONE WITH NUMBER OF INDIVIDUALS RECOVERED. X'S INDICATE PRESENCE, NO NUMBER DETERMINED.

United States National Museum Number	17372	17374	17378	17381	17364	17370	17376	17380	17367	17368	17379	17375	17373	17362	12372	12371	17369	17377	17371	17365	17366
FAUNAS																					
Cloudella matagamoni	18	3	9	9	1	2	54	23	1												
Plectonotus sp.					3	5	1												1		
Tropidodiscus sp.					3	1															
Cyclonema sp.						4	2														
Acrospirifer sp.							3				1	7	45				1				
Tropidocyclus sp.							1														
Leptostrophia cf. *L. magnifica*									5	21	2	46	71	1	x	x	15	7	7		
Leptocoelia flabellites											8	52	20	10	x	x	1	7	7	3	7
Platyorthis sp.											1	8	2				2				
Chonostrophiella sp.											2	1	6								
Meristella sp.											1	5	12	1	x						
Coelospira sp.												4	13	1							
Costellirostra sp.												20	4								
Plethorhyncha sp.												2									
"Schuchertella" sp.												8	2						1		
Costispirifer sp.													2								
Beachia thunii													2				29		4		
Globithyris(?) sp.													1								
Cyrtina sp.													3								
Platyceras sp.													1			x					
Loxonema sp.																	1				
Salopina sp.																		1			
"Brachyspirifer" sp.																		19			
Pleurodictyum sp.																				1	
Homalonotids					x	x															

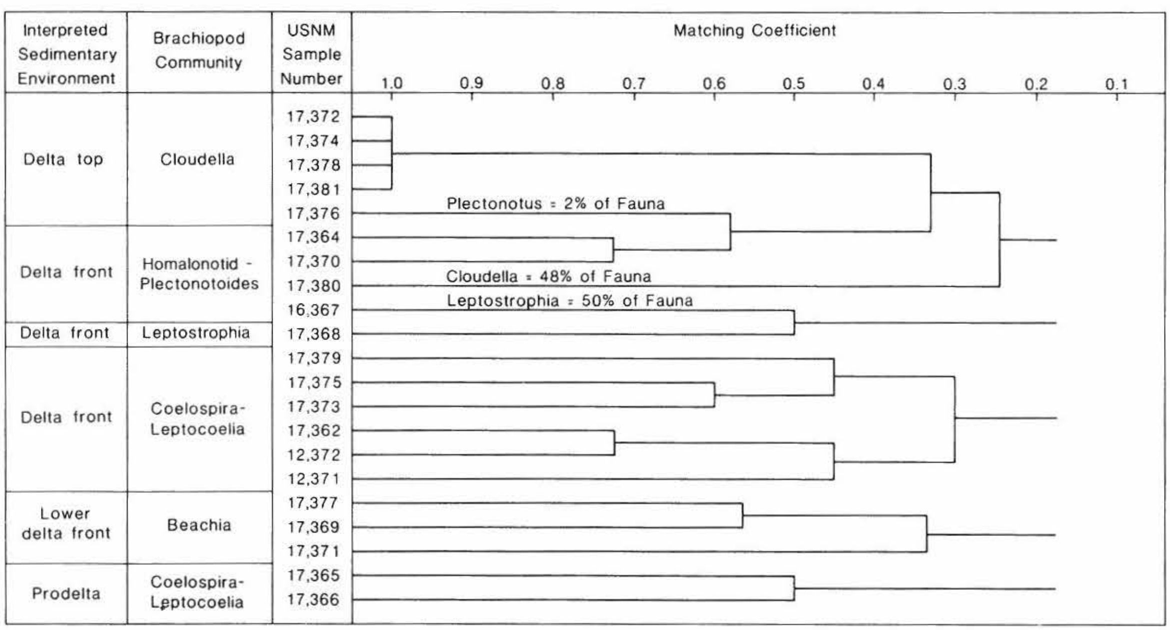

Figure 6. Dendrogram from cluster analysis of 21 collections of fossils from the Matagamon Sandstone and Seboomook Group. The dendrogram is correlated with the assigned communities and sedimentary environments from which they were collected. Collection numbers 17365, 17366, and 17371 were collected from the turbidite channels of the prodelta-marine slope environment of the Seboomook Group. Collection numbers 17377 and 17369 are from the base of the Matagamon Sandstone.

Beachia *Community*

The *Beachia* Community is used here in the same sense as in Boucot (1975, 1982). The *Beachia* Community is one of the members of the *Janius* Community Group. The *Janius* Community Group is well represented in the endemic Early Devonian Appohimichi Subprovince. Included with *Beachia* in this subprovince are the *Amphigenia, Cyrtina,* and *Etymothyris* Communities as discussed by Boucot (1982). These four communities are here designated the *Amphigenia* Community Subgroup. These Oriskany age faunas are concluded to be of about Siegenian age.

The *Beachia* Community is identified by the presence of *B. thunii*, a moderately sized centronellid. The community is a high-diversity, low-dominance community and occurs at the base of the Matagamon Sandstone in the thin-bedded, laminated sandstone subfacies. Its stratigraphic position is within the Seboomook Group-Matagamon Sandstone transition interval. It also occurs within the Seboomook Group-Tarratine Formation transition to the southwest in Somerset County (Boucot and Heath, 1969). The occurrence here is a heterogeneous assemblage, with its composition roughly conforming to that described by Boucot (1975, 1982) and Lesperance and Sheehan (1975). Other brachiopods reported to occur with *Beachia* are *Leptocoelia, Leptostrophia,* and *Acrospirifer*. In the Matagamon Sandstone these also occur with other brachiopods and members of other phyla. *Dawsonelloides* and *Meristella*, both abundant members of the *Beachia* Community as discussed by Boucot and Johnson (1967), Boucot (1975, 1982), and Lesperance and Sheehan (1975), have not been collected from the *Beachia* Community in the Matagamon Sandstone. We include Lesperance and Sheehan's (1975, Table VI) *Plicoplasia* Community in the *Beachia* Community, and not in Boucot's (1982) *Plicoplasia cooperi* or *P. plicata* Communities.

Leptostrophia *Community*

This community was introduced by Lesperance and Sheehan (1975) and is used here in their sense for rocks dominated by *Leptostrophia*. Lesperance and Sheehan's (1975) data suggest that this community should be assigned to the *Janius* Community Group. The *Leptostrophia* Community is identified by *L. cf. magnifica*, a large (to 6 cm), free-living strophomenid found in the laminated-sandstone facies. In the Matagamon Sandstone, the *Leptostrophia* Community stratigraphically overlies the *Beachia* Community, and underlies, or is interbedded with the Homalonotid-*Plectonotoides* Community. *Leptostrophia* dominates the community. It is also an important member of the *Beachia* and *Coelospira-Leptocoelia* Communities. It appears rarely in the Homalonotid-*Plectonotoides* Community. Lesperance and Sheehan (1975) indicate similar stratigraphic relations relative to the *Beachia* Community. Also, they note that the community is characterized by its overwhelming dominance of a single genus and its relative importance as a secondary member of other communities (their *Plicoplasia* Community in the Gaspé limestones).

Coelospira-Leptocoelia *Community*

The *Coeloespira-Leptocoelia* Community is used here for the first time. This is not the same as Boucot's (1975) *Coelospira-Leptocoelia* Community, which was altered by Boucot (1982) to the *Coelospira-Pacificocoelia* Community. The Cap Bon Ami Formation occurrences do *not* contain true *Leptocoelia*, whereas those in the Matagamon Sandstone do. This community is assigned to the *Janius* Community Group. True *Leptocoelia* is now known only from beds of Oriskany and Esopus age in the Appohimichi Subprovince.

In the Matagamon Sandstone, this community is identified as a high-diversity community primarily dominated by the plicate spiriferids *Coelospira* sp. and *Leptocoelia flabellites*. Other brachiopod genera consistently occurring in this community are *Leptostrophia, Acrospirifer, Meristella, Costispirifer,* and *Chonostrophiella*. This community occurs in the laminated-sandstone facies of the Matagamon Sandstone. Other occurrences of *Coelospira* and *Leptocoelia* are from turbidite sandstones of the Seboomook Group.

Cloudella matagamoni *Community*

This community is used here as in Boucot (1982). Community group assignment is uncertain as of this time (see Boucot, 1982). This community is geographically distributed in northern Maine, and is closely related to the Late Helderberg *Cloudella stewarti* Community. The *C. stewarti* Community occurs in northern New Brunswick and is possibly ancestral to the *C. matagamoni* Community. *C. matagamoni* is of Oriskany age.

The *Cloudella matagamoni* Community is composed exclusively, or dominantly, of the small mutationellid *C. matagamoni*. This low-diversity, high-dominance community occurs at the base of channels in the thin-bedded sandstone facies or from the mudstone and siltstone facies. Minor occurrences are in the thick-bedded, laminated sandstone at stratigraphically high levels beneath the thin-bedded sandstone facies. The commonly disarticulated nature of the shells from the Matagamon Sandstone may be a post-mortem effect, because the closely related *C. stewarti* shells occur in a fine-grained calcareous shale, and are commonly articulated.

Homalonotid-Plectonotoides *Community*

This community is used here in Boucot's (1982) sense. Community groups have not been worked out for either the trilobite or the bellerophontid. The community is of Late Helderberg to Oriskany age. The allied, possibly ancestral, *Plectonotus* Community is widespread in the Malvinokaffric Realm (Boucot et al., 1986).

The Homalonotid-*Plectonotoides* Community is composed of the bellerophontid gastropod *Plectonotoides*, in association with homalonotid trilobites, chonetids, and nuculoid and other

bivalves, and non-bellerophontid gastropods, together with *Cloudella* and *Leptostrophia*. The usual facies from which this community is collected is the thick-bedded, laminated sandstone with less common occurrences in the thin-bedded sandstone. The combined dominance of *Plectonotoides* and homalonotids ranges from a low of 2% in occurrence with *C. matagamoni* in thin-bedded sandstone, to 42% in the thick-bedded, laminated sandstone. Stratigraphically, this community occurs in the upper part of the thick-bedded, laminated sandstone subfacies.

Tentaculites *Community*

The *Tentaculites* Community as defined by Boucot (1982) is a low-diversity community consisting principally of tentaculitids (Mollusca: Cricoconarids). Community groups have not yet been proposed for tentaculitids, although it is obvious that these nearshore types are very distinct from the pelagic nowakiids. This is a single taxon community in the Matagamon Sandstone except for one locality where *Tentaculites* are associated with orthoceroid nautiloids and nuculoid bivalves. This community occurs only in the thick-bedded, laminated sandstone. *Tentaculites* are generally observed in random orientation on upper bedding planes, but rarely show preferred orientation with the long axes of the shell aligned approximately parallel to current direction, as ascertained from cross-lamination in the underlying bed.

ENVIRONMENTAL ANALYSIS

General

The facies occur in two natural groupings formed by repetitive sequences of a single facies, or repetition of several facies over vertical stratigraphic distances, and by facies occurring on equivalent stratigraphic horizons. The two groupings are further defined by the manner of vertical and/or lateral occurrence of facies, paleocurrent data, and faunal communities. They are interpreted as depositional sub-environments of a westerly prograding delta complex.

Deltas have received much attention over the last several years, and many studies and models are available from which to draw interpretations. Among the more recent summaries of deltaic environments are Broussard (1975), Coleman (1976), Coleman and Prior (1980), Miall (1984), Morgan and Shaver (1970), and Fisher et al. (1969). In addition to these summaries, many topical studies have been completed. It is not the intention of this paper to present a comprehensive literature review, but rather to discuss aspects of the Matagamon delta as they relate to existing models. Environments within the interpreted Matagamon delta are characterized by differences in paleocurrent patterns, faunal content, and depositional hydrodynamics, in addition to the previously described facies.

Ecological restraints are inferred from sedimentary parameters. These are: 1) distance from distributary influence as inferred from water depth, 2) energy conditions, as inferred from sedimentary structures and paleocurrent patterns, and 3) salinity. Substrate character is not considered to be a major ecological control. The major sediment type of the Matagamon Sandstone is uniformly fine-grained sandstone. From the available data, nothing can be inferred concerning other ecological constraints such as water temperature, turbidity, nutrient or dissolved oxygen content, climate, river discharge, etc.

The two environments of the Matagamon delta are delta front and delta plain. The stratigraphically lower portion of the Matagamon Sandstone is interpreted as the delta front, which is comprised dominantly of the thick-bedded, laminated sandstone subfacies, and to a volumetrically smaller degree of the mudstone and siltstone, sharp-based sandstone, thin-bedded, laminated sandstone, and laminated-shelly sandstone. Stratigraphically overlying, and partially laterally equivalent to the delta front is an environment interpreted as the delta plain. This comprises dominantly thin-bedded and massive-bedded channel sandstones and to a volumetrically smaller degree, mudstone, siltstone, sharp-based sandstone, ripple-laminated sandstone, and shale-pebble conglomerate.

Vertical and Horizontal Arrangement of Facies

The Matagamon Sandstone is approximately 1600 m thick. Figure 7 schematically illustrates the vertical and lateral distribution of facies. The lower 165 meters is a transition zone between the Matagamon Sandstone and the Seboomook Group. Facies which crop out in this interval are the mudstone and siltstone, sharp-based sandstone, and the thin-bedded, laminated sandstone. The sharp-based sandstone is more common in the lower part of interval and the thin-bedded, laminated sandstone is more common in the upper part of the interval.

Overlying the transition zone is thick-bedded, laminated sandstone which thickens from 900 m in the southeastern portion of the map area to 1360 m in the northwestern portion. The transition zone and the overlying thick-bedded, laminated sandstone are assigned to the delta front environment.

Overlying this thick section of predominantly thick-bedded, laminated sandstone is a section of interbedded facies which has a maximum thickness of 520 to 575 m. The facies grouping is considered representative of the delta plain environment. It consists of interbedded facies which include, in decreasing order of abundance, the thin-bedded sandstone, massive-bedded sandstone, ripple-laminated sandstone, sharp-based sandstone, and shale-pebble conglomerate. The vertical arrangement of the facies in this upper section is complex. Figure 7 illustrates examples of measured portions of this upper association. Specific relationships between the facies in this sequence are as follows:

1. Thin layers of the mudstone and siltstone are overlain by trough cross-beds of the thin-bedded sandstone, or by channel-fill deposits of the massive-bedded sandstone.
2. Shale-pebble conglomerate is most commonly observed to underlie massive-bedded sandstone and less commonly to underlie trough cross-beds of the thin-bedded sandstone.
3. Massive-bedded sandstone may also erode and overlie thin-bedded sandstone or ripple-laminated sandstone.

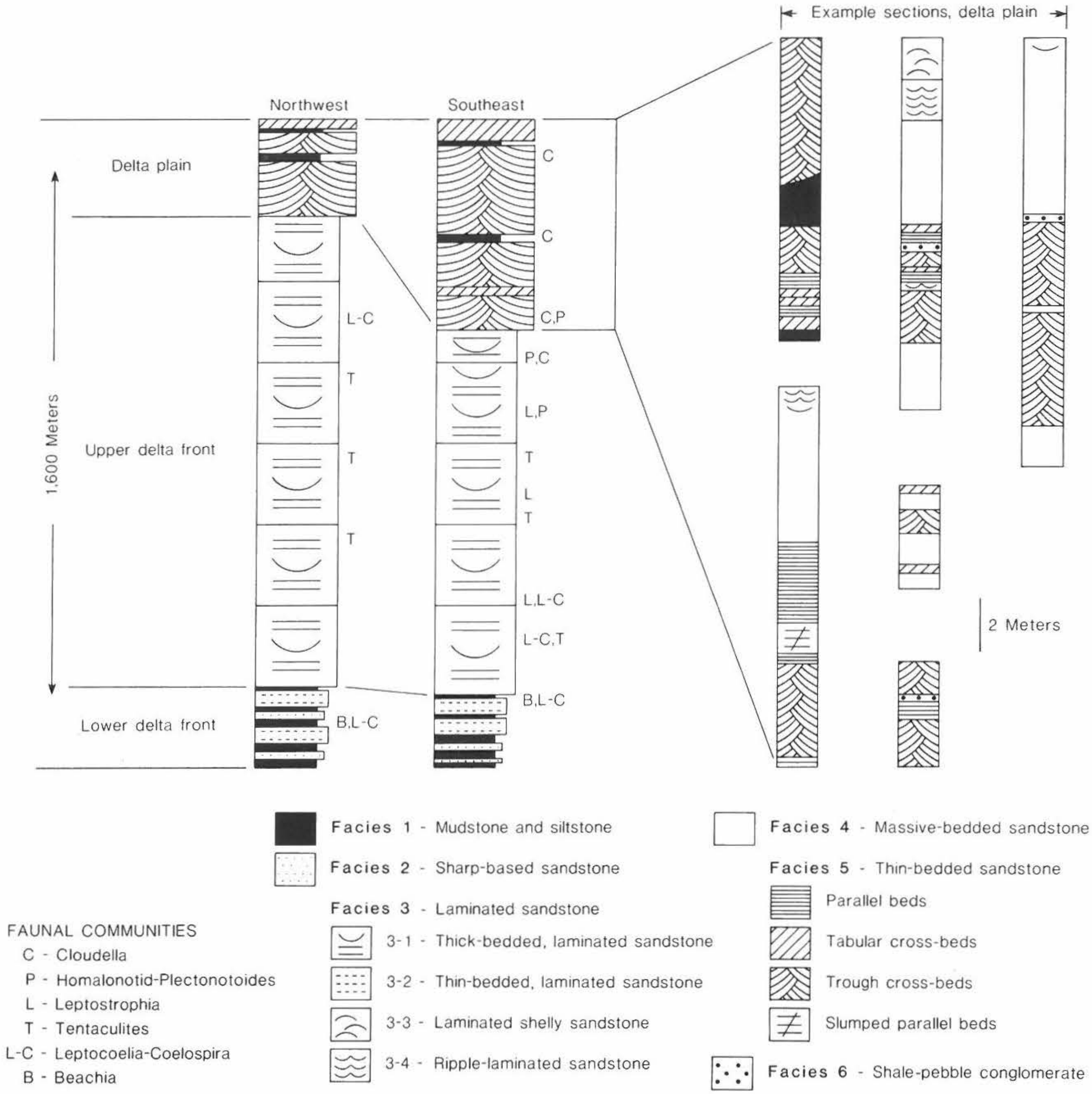

Figure 7. Schematic illustration of the vertical and horizontal arrangement of facies of the Matagamon Sandstone. The small columns to the right illustrate the localized differences in the distribution of the facies which comprise the delta plain.

4. Parallel, planar, and trough cross-beds of the thin-bedded sandstone may be present in interbedded sequences. Mudstone and siltstone may be thinly interbedded within these sequences.
5. Ripple-laminated sandstone most commonly occurs on top of the massive-bedded and thin-bedded sandstone channel-fill deposits. It is also interbedded with horizontal and tabular cross-beds of the thin-bedded sandstone.
6. Sharp-based sandstone is very minor and where present commonly is interbedded with the mudstone and siltstone.

Thickness of the individual facies in this upper sequence is extremely variable. Mudstone and siltstone may occur as beds a few centimeters thick or several meters of beds several centimeters to tens of centimeters thick. The same is true for the ripple-laminated, thin-bedded, and massive-bedded sandstones. Systematic cycles of vertical facies arrangements are not observed. Overall, the arrangement of facies demonstrates a general coarsening-upward sequence. Cyclically bedded mudstone and siltstone and very fine-grained sandstone at and near the base of the formation pass upward through very fine-grained

sandstone into fine to medium-grained sandstone in the upper portion of the formation.

The facies of the transition zone crops out sporadically and is not traceable horizontally for distances more than a few tens of meters. The overlying thick-bedded, laminated sandstone forms a prominent and laterally persistent sequence which is traceable for distances in excess of 5 km parallel to the strike.

The uppermost facies grouping is characterized by a general lack of lateral persistence of the facies. This is in part due to the geometric nature or channel morphology of the massive-bedded sandstone and trough cross-beds of the thin-bedded sandstone. Locally, thick beds of mudstone and siltstone, thin-bedded sandstone, or ripple-laminated sandstone are the lateral stratigraphic equivalent of these two facies. Lateral transition between these is sharp, and gradational textural changes are not observed. Also locally, ripple-laminated sandstone is gradationally traced into thin-bedded sandstone. Horizontal beds of the thin-bedded sandstone are locally and gradationally traceable into trough cross-beds of the thin-bedded sandstone. The channelized nature of the deposits which characterizes the second facies grouping is considered the reason for the lack of horizontal persistence of these facies. Channels of the massive-bedded sandstone are the largest of those observed, and the maximum width of a channel fill of this facies is observed to be slightly less than 1 km.

Detailed field mapping, measurement of partial stratigraphic sections, and construction of geologic cross sections suggests that the first facies grouping thickens from east to west. This suggests that portions of the delta front deposits are partial stratigraphic equivalents of the delta plain.

Paleocurrent Data

Azimuths of a total of 763 cross-bed sets were measured from the sharp-based, laminated, and thin-bedded sandstone facies. One measurement was taken per cross-bed set. Tilted bedding was rotated to horizontal following the procedures of Ramsay (1961) and Potter and Pettijohn (1977). Current rose diagrams were constructed from the tilt-corrected data for the different facies and different vertical and lateral positions within the stratigraphic column. For each diagram, vector mean, vector magnitude or strength, and the percent vector magnitude was calculated. These methods and their results are concluded to be adequate for the purposes of this study, in that they sufficiently discriminate and illustrate the salient differences between facies and within facies.

Delta Front

Lower Delta Front. The lower delta front environment is characterized by the transition zone between the Seboomook Group and the Matagamon Sandstone, and those portions of the thick-bedded, laminated sandstone that contain the *Beachia* and *Coelospira-Leptocoelia* Communities. The facies which comprise this environment are the mudstone and siltstone, sharp-based sandstone, thin-bedded, laminated sandstone, thick-bedded, laminated sandstone, and laminated-shelly sandstone.

Paleocurrent Patterns. Paleocurrent patterns of this facies grouping indicate, in general, a variable dispersal pattern. The variability is due to the differences in the dispersal pattern between the sharp-based and thin-bedded, laminated sandstones. Measurements from both facies are from small-scale (less than 6 cm) sets of cross-laminae. Azimuths for the sharp-based sandstone have an average vector, for 35 measurements, of N75°W (Fig. 8a). The vector magnitude is 67%. Flute casts are uncommon; however, those measured indicated a direction of current flow ranging between N35°W and N82°W.

Thin-bedded, laminated sandstone paleocurrent patterns indicate a different paleoflow direction and a much smaller percent vector magnitude. Figure 8b shows the results from 51 sets measured in the thin-bedded, laminated sandstones. Average vector direction is S51°E and the vector magnitude is 31%.

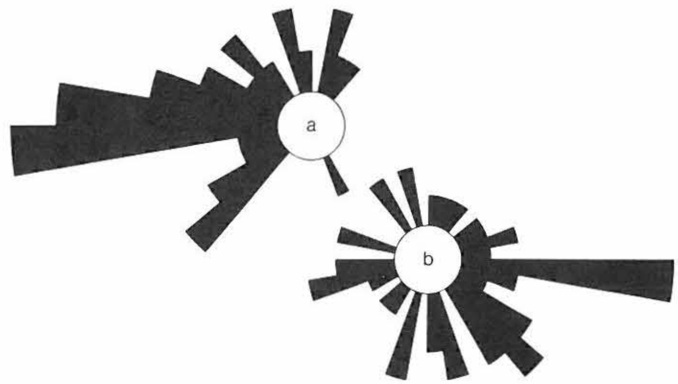

Figure 8. Current rose diagrams for the lower delta front environment. (a) is for the sharp-based sandstone while (b) is for the thin-bedded-laminated sandstone.

Faunas. Faunas collected from the lower delta front are assigned to the *Beachia* and *Coelospira-Leptocoelia* Communities. *B. thunii*, as previously mentioned, is a sessile, medium-sized, centronellid. The sessile nature of this form suggests lower energy conditions where the substrate was not affected by continual erosion and redeposition. Parallel and shallow trough cross-bedding of the thin-bedded, laminated sandstone from which this community was collected indicate energy conditions comparable to the lower flow regime. This community has previously been interpreted to occur in an offshore or "deeper water" environment (Boucot and Johnson, 1967; Boucot, 1975, 1982; Lesperance and Sheehan, 1975). The stratigraphic position of the *Beachia* Community within the Matagamon Sandstone, the interpreted depositional environment, and depositional mechanisms of the units in which this community occurs, support this conclusion.

Faunas of the *Coelospira-Leptocoelia* Community are secondary members found in association with the *Beachia* Community and probably lived in association with *Beachia*. These

faunas along with the *Beachia* Community occur in the lower delta front and are associated with sediment and gravity flow deposits of the Seboomook Group. Where these faunas are found in the Seboomook Group they are interpreted as transported from the lower delta front into the prodelta marine slope environment.

Interpretation. The lower delta front of the Matagamon delta is locally characterized by five key characteristics. These include: 1) cross and parallel lamination in the sandstone facies, 2) finer-grained textures of the sandstones, 3) individual beds of non-uniform thickness, 4) bioturbation of all facies, and 5) abundant macerated plant debris.

Processes which were responsible for the deposition of sediments in the lower delta front are interpreted to have been northwest-directed turbidity currents (Fig. 8a), east-directed storm, wave surge or near shore traction currents (Fig. 8b), and suspension. These processes are suggested by the alternating rock types, sedimentary structures, and dispersal patterns. The sharp-based sandstone is interpreted to have been deposited by turbidity currents. Sedimentary structures such as flute and groove casts and sequentially arranged structures (which consist of massive or graded zones overlain by parallel or ripple cross-lamination) are consistent with structures reported in other well documented turbidity current deposits such as those described by Bouma (1962) and Dzulynski and Walton (1965). Turbidite sands and silts have been reported from inner-shelf or prodelta environments and are also interpreted to have been deposited from turbidity currents. The dispersal pattern for the sharp-based sandstone is strongly unimodal to the west (Fig. 8a). This is consistent with dispersal patterns of turbidite sandstones from the prodelta-marine slope of the Seboomook Group (Hall and Stanley, 1973; Hall et al., 1976; Pollock, 1987).

The thin-bedded, laminated sandstone is interbedded with the sharp-based sandstones. The thin-bedded, laminated sandstones differ primarily in greater bedding thickness, lack of sequentially arranged sedimentary structures and sole marks, and dispersal patterns. Textural grading occurs commonly in the upper few centimeters of the bed which commonly overlies parallel or cross-lamination. The origin of swaley cross-stratification similar to that observed in this subfacies is unclear but appears to be related to storm waves (Leckie and Walker, 1982; Duke, 1985). The interpretation is that swaley cross-stratification is formed in environments shallower than fair weather wave base (Leckie and Walker, 1982). If this interpretation is correct, then the lower delta front of the Matagamon delta was probably fairly shallow water.

Both of these sandstone facies are presumed to have originated in the delta plain and were most probably transported onto the lower delta front by storm-generated currents, seasonal floods, or peak discharges affecting distributaries and the upper delta front. The thin-bedded, laminated sandstones were reworked and transported shoreward, possibly during storms. Similar processes are recognized in several modern deltas (Coleman and Prior, 1980; Donaldson et al., 1970; Hayes, 1967; and Swift, 1969). However, the paleocurrent patterns suggest that depositional processes differed somewhat for each sandstone. Figure 8 clearly demonstrates a more random orientation of cross-bed vectors in the thin-bedded, laminated sandstone (max. vector mag. 31%) than in the sharp-based sandstone (max. vector mag. 67%). The thin-bedded, laminated sandstone dispersal pattern is characteristic of dispersal patterns of the overlying thick-bedded, laminated sandstone.

Suspension is believed to be the major mechanism of deposition of the mudstone and siltstone within the lower delta front. Thin silt laminations are rarely present. Suspension is the major mechanism of deposition of parallel-laminated silts and clays in the prodelta and delta front environments of modern deltas (Coleman, 1976; Reineck and Singh, 1980) and their ancient counterparts (Collinson, 1969; Walker and Sutton, 1967).

Upper Delta Front. This environment is characterized by the thick-bedded, laminated sandstone and to a much lesser extent by the thin-bedded sandstone and shelly-laminated sandstone. The environment is represented by between 700 m and 1200 m of sandstone and is laterally persistent for distances exceeding 5 km. The large sediment volume, great bed thickness, abundance of cross and parallel laminae sets, and homogeneity of grain size suggests a large and continuous sediment supply of uniform grain size being acted upon by a constant current system. Concentration, disarticulation, and orientation of marine fossils along local erosion surfaces indicate reworking of the sediment by current in a marine environment.

Paleocurrent Pattern. Current rose diagrams based on 302 cross-bed set measurements are shown for equivalent or near-equivalent lateral stratigraphic horizons and different vertical horizons throughout the upper delta front (Fig. 9). This was performed to determine if the dispersal pattern is random throughout, or whether it exhibits a preference based on a lateral or vertical position within the environment. Figure 9 shows readings from rocks that underlie or are the stratigraphic equivalents of the delta plain. Figures 9e and 9f are from the uppermost 400 m of the Matagamon Sandstone underlying the Traveler felsite. Thick-bedded, laminated sandstone paleocurrent patterns exhibit vector magnitudes which range from 16% to 39%. A composite vector direction and magnitude for these paleocurrents is N78°E and 16% respectively.

Currents in the delta front were polymodal as is indicated by the various current rose diagrams. Distributary, tidal, near shore, and possibly storm- or wind-driven currents are considered to be responsible for this dispersal pattern. Currents of similar origin are extensively documented on modern delta fronts (cf. Coleman, 1976; Coleman and Prior, 1980; Klein, 1967). The cross-bedding typical of this facies is similar to the swaley cross-stratification discussed by Leckie and Walker (1982) and quite possibly hummocky cross-stratification, but the radial domal structures are not observed. This is possibly due to erosion which would preferentially preserve swales over domal tops. If these features were originally formed below fair-weather wave base as hummocks during storms, the interpretation of the Matagamon delta as having been deposited in a storm- or tide-dominated environment where reworking and redistribu-

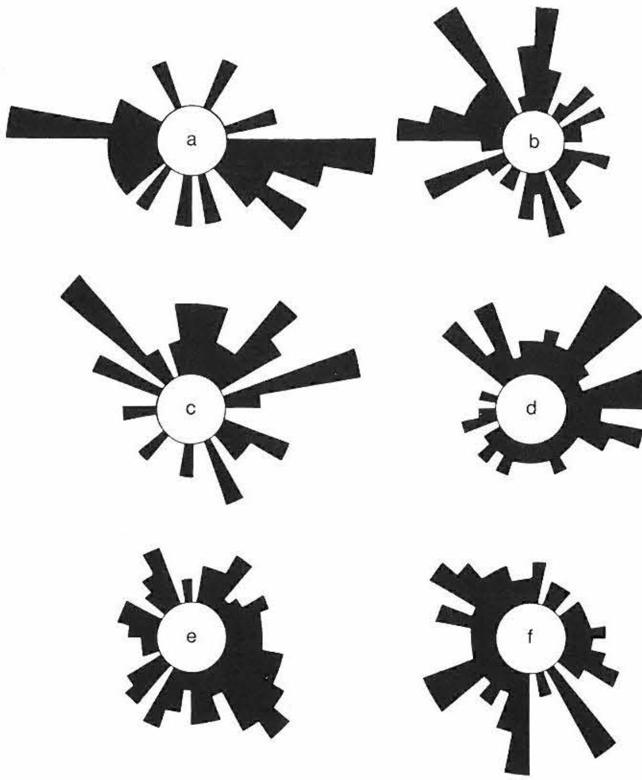

Figure 9. Current rose diagrams for the upper delta front environment.

tion of distributary-supplied sediment are important (Miall, 1984; Leckie and Walker, 1982) would be reinforced.

Faunas. Faunas collected from the upper delta front are assigned to all benthic faunal (except the *Beachia*) communities. The communities in this environment are dominated by free-living rather than sessile forms. The approximate vertical arrangement of the communities is from *Coelospira-Leptocoelia* through *Leptostrophia*, into Homalonotid-*Plectonotoides*. The *Tentaculites* Community occurs throughout the delta front. Faunas of the *Cloudella* Community are believed to have been transported from the delta plain environment. This vertical relationship suggests that these communities were most probably environmentally influenced by some aspect of the delta. The arrangement indicates that the communities were most probably controlled by water depth, distance from shore, or strength and nature of currents. Other controls likely to have exerted influences over the communities are not recognizable or preserved in this rock.

Interpretation. The vertical and horizontal sequence of facies, sedimentary structures, dispersal patterns, and uniform rock type in conjunction with the sedimentary features of the overlying delta plain suggest that the delta front was probably associated with a non-barred or non-barrier coast. In that setting the dominant processes consisted of the reworking of distributary-supplied sediment by storm or tide-generated marine currents.

Similar dispersal patterns and sedimentary structures occur in the non-barred, high-energy, near-shore environment (Clifton et al., 1971). Also, Reineck and Singh (1980) summarize the dispersal patterns and sedimentary structures of coastal sand bodies and shelves, which conform, in general, to those observed in the delta front of the Matagamon Sandstone. Coleman (1976) and Coleman and Prior (1980) discuss the different modern delta types. Those most similar to the observed sedimentary features of the Matagamon delta front have relatively high tidal ranges and wave energies, and the primary sediment consists of river-supplied sand. The sediment is reworked and redistributed by marine processes.

Tide or wave dominated currents produce bipolar or bimodal patterns similar to Fig. 9a of the Matagamon Sandstone. These have been documented in delta front or tidally influenced shelf environments (Klein, 1967, 1972; Young and Long, 1978; and Coleman, 1976; among others). Also, Van de Graaff (1972) proposed landward-flowing bottom currents from the delta front facies of the Castlegate Sandstone. Similar landward flowing currents (Fig. 9e) are interpreted from the dispersal patterns of the Matagamon Sandstone. We conclude that marine currents prevailed over distributary currents in the upper delta front. Substantial reworking of sediment by easterly flowing nearshore marine currents occurred as the currents impinged on the delta front. The net effect of the reworking was to produce a dispersal pattern indicating landward redistribution of sediment.

Delta Plain

General. The delta plain environment beds reach a maximum thickness of 570 m. The vertical and horizontal arrangement of individual facies is not persistent for more than a few meters or few tens of meters. This is generally interpreted to be the result of laterally migrating or meandering distributaries. Distributary channel fills and channel bars are the primary preserved features of the delta plain. The delta plain bay environment, as identified by the mudstone and siltstone facies, is a minor component of the delta plain.

Delta Plain Distributaries. Distributary channels of the Matagamon Sandstone are composed ideally of a basal lag shale-pebble conglomerate overlain by massive-bedded or trough cross-bedded, thin-bedded sandstone. This in turn is overlain by horizontal or tabular cross-beds of the thin-bedded or ripple-laminated sandstone. Specific vertical sequences have been discussed previously (Fig. 7). The channels are of two sizes. The larger are represented by channel fills of massive-bedded sandstone with maximum preserved fill slightly less than 1 km wide and 15 m thick. Smaller channels are represented by the thin-bedded sandstone facies (Fig. 5). Maximum preserved fill of these channels is 20 m wide and 2 m thick. Arrangement of sedimentary structures and textures within channels reflects fining-upward sequences and changes in flow regime.

Paleocurrent Pattern. Analysis of 230 medium- and large-scale cross-beds from the thin-bedded sandstone facies shows an average vector of N85°W and magnitude of 52%. Figure

10 shows current rose diagrams from five vertically and laterally separated parts of the thin-bedded sandstone facies. The dispersal pattern for this environment differs from the delta front in that it is more nearly unimodal and average vectors are consistently to the northwest. The average vectors range between N84°W and N24°W. The vector magnitudes also show less dispersion, ranging from 33% to 61%, with four of the vectors exceeding 46%.

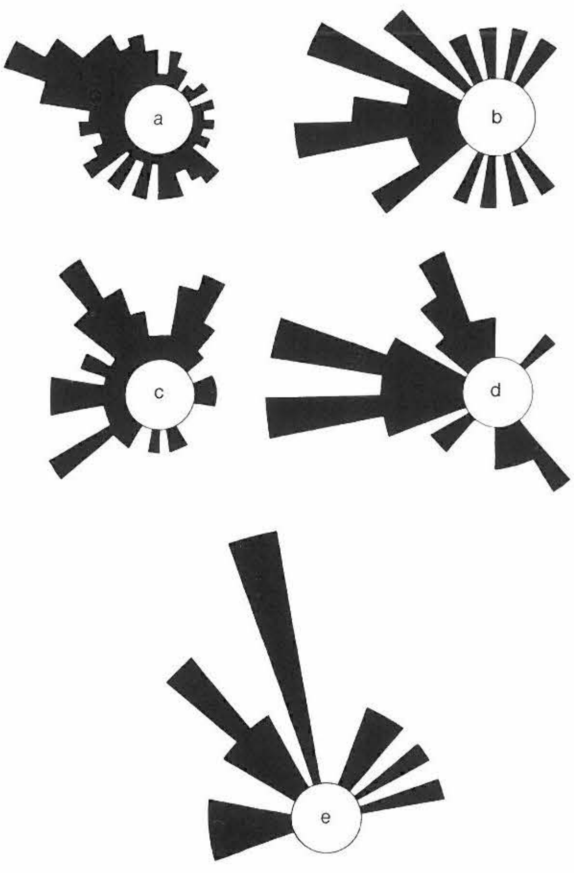

Figure 10. Current rose diagrams for the delta plain environment.

Because of the multidirectional nature of the dispersal pattern of the delta front, the importance of a more unidirectional dispersal pattern of the delta plain is heightened. It is concluded from the dispersal pattern of this facies grouping that the source area for the Matagamon Sandstone lay to the southeast. The northwesterly dispersal pattern of the delta plain environment is a better indicator of paleoslope than the dispersal patterns of the delta front because of the substantial sediment reworking in the latter environment. It supports the argument for westward progradation during Matagamon delta deposition.

Long and Young (1978) discuss the use of cross-stratification dispersion as a tool to augment the interpretation of sedimentary environments. From Long and Young's study, and from the paleocurrent data summary provided by Potter and Pettijohn (1977), it would appear that more unimodal dispersal patterns are characteristic of fluviatile or channeled environments, and polymodal dispersal patterns are more characteristic of marine or "mixed" sedimentary environments.

Faunas. Faunas from the distributaries of the delta plain environment are assigned to the *Cloudella* Community. This community occurs at the base of channels usually overlying mudstone and siltstone of the bay environment discussed below. One collection also contains *Plectonotoides*. *Cloudella* is also found to be a minor element of the upper delta front in rocks stratigraphically below the delta plain. The *Cloudella* Community is essentially a single taxon community. *Cloudella* is a small sessile mutationellid and most probably lived in the bays or channels of the delta plain. Its sessile nature probably made it more adaptable to the bay environment where reworking of the sediments by relatively high energy currents and sediment influxes was minimal. The occurrence of the *Cloudella* Community in the delta front is considered to represent transportation of the faunas out of the delta plain onto the delta front during periods of peak distributary flow.

One control which may have influenced the distribution of the *Cloudella* Community, in addition to energy and substrate, is salinity, where bay water presumably had salinity less than normal marine as a result of dilution by the distributaries and precipitation.

Interpretation. The interpretation is that these deposits were produced by vertical and perhaps lateral accretion of migrating channels (Allen, 1964, 1965a,b; Collinson, 1969; Kelling, 1969; Smith, 1970; and Visher, 1965). Deposition in the larger channels was probably in the upper flow regime and in the smaller channels from the lower flow regime. This interpretation is based upon comparison of the sedimentary features of the channels of the Matagamon Sandstone with those generated experimentally where stratification is used as an indicator of flow environment (see for example, Harms and Fahnestock, 1965; Simons et al., 1965; Middleton and Southard, 1978; Reineck and Singh, 1980; for discussions of flow regimes, bed forms, and stratification types). Channel deposits locally are overlain by possible point and longitudinal bar deposits. These are represented by interbedded horizontal beds, planar cross-beds and ripple-laminated sandstone where bar growth occurred by avalanche deposition of the sand within the channel. Interpreted channel bar deposits are less than 2 m thick. These observed sequences partially correspond to the fluvial and deltaic channel models discussed by Allen (1964), Collinson (1969), Miall (1984), McGowen and Garner (1970), Smith (1970), Visher (1965), Reineck and Singh (1980), and Reinson (1984).

Faunas (*Cloudella* Community) associated with the channel lag conglomerates might suggest that these channels may have been partially influenced by tides, rather than being totally fluvial in nature. Lag deposits in the form of shale-pebble conglomerates at the base of tidal channels have been described by Johnson and Friedman (1969), Oomkens and Terwindt (1960), Reineck and Singh (1980), Reinson (1984), and Van Straaten (1954). However, in the Matagamon channels there

is a complete lack of herringbone cross-stratification, bimodal dispersal pattern and intense bioturbation that is characteristic of tidal channels, (Miall, 1984; Reineck and Singh, 1980; Klein, 1976; Reinson, 1984).

Distributary Plain Bays. The distributary plain bay environment is represented by the mudstone and siltstone primarily, and sharp-based sandstone facies. The mudstone and siltstone is characterized by rare macerated plant fossils. Thin silt or sand laminae are characterized by flaser bedding, small-scale ball and pillow structures, and small-scale ripple cross-lamination. The uncommon sharp-based sandstone is usually graded in the lower interval, which in turn is overlain by an intensely bioturbated zone containing uncommon, unidentified gastropods. Upper bedding surfaces are irregular and beds commonly pinch and swell. Both facies underlie or overlie distributary deposits (Figs. 4, 5).

Interpretation. The mudstone and siltstone facies and the sedimentary structures which occur within the facies are believed to have been deposited in bays or lagoons adjacent to the distributaries. Sediment was intermittently introduced, as is evidenced by the rippled silt and sand laminations, ball and pillow structures, and the sharp-based sandstone. Currents operating in the lower flow regime were presumably responsible for the deposition of silts with small-scale structures. The sharp-based sandstone is concluded to represent crevasse splay deposits resulting from flooding of distributaries over their banks. The sediment types and sedimentary structures of the bay and crevasse splay deposits of the Matagamon are similar to those discussed by Reineck and Singh (1980), Coleman and Prior (1980), Morgan (1970), and Johnson and Friedman (1969), for modern and ancient delta plain bays and lagoons.

DISCUSSION

The Matagamon Sandstone is presented as an example of deltaic sedimentation in a relatively high energy non-barred coastal setting during the Lower Devonian (Siegenian) in northern Maine. Figure 11 summarizes the Matagamon-Seboomook delta. The sequence of the two facies associations coupled with the paleocurrent data and the environmental interpretations records a westerly prograding delta complex with a prodelta-marine slope (Seboomook Group), (Hall and Stanley, 1973; Hall et al., 1976) at the base passing upward through the delta front into the delta plain. Spatial relationships between the depositional environments are relatively simple and appear to represent a shoreline migration to the northwest. This may have been modified in the later depositional stages by contemporaneous volcanism to the west of the delta in late Siegenian through Emsian time. Rankin (1968) documents the existence of a paleocaldera structure at the northern end of the Piscataquis volcanic belt, and to the immediate west of the outcrop area of the Matagamon Sandstone. Volcanic products associated with this caldera are the Traveler felsite, which stratigraphically overlies the Matagamon Sandstone and crops out to the west of the Matagamon Sandstone. Contemporaneous volcanism associated with the later depositional stages of the Matagamon Sand-

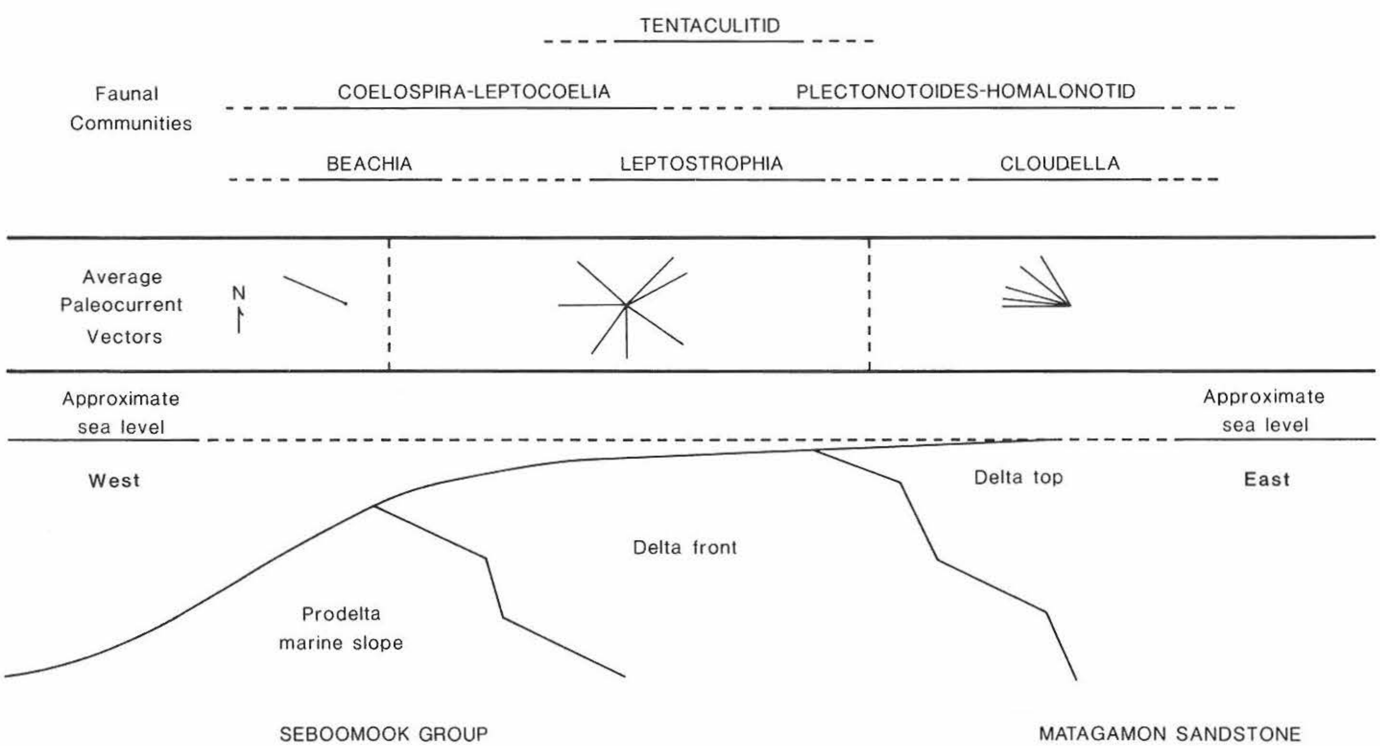

Figure 11. Summary diagram showing the schematic distribution of the major sedimentary environments, average paleocurrent vectors and faunal communities from the Matagamon-Seboomook delta complex.

stone is represented by an increase in the volcanogenic component of the sandstone.

The vertical sequence is also interpreted as shallowing upward from the deeper water of the prodelta-marine slope of the Seboomook Group to the delta plain as represented by distributary fill and bay sediments.

The delta front environment is interpreted to have been extensively reworked by easterly flowing nearshore currents. The faunas of the delta front are spatially distributed such that the *Beachia* Community is overlain by, and associated with, faunas of the *Coelospira-Leptocoelia* Community. The latter is overlain by, or the horizontal equivalent of, the *Leptostrophia* and Homalonotid-*Plectonotoides* Communities respectively (Figs. 6 and 11). The *Cloudella* Community is present in the delta plain environment.

Ecological constraints on these communities, as inferred from sedimentary parameters, presumably included substrate, salinity, and water depth and energy conditions.

Substrate character is not considered to be a major ecological control. The major rock type is uniformly fine to very fine-grained sandstone in all of the depositional environments. The sand units from which the *Cloudella* and *Beachia* Communities were collected are, however, interbedded with mudstones, and it is probable that elements of these two communities tolerated or lived in conjunction with multiple sediment types. The other communities were collected from environments in which sand was the sole accumulating sediment.

Among the inferred ecological parameters, energy conditions are considered important in faunal distribution. *C. matagamoni* and *B. thunii* are attached forms found on the delta plain and lower delta front respectively. The sessile nature of these faunas suggests relatively lower energy conditions where the substrate was not susceptible to continual erosion and redeposition.

The other communities, *Coelospira-Leptocoelia, Leptostrophia,* Homalonotid-*Plectonotoides*, and *Tentaculites* are dominated by free-living forms found within the delta front. The delta front is characterized by complex dispersal patterns believed to be the result of dominantly southeasterly (landward) flowing marine currents which reworked sediment supplied by northwesterly flowing distributaries. We suggest that the complex current and energy systems partially controlled the composition of these communities. The *Coelospira-Leptocoelia* Community occurs in conjunction with members of the Beachia and *Leptostrophia* Communities, near the stratigraphic base and top of the Matagamon Sandstone. Near the stratigraphic top it is presumed to occur in the part of the delta front that is the lateral equivalent of the delta plain. Dispersal patterns associated with this community are to the southeast, versus a northwesterly dispersal pattern of the delta front in areas from which this community was not collected. We conclude that the *Coelospira-Leptocoelia* Community inhabited areas away from distributary influence. The *Leptostrophia* Community occurs in conjunction with members of the *Coelospira-Leptocoelia* and Homalonotid-*Plectonotoides* Communities. The Homalonotid-*Plectonotoides* Community occurs stratigraphically below the delta plain. Dispersal patterns in rocks containing the *Leptostrophia* and Homalonotid-*Plectonotoides* Communities have average vectors to the northwest. These two communities inhabited areas that were presumably under distributary influence. The stratigraphic structuring of benthic faunal community distribution across the delta front is in accordance with proximity to shoreline (i.e. delta plain).

Salinity may also have exerted an ecologic control over the faunas of the Matagamon Sandstone. If salinity distribution within the Matagamon delta paralleled that of modern deltas, then the *Cloudella* Community living on the delta plain presumably experienced less than normal salinity due to dilution from the distributaries. This may also have been true for the Homalonotid-*Plectonotoides* and perhaps the *Leptostrophia* Community, while *Coelospira-Leptocoelia* and *Beachia* Communities lived in areas with "normal" salinities.

Other investigations (Ziegler, 1965; Boucot and Johnson, 1967; Boucot, 1975; among others) have related Lower Devonian faunal communities to water depth rather than other types of ecological controls. This study tends to support the conclusion that community distribution is largely controlled by depth or relative distance from shore. Salinity and depth presumably increased across the delta, structuring community distribution as shown in Figure 11. However, energy conditions were also a factor in community distribution.

An alternative to the ecologic controls for community structure is the hypothesis that communities were structured by a physical mixing of the dominant community elements. The general problem of transported faunas needs to be considered due to the complex dispersal pattern of the Matagamon Sandstone. The faunal distribution, as indicated in Table 2 and Figure 6, demonstrates that the various faunas are not uniformly distributed throughout the sedimentary environments, but rather are preferentially distributed within a particular sedimentary environment. Walker and Bambach (1971) concluded that macrobenthonic death assemblages represent time-averaged life assemblages. Also, Schaefer (1972) and Turney and Perkins (1972) have found that transport across sedimentary environments is not appreciably significant. The data of Table 2 and Figure 6 suggest that the communities of the Matagamon Sandstone represent time-averaged life assemblages with minimal physical transport into sedimentary environments different from that in which they lived.

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Occurrence of the Crinoid Rhodocrinites nortoni *(Goldring) from the Lower Devonian Seboomook Formation in the Telos Lake Area, North-Central Maine*

J. Steven Kite
Thomas W. Kammer
Department of Geology and Geography
West Virginia University
Morgantown, West Virginia 26506

ABSTRACT

Specimens of the camerate crinoid *Rhodocrinites nortoni* (Goldring) have been recovered for the first time from the Siegenian (Lower Devonian) Seboomook Formation and are described and illustrated. Two species of brachiopods and arms of the crinoid *Ctenocrinus* sp. also were found at the same locality. The mode of preservation of the arms of both crinoid genera suggests that they were buried alive during a rapid sedimentation event, as might be expected in a flysch sequence such as the Seboomook Formation.

The only other reported occurrence of *R. nortoni* is in a glacial erratic containing the species holotype. Contrary to previous conclusions, the erratic that contains the holotype cannot confidently be assigned to the Siegenian Tarratine Formation and may have been derived from the equivalent-age Seboomook or Matagamon Formations. *Rhodocrinites nortoni* is the oldest known representative of its genus; *Rhodocrinites* is more typically found in Lower Carboniferous rocks.

INTRODUCTION

The Seboomook Formation is a widespread bedrock unit underlying more than half of northwestern Maine (Osberg et al., 1985). Cyclically layered dark slate and sandstone dominate the Seboomook Formation (Boucot and Heath, 1969), but there is considerable lithologic variability and its internal stratigraphy is not well established (Pollock, 1983). Several papers have urged careful, detailed work on the stratigraphy of the Seboomook Formation (Boucot, 1970; Roy, 1980; Pollock, 1983), but progress has been slow because of a lack of marker beds and structural repetition due to near-isoclinal folding. Biostratigraphy has been of limited use in sorting out the Seboomook Formation because it is virtually devoid of macrofossils throughout much of its outcrop belt. In the southern end of the outcrop belt of the formation, where macrofossils are most common, brachiopods dominate collections (Boucot and Heath, 1969; Hall, 1970) and there is little evidence of other faunal elements. Thus, almost any fossil locality in the Seboomook Formation is noteworthy, especially if the locality contains fauna other than brachiopods.

This paper will focus on the occurrence of the crinoid *Rhodocrinites nortoni* (Goldring) at a locality in the Seboomook Formation in the Telos Lake area, Maine (Fig. 1). The larger problems of the internal biostratigraphy of the Seboomook Formation are beyond the scope of this paper, but we hope this paper will help in their eventual resolution.

THE CRINOID LOCALITY

The crinoid-bearing fossil locality was discovered during mapping of the surficial geology of the Telos Lake 15-minute quadrangle in July 1984 (Locality SK-84-508 in T6 R11 WELS; UTM: 5114700 N, 489750 E, Zone 19; 69° 07′ 45″ W, 46° 11′ 15″ N; see Fig. 1). The locality is in a shallow quarry in fine-grained sandstones and siltstones used for road material. The beds containing the crinoids were disturbed by quarry activity, but all of the specimens were obtained from a 1.5 m by 3 m strip in which virtually all clasts contained fossil crinoids. All of the crinoids in our original collection were of the same species, *R. nortoni* (see systematic paleontology, Appendix A). Fossil brachiopods, *Acrospirifer murchisoni* (Castelnau) and

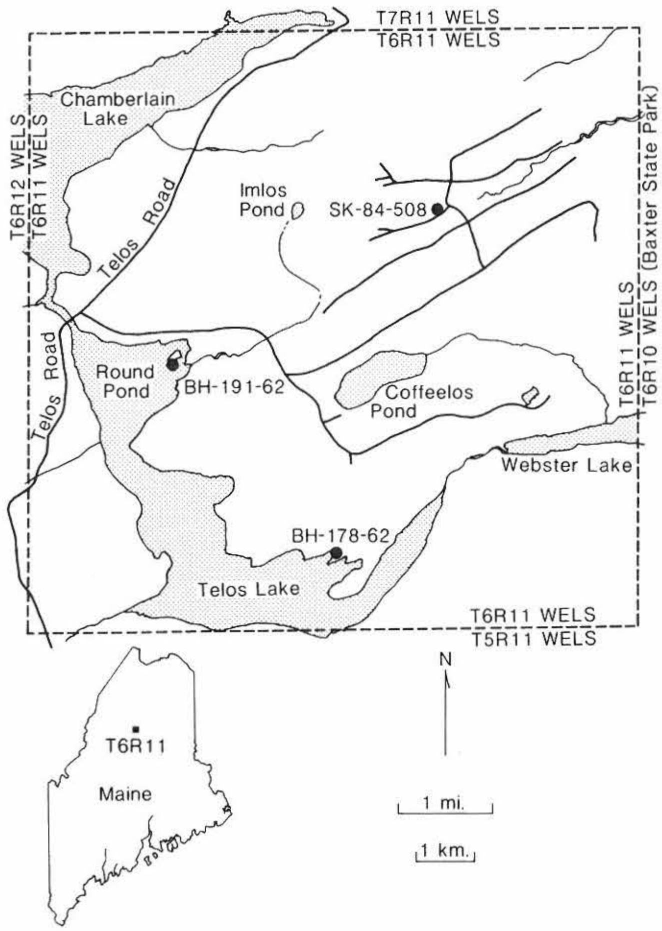

Figure 1. Location map of T6 R11 WELS, including the *Rhodocrinites nortoni* locality (SK-84-508). Crinoid localities reported by Hall (1970) are designated by BH. Selected passable roads (as of 1984) are shown.

Leptostrophia cf. *L. magnifica* (Hall), were obtained from the same general area in the pit. No erratic lithologies occurred in the quarry, so it is assumed that the fossiliferous clasts came from a single lens or bed exposed during the quarry operations and that the fossils were not moved more than a meter or two by the quarry operations.

Thomas Lowell collected additional specimens from the same locality in 1986 and has donated his collection for our study. Photographs and field descriptions show that his collection was from the same 1.5 m by 3 m strip. In addition to species in our original collection, Lowell's collection also includes two occurrences of arms from the crinoid *Ctenocrinus* sp. *Ctenocrinus* is very distinctive and can be recognized by the fused inner arms with biserial outer arms. Unfortunately, no *Ctenocrinus* calyces were found, so the specimens cannot be identified to species level.

Rhodocrinites nortoni was first described by Goldring (1933) from a fossil in a Lower Devonian sandstone erratic found at Caratunk, Maine. The holotype and the fossils described in this paper represent the only known occurrences of *R. nortoni*. We have examined the holotype, but cannot determine the formation from which the erratic was derived.

The preservation of the arms of *R. nortoni* indicates that the crinoids were buried alive during a rapid sedimentation event, such as a turbidity current or a storm. The arms are preserved in three dimensions within the rock, indicating that the sediment was water-saturated. The arm plates still contain feathery pinnules. The calyces generally are not well-preserved and show evidence of disruption, perhaps from gases produced by decay. The calyx contains most of the biomass of a crinoid and, thus, would produce the most gas during decay. Two reasonably well-preserved calyces are illustrated in this paper (Figs. 3a-3d). Approximately ten additional calyces and many calyx fragments were recovered, but most are poorly preserved.

Several slabs of rock show a concentration of brachiopod shells on their base and feathery crinoid arms "floating" within the matrix above the brachiopods. This fossil distribution, which also occurs in the erratic containing the holotype, suggests graded sedimentation wherein the arms of the crinoids caused them to settle out after the denser brachiopod shells.

STRATIGRAPHIC SETTING

The locality occurs in an area mapped as Seboomook Formation on the state bedrock geologic map (Osberg et al., 1985). The fossil locality is 1.1 km beyond the area mapped by Hall (1970), but the locality is on strike with his mapped Seboomook Formation outcrops. The lithologies present in the quarry are similar to sandstones and siltstones mapped as Seboomook Formation elsewhere in northern Maine. Bedrock stratigraphy and paleontology were outside the scope of the project for which this field work was conducted, but samples were collected because of the excellent preservation of the crinoids and the general scarcity of fossils in the Seboomook Formation.

The thickness of the Seboomook Formation ranges from 1200 m near Telos Lake to over 5000 m at the western limits of the unit (Neuman and Rankin, 1980; Pollock, 1983). Pollock (1983) has characterized the Seboomook Formation as monotonous, cyclically bedded slates and sandstones – a classical shaley flysch. Hall and Stanley (1973) report that the Seboomook Formation represents submarine-slope and prodeltaic deposits. Levee-bounded channels and Bouma sequences of graded muds and graywackes indicate deposition by turbidity currents (Hall and Stanley, 1973; Pollock, 1983). Boucot and Heath (1969) inferred high rates of deposition for the Seboomook Formation, generally at depths below the bathymetric limits of the typical Lower Devonian fauna in adjacent formations. Scattered shell beds in the Seboomook Formation may be either reworked turbidite deposits, or shallow-water life assemblages associated with isolated structural highs in the basin (Boucot, 1953; Boucot and Heath, 1969).

Several substantial sandstone bodies occur within or adjacent to the Seboomook Formation (Roy, 1980; 1982). Two of the adjacent sandstones are differentiated as separate formations: the Tarratine Formation (Boucot, 1961) and the Matagamon Sandstone (Rankin, 1965) (Fig. 2). These formations have been

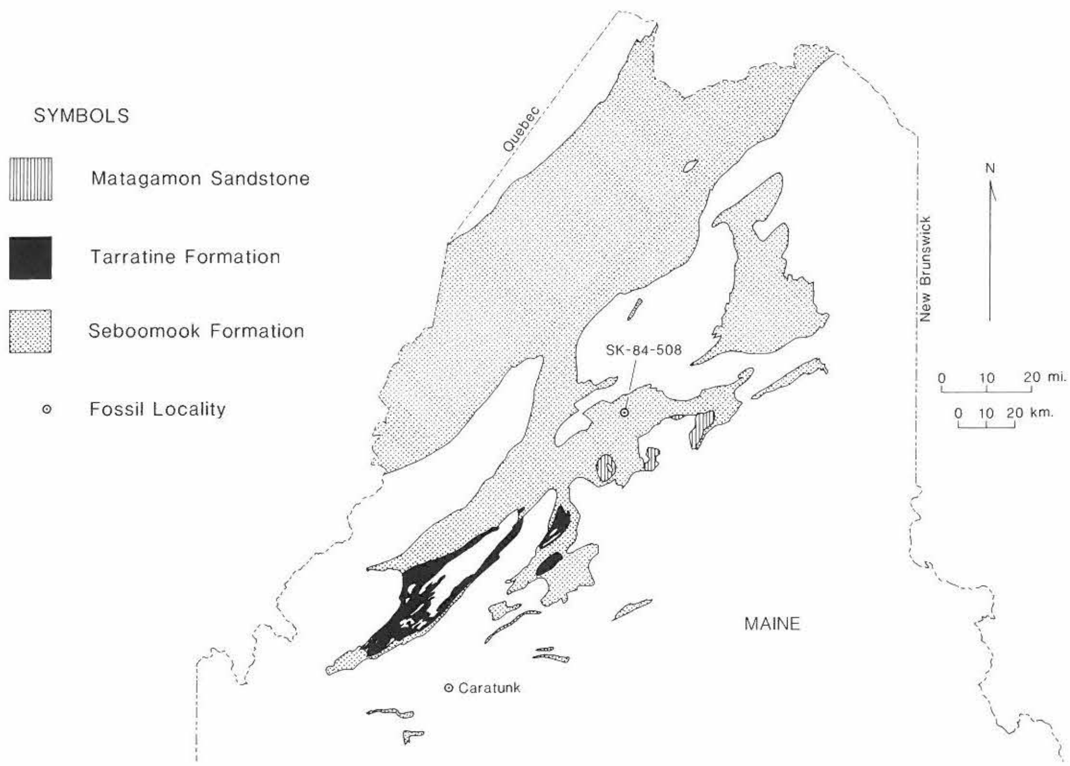

Figure 2. Generalized geologic map showing distribution of Seboomook, Tarratine, and Matagamon Formations north of 45°N latitude (after Osberg et al., 1985). Although lithologically similar, the Tarratine Formation was derived from a source to the west (Boucot and Heath, 1969; Roy, 1980), whereas the Matagamon Sandstone was derived from an emergent terrain in New Brunswick (Roy, 1980; Hall and Stanley, 1973). Fossils described in this paper are from a quarry at SK-84-508, and a glacial erratic found at Caratunk (Goldring, 1933).

interpreted as shallow-water facies in close association with the relatively deep-water basin of the Seboomook Formation (Roy, 1980; Pollock et al., this volume).

Southwest of the Telos Lake area, the Tarratine Formation, which is restricted to the Moose River synclinorium (Fig. 2), was derived from the Lower Devonian Somerset Island and nearby Lobster Mountain Volcanics (Boucot and Heath, 1969; Roy, 1980). The facies, fauna, and lithologies of the Tarratine Formation grade laterally into the Seboomook Formation because the depocenter of the Tarratine Formation intermittently supplied the basin of the Seboomook Formation (Boucot and Heath, 1969). The Tarratine Formation is dominated by dark sandstones similar to those within the Seboomook Formation, and it is, at best, very difficult to differentiate sandstones of the two formations in hand specimen. The gradational contact between the formations is arbitrarily defined; the Seboomook Formation contains less than 50 percent sandstone beds (Boucot and Heath, 1969).

The Matagamon Sandstone, which crops out south and east of the fossil locality (Fig. 2), has been interpreted as delta-top and delta-front deposits, derived from an emergent terrain in New Brunswick (Roy, 1980; Hall and Stanley, 1973; Pollock et al., this volume). The Matagamon Sandstone is dominated by dark sandstones similar to those within the Seboomook and Tarratine Formations (Rankin, 1965; Neuman and Rankin, 1980). The Seboomook Formation interfingers with, and is overlain by, the Matagamon Sandstone. The transition is marked by increases in cross-bedding, bed thickness, and sandstones, and by a decrease in graded bedding (Rankin, 1965). The contact between the formations is arbitrary, based on the Seboomook Formation containing less than 50 percent sandstone beds (Rankin, 1965). As with sandstones from the Tarratine Formation, sandstones from the Matagamon Sandstone and the Seboomook Formation cannot be distinguished in hand specimen because the depocenter of the Matagamon Sandstone was a source of sandstone for the Seboomook Formation.

AGE OF THE SEBOOMOOK FORMATION

Brachiopods found at locality SK-84-508 include abundant specimens of *Acrospirifer murchisoni* (Castelnau) and a few specimens of *Leptostrophia* cf. *L. magnifica* (Hall). According to Boucot and Heath (1969), *A. murchisoni* occurs only in Siegenian (Oriskany) age rocks of the Lower Devonian. Neither Boucot and Heath (1969) nor Hall (1970) reported *Leptostrophia* cf. *L. magnifica* (Hall) from the Seboomook Formation, but Boucot and Heath (1969) report its occurrence in the *Beachia* community in the Siegenian Tarratine Formation, and Rankin (1965) reports it in the Matagamon Sandstone. The fossils discussed in this study are consistent with Dr. Edwin Kirk's con-

clusion that the holotype of *R. nortoni* belongs to the *Beachia* faunal community (Boucot and Heath, 1969). A Siegenian age assignment is consistent with ages determined from other brachiopod localities in the Seboomook Formation of north-central Maine, including three localities within 10 km of SK-84-508 (Hall, 1970), a locality on Grand Lake Matagamon (25 km to the east; Rankin, 1965), and a locality on Gero Island (22 km to the west; Boucot, 1954). Microfossils collected by Pollock from the Seboomook Formation and identified by William Forbes also indicate Siegenian age (Pollock, 1983).

Although a Siegenian age for the Seboomook Formation is well documented in the Telos Lake area, the formation is virtually unfossiliferous and age assignments are tentative throughout much of its wide outcrop belt. Gedinnian (New Scotland) fossils have been reported from the lowermost Seboomook Formation in the Moose River synclinorium; however, all other fossils of the Seboomook Formation from that area are Siegenian (Boucot, 1961). Gedinnian and Siegenian fossils are also reported from the Seboomook Formation in northeastern Maine (Boucot, 1970). Reconnaissance study in northwestern Maine has uncovered three localities in the Seboomook Formation with Siegenian brachiopods (Boudette et al., 1976), but Roy (1982) suggested that parts of the Seboomook Formation in northwestern Maine are as young as Middle Devonian. The Temiscouata Formation in western New Brunswick, which is correlative to the Seboomook Formation, has yielded Emsian (latest Early Devonian: Esopus-Schoharie) brachiopods (St. Peter and Boucot, 1981). All of the fossils reported from the Tarratine Formation or the Matagamon Sandstone are Siegenian (Boucot and Heath, 1969; Rankin, 1965).

OTHER CRINOID LOCALITIES IN THE SEBOOMOOK FORMATION

Hall (1970) reports two crinoid-bearing localities in the Telos Lake area (Fig. 1). *Edriocrinus* sp. occurred in collection BH-178-62 (USNM 11354), made 5.9 km south-southwest of SK-84-508. Hall's collection BH-191-62 (USNM 12464), obtained 5.0 km southwest (along strike) of SK-84-508, included an unidentified crinoid. The crinoid fossil was sent to Dr. G. Arthur Cooper of the United States National Museum (Hall, 1970), but attempts to relocate this specimen have been unsuccessful, and there is no record of its positive identification (Martha Hays, pers. commun., 1986). We are not aware of other crinoid-bearing localities in the Seboomook Formation.

PROVENANCE OF THE CRINOID DESCRIBED BY GOLDRING

The discovery of *R. nortoni* in the Seboomook Formation suggests re-evaluation of the provenance of the crinoid-bearing erratic described by Goldring. Originally, E. S. C. Smith assigned the erratic boulder to the Moose River Sandstone (Goldring, 1933). Later, Boucot (1961) redefined the Moose River to group status and assigned the Oriskany-age sandstones in the Moose River synclinorium to the Tarratine Formation, but he excluded other dark Oriskany-age sandstone units from the Tarratine Formation, particularly the Matagamon Sandstone. Following this revision, Boucot and Heath (1969) stated that the erratic "boulder was undoubtedly derived from the Tarratine Formation to the north as shown by *Leptocoelia flabellites* (this form is present to the north only in Tarratine age rocks)."

Four considerations cast doubt on a Tarratine Formation provenance for the erratic. First, the Moose River Sandstone, as loosely defined early in the twentieth century, included rocks outside of Boucot's Tarratine Formation, such as the Matagamon Sandstone (Rankin, 1965). Hence, Smith's interpretation of a Moose River Sandstone provenance should not be translated directly into a Tarratine Formation provenance. Second, dark sandstone erratics derived from the Seboomook Formation would be virtually indistinguishable from dark sandstone erratics from the Tarratine Formation. The formations differ in amount of sandstone, but not in sandstone lithology (Boucot and Heath, 1969). Third, *Leptocoelia flabellites* also occurs in the Seboomook Formation (Boucot, 1954; Boucot and Heath, 1969; Hall, 1970) and the Matagamon Sandstone (Rankin, 1965), so this species cannot be used to assign a Tarratine Formation provenance.

The fourth consideration is that ice-flow history reveals little about the possible provenance of the Caratunk erratic. Boucot and Heath (1969) implied that the crinoid-bearing erratic was transported by southward-flowing glaciers. Striations and other ice-flow indicators in north-central Maine show ice-flow directions ranging from east-northeast to slightly west of south (Kite and Lowell, 1985; Thompson and Borns, 1985; Flint, 1971). Indeed, the till stratigraphy of eastern Quebec records an early or middle Wisconsin glaciation with westward and southwestward flow from the maritime Provinces into Northern New England and adjacent Canada (Shilts, 1981). It is possible that the crinoid-bearing erratic discovered at Catatunk was transported southwestward from outcrops of the Seboomook Formation or Matagamon Sandstone in the Telos Lake area. In short, the provenance of the erratic studied by Goldring is unclear, and *Rhodocrinites nortoni* cannot confidently be assigned to the Tarratine Formation at this time.

SUMMARY

A shallow quarry in the Lower Devonian Seboomook Formation near Telos Lake, Maine, has yielded specimens of the camerate crinoid *Rhodocrinites nortoni* (Goldring). Two brachiopods, *Acrospirifer murchisoni* (Castelnau) and *Leptostrophia* cf. *L. magnifica* (Hall), and the crinoid *Ctenocrinus* sp. have been collected from the same locality. This is the first report of *R. nortoni*, *Leptostrophia* cf. *L. magnifica* (Hall), and *Ctenocrinus* sp. from the Seboomook Formation. The brachiopods suggest a Siegenian (Oriskany) age for *R. nortoni*, as was proposed by Dr. Edwin Kirk (Boucot and Heath, 1969).

The crinoids appear to have been preserved by rapid burial, possibly because of storm or turbidity currents. Rapid burial

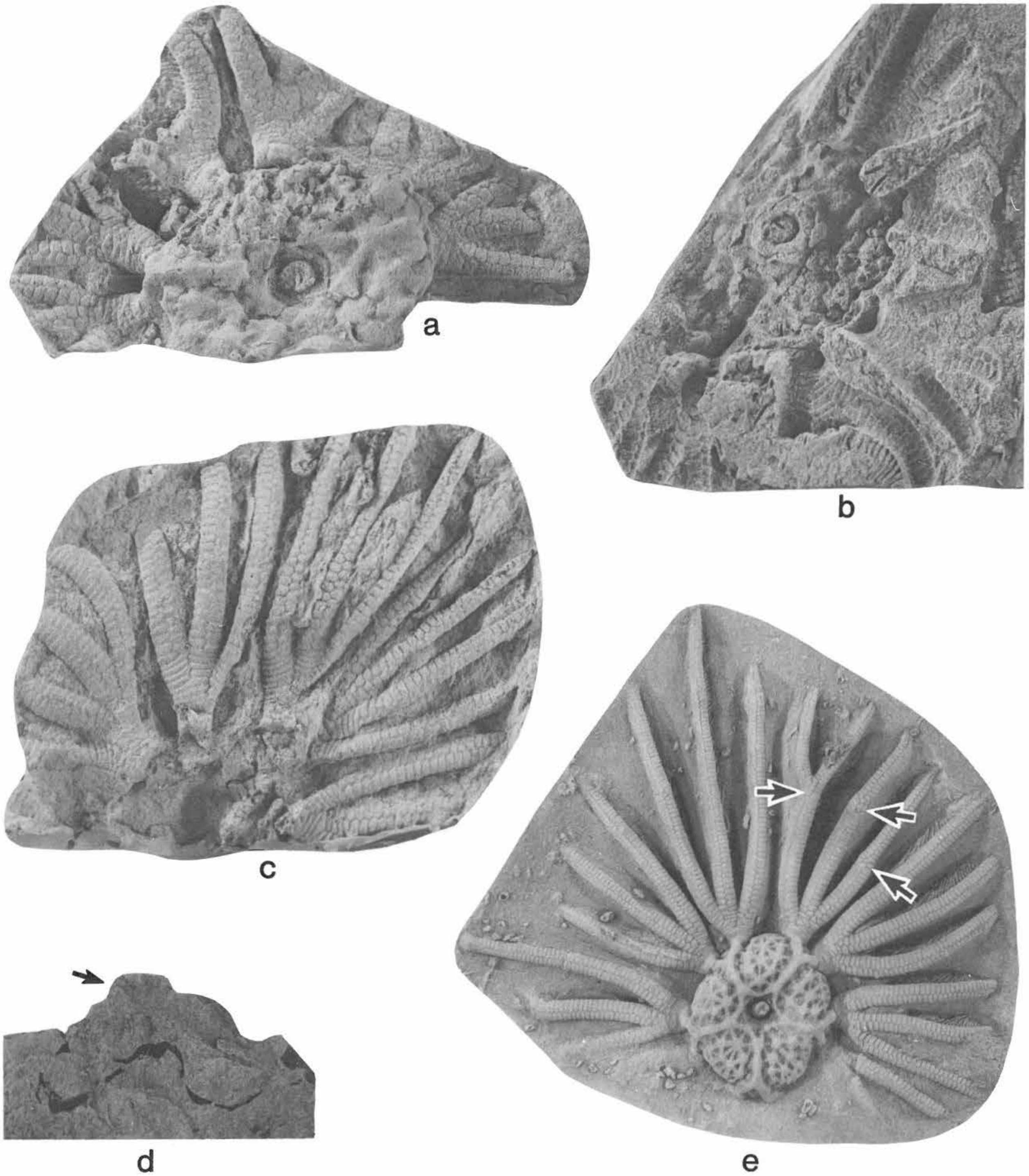

Figure 3. *Rhodocrinites nortoni* (Goldring); all photographs are enlarged 2 times, except e, which is actual size. (a,b) Latex cast and original mold, respectively, of the best preserved specimen from the Seboomook Formation (WVU 1101); note the stem fragment centered on the infrabasals, the radial ridges on the calyx plates, and the arrangement of the arms. (c) Latex cast of upper surface of specimen shown in d (WVU 1102), showing several well-preserved arms as well as a concavity formed by the internal mold of the theca shown in d; note that a set of arms on the left, and a set on the right, each have four branches instead of three; also note the tectonic strain shown by the wide arms on the left and the narrow arms on the right. (d) Outline of theca from specimen (WVU 1102) preserved as an internal mold and exposed on a fractured rock surface; anal tube indicated by arrow. (e) Latex cast of holotype, which is a mold, (University of Maine at Presque Isle, No. 10500); note the regenerated arms on the E-ray indicated by arrows.

and the graded bedding displayed in several crinoid-bearing slabs, including the erratic containing the holotype of *R. nortoni*, are consistent with the general stratigraphic framework for the basin of the Seboomook Formation and the adjacent depocenters associated with the Tarratine and Matagamon Formations. The occurrence of *R. nortoni* in the Lower Devonian of northern Maine apparently marks the first occurrence of the genus *Rhodocrinites* in the paleontological record. The holotype and only other known occurrence of *R. nortoni* is in a glacial erratic from Caratunk, Maine. Boucot and Heath (1969) assigned this erratic to the Tarratine Formation, but the similar lithology of Siegenian sandstones in northern Maine and the area's complex Late Quaternary ice-flow history indicate that the exact provenance of the holotype cannot be determined with certainty.

ACKNOWLEDGMENTS

The authors are in great debt to the field assistants for the 1984 field season: Robert Guice, who discovered the crinoids and insisted they be collected, and Dan Hostettler, who helped collect the specimens. Thomas Lowell kindly donated specimens collected at the same locality in 1986, and William Forbes allowed examination of the holotype of *R. nortoni* from the Paleontology Collection at the University of Maine at Presque Isle. This work benefited from discussion and encouragement from Arthur Boucot, Bradford Hall, and Stephen Pollock, and from thoughtful reviews of an earlier version by Boucot, William Ausich, and Dale Springer. However, we are responsible for all errors in the text. Support for the surficial geology field work came from the Maine Geological Survey.

APPENDIX A. SYSTEMATIC PALEONTOLOGY

Class CRINOIDEA Miller, 1821
Subclass CAMERATA Wachsmuth and Springer, 1886
Order DIPLOBATHRIDA Moore and Laudon, 1943
Suborder EUDIPLOBATHRINA Ubaghs, 1953
Superfamily RHODOCRINITACEA Roemer, 1855
Family RHODOCRINITIDAE Roemer, 1855
Genus RHODOCRINITES Miller, 1821

Rhodocrinites nortoni (Goldring) 1933

Figures 3a-e, 4; Table 1

Rhodocrinus nortoni Goldring, 1933, p. 153-155, Pl. 3, Fig. 1-2, Pl. 4, Fig. 1-2.

Rhodocrinites nortoni (Goldring), Bassler and Moodey, 1943, p. 664; Webster, 1973, p. 232.

Diagnosis. A species of *Rhodocrinites* with a flat, bowl-shaped calyx and an invaginated base; ornamentation consisting of stellate ridges, which cover each of the calyx plates.

Description. The theca is flattened, with a shallow, bowl-shaped calyx that has an invaginated base; the anal tube is small (Fig. 3d). The best preserved calyx from the Seboomook Formation (Figs. 3a-b) measures 23 mm at its greatest width and is 6 mm high (Table 1). The holotype (Fig. 3e) is between 19.5 and 23 mm in width and is approximately 4 mm high. The tegmen plates are unknown; only an outline of the theca is preserved (Fig. 3d). All of the major calyx plates are on the flattened base of the calyx. The primaxil brachial plate of each ray is at the outer margin of the flattened calyx. There are either two or three secundibrachials on the top of the calyx, which give rise to the free arms.

TABLE 1. MEASUREMENTS (IN MM) OF *RHODOCRINITES NORTONI* (GOLDRING).

Specimen	Width of theca	Height of theca or calyx
UMPI 10500	19.5-23	4
WVU 1101	23	6
WVU 1102	20	6

The calyx plate arrangement (Fig. 4) essentially matches that of *Rhodocrinites* illustrated in the Treatise of Invertebrate Paleontology (Ubaghs, 1978, p. T422). Five infrabasals, forming a pentagon, are visible within a circlet of five basals (Figs. 3a and 4). The radials are separated by well-developed interbrachial plates. Prominent radial ridges cover the calyx plates forming an interlocking stellate pattern (Figs. 3a and 3e). A robust radial ridge extends upward from each radial, across the primibrachials, forking at the primaxil, and extending onto the fixed secundibrachials (Fig. 3e).

The stem attachment scar does not completely cover the infrabasal plates (Figs. 3a and 4). The stem is circular in outline as indicated by dissociated pluricolumnals and the stem fragment attached to the holotype (Fig. 3e).

Each fixed arm branches once to form two sets of free arms (Figs. 3a, c, and e); free arms divide further to yield three or four arms (Figs. 3c and e). Thus, there are six or seven, but not eight, arms in each ray. Each set of arms branches endotomously with new branches added on the inside of the arms. The arms are biserial, with long pinnules (10 mm) above the secundibrachials. Partial arms from several individuals were found in the Seboomook Formation, but no complete crowns were recovered. Thus it was not possible to determine which rays produced the sets of four arms. No more than three rays, and their attached arms, were preserved on any specimen from the Seboomook Formation. Goldring (1933) indicated that there were 30 arms on *R. nortoni*, but the holotype actually has only 19 arms preserved (Fig. 3e). Thus, the total number of arms

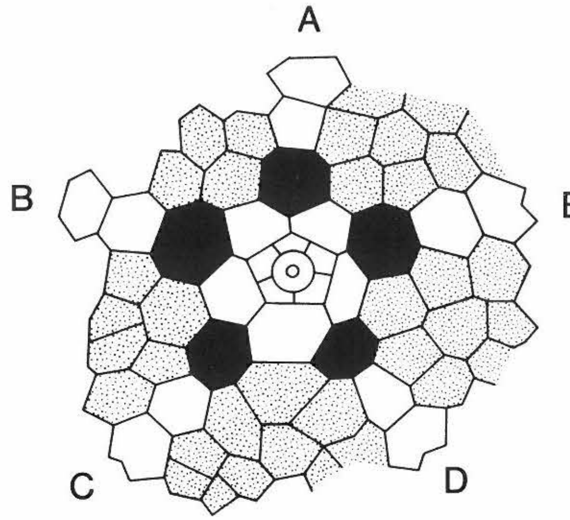

Figure 4. Camera lucida drawing of actual (not idealized) calyx plate arrangement of *Rhodocrinites nortoni* based largely on the holotype (Fig. 3e). Arrangement of infrabasal plates determined from specimen WVU 1101 (Figs. 3a and b). Radials shown in black; interbrachials stippled. Individual rays identified by letter.

could have ranged from 30-35.

Remarks. The above description is based on several partial specimens of *R. nortoni* found in the Seboomook Formation, plus the holotype, which was obtained for comparison with the Seboomook Formation material.

All the specimens from the Seboomook Formation, plus the holotype have a flattened theca. Because the calyx plates are all tightly interlocking, rather than ruptured and dissociated, the original calyx was probably not globose with vertical sides as is typical of *Rhodocrinites* (Ubaghs, 1978, p. T420). However, compression associated with sediment compaction has probably exaggerated the flatness of the theca and calyx.

The holotype shows evidence of predation (Fig. 3e). Three of the arms on the E-ray have been partially regenerated as shown by the more slender segments on the distal portions of the arms. Additionally, the brachial plates are disrupted where the arms were nipped, and one arm regenerated two branches rather than a single branch.

The Early Devonian age of *R. nortoni* apparently makes it the oldest representative of the genus. Ubaghs (1978, p. T421) reports that the genus is found only in the Lower Carboniferous. However, *R. insculptus* (Goldring) 1935, *R. quinquelobus* (Schultze) 1866, and *R. ornatus* Dubatolova, 1964 have been reported from the Middle Devonian of New York, Belgium, and the Soviet Union, respectively (Bassler and Moodey, 1943; Webster, 1973). We have not determined whether these three species have been correctly assigned to the genus R. *Rhodocrinites*.

Repository. The specimens from the Seboomook Formation are deposited in the paleontological collections of the Department of Geology and Geography at West Virginia University. The two illustrated specimens are WVU 1101 and 1102. Many additional fragmentary specimens are grouped together under WVU 1110. The holotype is in the paleontology collections at the University of Maine at Presque Isle (UMPI), No. 10500 (formerly No. 4020, Portland Society of Natural History).

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Plant Paleontology in the State of Maine – A Review

Andrew E. Kasper, Jr.
Department of Biological Sciences
Rutgers University
Newark, New Jersey 07102

Patricia G. Gensel
Department of Biology
University of North Carolina
Chapel Hill, North Carolina 27599-3280

William H. Forbes
Division of Mathematics and Science
University of Maine, Presque Isle
Presque Isle, Maine 04769

Henry N. Andrews, Jr.
Professor Emeritus
University of Connecticut
Storrs, Connecticut 06268

ABSTRACT

The history of paleobotany in Maine is presented along with reviews of major Devonian floras and their impact on our understanding of early vascular land-plant evolution. The Late Silurian and Devonian represent a time span in which plants emerged onto the land surface, diversified, and evolved major structural features. Data for these events are preserved in fossil floras present in strata ranging from Lower to Upper Devonian in north-central and easternmost coastal Maine. Some critical floras reviewed are those of the Lower Devonian Trout Valley Formation, Middle Devonian Mapleton Sandstone, and Upper Devonian Perry Formation. An updated assessment of geological age, species composition, and evolutionary significance is given for each plant-bearing formation in Maine.

INTRODUCTION

Continental strata in the state of Maine are relatively sparse, but their significance in the study of early land plant evolution is considerable. Five formations distributed throughout the state and representing a time interval from Late Silurian to Late Devonian have yielded numerous plant megafossils (Fig. 1). These fossils have provided an important database for our understanding of the emergence and subsequent dominance of vegetation on the land surface. The quantity and quality of paleobotanical information obtained from the Maine paleofloras are somewhat out of proportion to the limited number, size, and exposure of these continental formations. Furthermore, the prognosis is excellent. Known fossil localities are expected to be productive for years to come and the discovery of new localities is anticipated with continued field work. Maine's contribution to paleobotany, in particular the problem of early vascular land plant evolution, has been and will be outstanding.

Late Silurian and Early Devonian events are critical to our understanding of the transition of plants from aquatic to terrestrial environments. The scenario as often painted is that aquatic plants, presumably algal in nature and estuarine in habitat, became adapted to dry land, perhaps as a result of their competition for space (i.e., for light as an energy source and/or for mineral nutrients necessary for growth). The change may have been gradual, involving adaptation first to periodic dessication

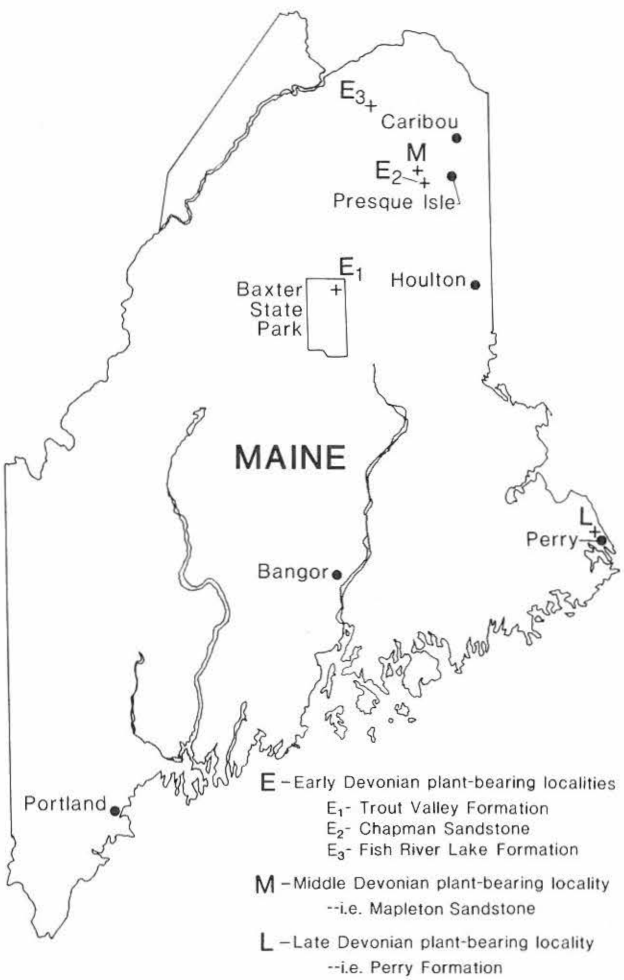

Figure 1. Map of Maine showing Devonian plant-bearing localities.

caused by tidal fluctuations and then to permanent dessication in supratidal zones. A host of morphological, anatomical, and physiological adaptations would have to be orchestrated for success. For example, these plants would also have to evolve a means of coping with the greater amounts of UV light present outside of water and a way of acquiring water internally. Knoll et al. (1986) present current theories about such a transition and review the Silurian fossil evidence relating to it.

Early Devonian plant fossils from Maine and other regions continue the story of this transition. The rapid evolution of plants newly adapted to the land surface is shown by the extensive diversification of vascular plant taxa or plant structures during the Early Devonian (Gensel and Andrews, 1984; Edwards and Fanning, 1985; Knoll et al., 1986; Gensel, 1986). The Maine plant fossils, plus others from nearly contemporaneous strata in New Brunswick and the Gaspé Peninsula, provide a detailed view of such evolutionary change for the North American portion of the Old Red Sandstone Continent.

Middle Devonian events continue the theme of morphological diversification in vascular land plants with the appearance of several new plant lineages and innovations in plant structure. Differentiation of the plant stem into various specialized organs (the incipient megaphyllous leaf and the root), the development of secondary stem-thickening tissues, and the arborescent habit are prominent features present during this time. Historically, Middle Devonian strata in Maine have yielded less information than Lower and Upper Devonian strata. Part of this is due to the limited surface exposure of Middle Devonian beds, but some of it may be the result of our emphasis on the highly productive Lower Devonian localities. The brief studies already published on Maine's Middle Devonian flora indicate potential for further contributions. Recent finds of permineralized (petrified) remains will contribute further information about the structural evolution occurring within this epoch.

Late Devonian time marks the culmination of the rapid diversification of vascular land plants. Leafy arborescent species of great stature are now present. Of extreme importance are the innovations in reproductive biology. Several taxa display heterospory—production of two kinds of spores resulting in unisexual gametophyte plants. This is a major step toward the evolution of the seed. In rare instances seeds have been reported from the Late Devonian in other geographical regions. From a historical point of view, many genera and species of plants were first recognized and named from the Upper Devonian strata of Maine. Even though Upper Devonian localities in Maine are few and surface exposure is minimal, new collections from old sites and restudy of curated specimens will further clarify our understanding of the biology and systematics of these historical taxa.

LINEAGES OF EARLY LAND PLANTS

It seems appropriate, here, to briefly introduce some concepts concerning the structure and classification of early vascular land plants. The majority of these fossils represent fragmentary remains of stems, reproductive organs and, in some cases, roots and leaves of the sporophyte phase (spore-producing plant) of the life cycle. The smaller, more delicate gametophyte phase (egg and sperm-producing plant) of the life cycle was either not preserved or was so different in form from modern counterparts that gametophyte plants have not as yet been recognized as such.

Early and even some Middle Devonian plants are comparatively small and simple. For many years they were collectively referred to as "psilophytes." However, as the data accumulated, recent workers have recognized that several evolutionary lineages were included in this catchall group. This recognition culminated in the classification scheme proposed by Banks (1968) in which several new early land plant groups were distinguished. The present review examines many different types of plants so it is appropriate at this point to summarize the characteristics of the major lineages.

(1) The RHYNIOPHYTES are small plants (one to several decimeters tall) consisting of leafless, dichotomizing stems, some of which are terminated by a sporangium which has no

special means of dehiscence (splitting open). The vascular strand consists of a few tracheids making up a solid cylinder of xylem with first-formed elements in the middle (a centrarch protostele).

(2) The TRIMEROPHYTES are larger (a decimeter to one or more meters tall) and more complexly branched plants (with main axes and lateral branches) bearing large clusters of sporangia at the tips of some side branches. Their vascular strand is more massive, but still a centrarch protostele. Sporangia dehisce longitudinally. Trimerophytes are regarded as being descendants of the rhyniophytes and precursors to several later-appearing plant groups, such as progymnosperms, ferns, etc.

(3) The ZOSTEROPHYLLOPHYTES are plants several decimeters tall with both equal and unequal branching, an exarch protostele (first-formed xylem occurs at margins of vascular strand), and large, laterally borne, globose or reniform sporangia. The latter open along their outer margin, often by means of specialized cells.

(4) The LYCOPODS are herbaceous or woody plants of predominantly rhizomatous habit, varying greatly in size (from a few decimeters to many meters) and in extent of branching. They possess microphyllous leaves and exarch protosteles. In most species the globose or reniform sporangia are borne on the upper surface or in the axils of the leaves.

(5) The PROGYMNOSPERMS are plants of shrub or tree size with much-divided branch systems, some of which terminate in laminar leaves. The stems have well-developed secondary vascular tissue as is found in modern gymnosperms. Their sporangia are borne on the tips of branches or on the surfaces of leaves. These plants show a more "modern" growth habit and structure and are more advanced in reproduction since some are heterosporous.

(6) The CLADOXYLS and IRIDOPTERIDS are two separate lineages of plants whose affinities are less well understood. They apparently are derived from the trimerophyte line and one or both may represent precursor(s) to the horsetails. Cladoxyls are distinguished anatomically by their much-dissected vascular strand and iridopterids by their deeply lobed vascular strand and distinctive lateral branch trace formation. External morphology of many taxa is still unknown, as most species are based on short lengths of permineralized stems. Where known, branching in some cladoxyls is dichotomous and digitate. Finely divided ultimate appendages are borne on some of the branches in dense spirals or whorls and some terminate in oval-elongate sporangia.

HISTORY OF PALEOBOTANY IN MAINE

Significant paleobotanical studies in Maine began in the 1860's with the work of the Canadian geologist, Sir John William Dawson (Table 1); these were followed much later by a contribution from David White in 1905, and then little was done until the mid 1960's. From that time on, numerous studies have appeared. In view of the pioneering work of Dawson and White we have included brief, general biographical sketches along with summaries of their paleobotanical contributions.

TABLE 1. HISTORY OF PALEOBOTANY IN MAINE

Date	Author	Occurrence
1861	Dawson	Plants from Perry Formation
1862	Dawson	Plants from Perry Formation
1863	Dawson	Plants from Perry Formation
1900	Williams	*P. princeps* var. *ornatum;* Mapleton Sandstone
1905	Smith and White	Plants from Perry Formation
		Psilophyton from Fish River Lake Formation
1916	Williams and Breger	*Psilophyton princeps* from Chapman Sandstone
1941	Kräusel and Weyland	Plants from Perry Formation
1943	White	*Psilophyton princeps* from Mapleton Sandstone
1962	Dorf and Rankin	Plants from Trout Valley Formation
1964	Schopf	Plants from Mapleton Sandstone
		Pachytheca from Chapman Sandstone
1964	Schopf *in* Boucot	Plants from Chapman Sandstone
	et al.	Plants from Mapleton Sandstone
1965	Pettitt	Heterospory in *Barinophyton richardsonii*
1966	Schopf et al.	*Eohostimella heathana;* Frenchville Formation
1968	Andrews et al.	*Psilophyton forbesii;* Trout Valley Formation
1969	Gensel et al.	*Kaulangiophyton* from Trout Valley Formation
1970	Andrews and Kasper	Review of Trout Valley Formation flora
1972	Andrews et al.	Spores from Trout Valley Formation
1972	Kasper and Andrews	*Pertica* from Trout Valley Formation
1974	Kasper et al.	*Psilophyton* spp. from Trout Valley Formation
		Dawsonites from Fish River Lake Formation
1977	Andrews et al.	Review of Trout Valley Fm. micro- & macroflora
1979	Kasper and Forbes	*Leclercqia* from Trout Valley Formation

The first important reports describing plant fossils from the state are those of Dawson (1861, 1862, 1863; Fig. 2) who studied the flora of the Perry Formation in southeastern Maine. Many of these fossils were described as new species since so little was known about Devonian plants. The Perry plants are not well preserved, being quite fragmentary, but Dawson's reports were a significant beginning.

Dawson's studies also provided the first substantial paleontological evidence for the Devonian age of the Perry Formation, until then a controversial issue in Maine geology. Jackson (1837), the state geologist for both Massachusetts and Maine, at first regarded the Perry Formation as equivalent to the Carboniferous strata of New Brunswick and Nova Scotia. Later he suggested that it was the same age as the red beds in Connecticut and New Jersey and might possibly be Silurian (Jackson, 1851). Using plant fossils from the Perry Formation, Rogers (1859) correlated the formation with the "Kiltorcan" of Ireland which he considered to be Lower Carboniferous—all of which summed up to a rather confused situation! In 1861, C. H. Hitchcock became the State Geologist of Maine and asked Dawson to establish the age of the beds of the Perry Formation. Dawson's study, presented in Hitchcock's (1861) report on the geology of the Perry Basin, assigns a Devonian age to the formation which has held up to this day.

J. W. Dawson (1820-1899) was born in Pictou, Nova Scotia, and started making collections of fossil plants when he was twelve. He studied at Pictou College and then went to Edinburgh where he met such eminent men of the times as Forbes,

Figure 2. Sir John William Dawson in the 1860's; from the W. C. Williamson photo collection courtesy of the Burndy Library, Norwalk, Connecticut.

Balfour, and Jamieson. Upon returning to Canada in 1842, he met Charles Lyell who was on one of his American tours, and the two spent considerable time together. In days when travel conditions were quite different from today, Dawson explored the coastal areas of the Gaspé Peninsula, New Brunswick and Maine. Of particular interest are his studies of the land plants of the Early Devonian. However, the importance of these studies to understanding early evolution in plants was not appreciated until some decades later. Probably his best known work is his monumental *Acadian Geology* (Dawson, 1891). Dawson's paleobotanical studies are in part summed up in *The Geological History of Plants* (Dawson, 1888). He was one of Canada's great educators, serving as Superintendent of Education for Nova Scotia and as Principal of McGill University in its early years. Some interesting aspects of his life as a naturalist in those years are to be found in his autobiography, *Fifty Years of Work in Canada* (Dawson, 1901).

The next work of importance in paleobotany in Maine was a monograph by Smith and White (1905) on *The Geology of the Perry Basin in Southeastern Maine*. The report (the paleobotanical part of which was by David White) seems to have been spurred by persistent rumors and economic expectations that coal in commercial quantities might be available in the Perry Formation. The significance of the monograph lies in its meticulous treatment of the flora. The plants are carefully described and illustrated, and a thorough synonymy and discussion are given. White lists 29 taxa of plants, 16 of these being Dawson's.

David White's (1862-1935; Fig. 3) first important work in paleobotany and stratigraphy involved collecting and studying the plants from Gay Head on Martha's Vineyard and his demonstration that they were of Cretaceous age. He is probably best known for his extensive work with late Paleozoic plants which included monographic treatments of the Coal Measure plants of Missouri and the Permian floras of Brazil and the Grand Canyon. He served as Chief Geologist of the U.S. Geological Survey, as Home Secretary of the National Academy of Sciences, and Chairman of the National Research Council's Committee on Paleobotany. E. W. Berry (1935, p. 391) wrote of David White: "No geologist of his time had a wider influence on the scientific life of the nation, or took a more active part in that of its capital."

The only major contribution to Maine paleobotany during the hiatus between Smith and White (1905) and the revival of the 1960's is the work of two German paleobotanists, Kräusel and Weyland (1941). They present the first photographic illustrations of the Perry flora and, more significantly, their approach is thoroughly modern in aspect. Besides careful specimen description, they are concerned with the biology of whole organisms, i.e., their concept of species allows for some variation in morphology between organisms and, hence, between specimens. This philosophy is reflected in the fact that they reduce Smith and White's 29 taxa to 9.

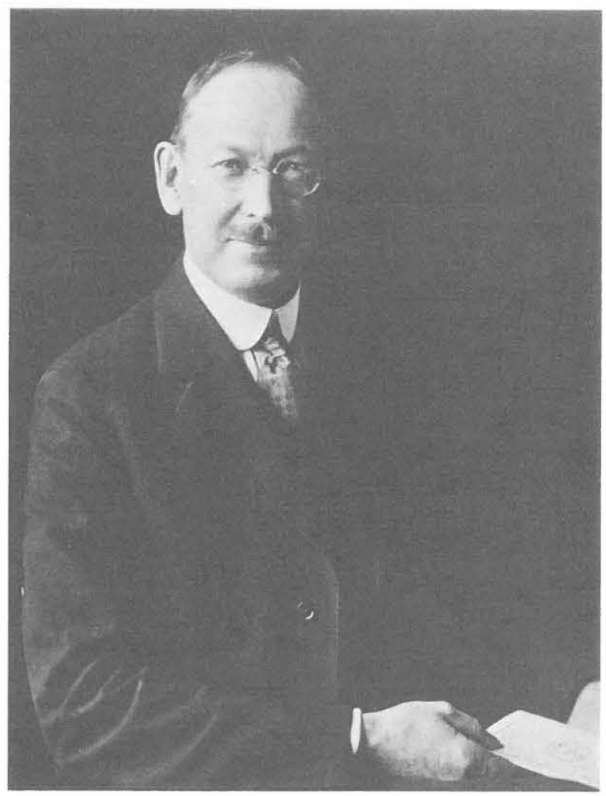

Figure 3. David White in the early 1900's; photo courtesy of the U.S. Geological Survey, portrait # 68.

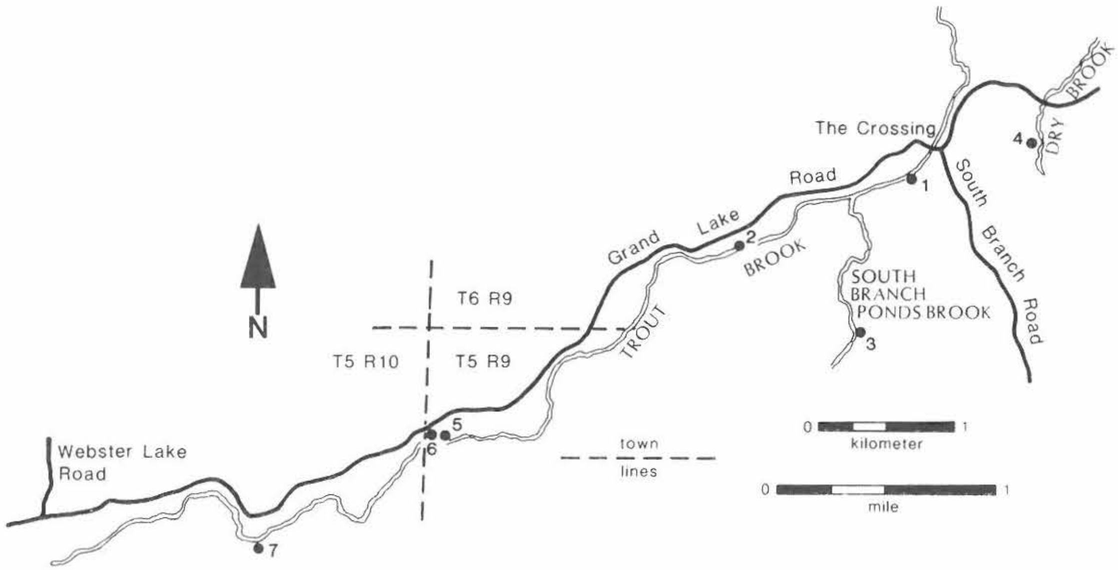

Figure 4. Major plant-bearing outcrops of the Lower Devonian Trout Valley Formation in Baxter State Park, Maine.

It is important to realize that the paleobotanical studies in Maine summarized thus far have all dealt with the Late Devonian flora from the Perry Formation. The paleobotanical revival in the 1960's centered in new areas (northern Maine) and at lower stratigraphic levels (Lower and Middle Devonian). The historical development of these studies will be incorporated into the various discussions of the floras themselves, starting with that of the Trout Valley Formation in Baxter State Park.

EARLY DEVONIAN FLORA IN MAINE

Plants of Early Devonian age have been found in northern Maine in two different formations: the Trout Valley Formation and the Fish River Lake Formation. These two floras will be treated sequentially below.

Flora of the Trout Valley Formation

Of the various localities in Maine where fossil plants have been found, Trout Brook and its tributaries in Baxter State Park have produced the most significant information. Therefore, it is especially appropriate to record in some detail the history of the work that has been undertaken there and the plants that have been discovered. The area, shown in Figure 4, includes the banks of Trout Brook from "The Crossing" upstream for about 6 kilometers and localities along Dry Brook and South Branch Ponds Brook.

The first published report of fossil plants found in this area is that of Erling Dorf and Douglas Rankin (1962) (Table 2). The plants that they described were actually discovered between 1955 and 1957 by Rankin during explorations sponsored by the Maine Geological Survey and Harvard University. Since the discovery of plants in this previously undescribed formation is so important to Maine paleobotany, we quote the following from their account (p. 999):

> All the plant fossils occur in a heretofore unnamed formation in the Traveler Mountain and Telos Lake quadrangles, north-central Maine (Rankin, 1958 and 1961). This formation, here named the Trout Valley formation, is composed predominantly of clastic rocks and occupies the valley of Trout Brook between the west shoulder of Trout Brook Mountain and the north shoulder of Burnt Mountain. . . .

With our more recent contributions to the Trout Valley flora, it seems appropriate to record something of the history of the developments that followed the Dorf-Rankin introduction. In the early 1960's Ely Mencher, a professor at M.I.T., was engaged in stratigraphic studies in northern Maine. Mencher was aided by William Forbes of Washburn who had an extensive field knowledge of northern Maine geology. With the hope that the fossil plants they were finding would be helpful with age determinations, Mencher invited James M. Schopf of the U.S. Geological Survey to participate in the field work. Schopf was one of the leading authorities of his time on Paleozoic plants and he, in turn, asked Henry Andrews (University of Connecticut) to join the group in the summer of 1965.

The pathways to success in paleontology are certainly varied and often unpredictable. Much of the plant material that was being found was quite fragmentary and not too encouraging. It was no secret that the forested areas of north-central Maine were not generally regarded as a likely place to look for fossil plants. The "breakthrough" came when Schopf and Andrews were invited to examine some large specimens at Forbes' home. The specimens contained some spectacular fossil plants, described later as *Psilophyton forbesii* (Andrews et al., 1968). These had been collected by Forbes at the South Branch Ponds

TABLE 2. FLORAL COMPOSITION OF SILURIAN/DEVONIAN PLANT-BEARING FORMATIONS IN MAINE

Age	Formation	Taxon	Selected references
UD	Perry Formation	*Archaeopteris jacksonii*	Posnick, 1982
		Barinophyton richardsonii	Posnick et al., 1983
		Leptophloeum rhombicum	Posnick, 1982
		Platyphyllum brownianum	"
		cf. *Rhacophyton incertum*	"
		Barinostrobus spicatus	"
		cf. *Cyclostigma*	"
		cf. *Tetraxylopteris*	"
MD	Mapleton Sandstone	cf. *Sawdonia ornata*	Schopf, 1964
		Barrandeina(?) *aroostookensis*	"
		Calamophyton forbesii	"
		Hostinella sp.	"
		Aphyllopteris sp.	"
		cf. Stolbergia	present report
		cf. *Cladoxylon*	"
		cf. *Calamophyton*	"
		cf. *Schizopodium*	"
		cf. *Rhacophyton*	"
MD	Chapman Sandstone	*Pachytheca* sp.	Schopf, 1964
LD	Trout Valley Formation	*Sawdonia ornata*	Dorf and Rankin, 1962
		Psilophyton forbesii	Andrews et al., 1968
		Kaulangiophyton akantha	Gensel et al., 1969
		Taeniocrada sp.	Andrews and Kasper, 1970
		Prototaxites sp.	"
		Pertica quadrifaria	Kasper and Andrews, 1972
		Psilophyton dapsile	Kasper et al., 1974
		Psilophyton microspinosum	"
		Psilophyton princeps	"
		Drepanophycus sp.	Andrews et al., 1977
		Thursophyton sp.	"
		Leclercqia sp.	Kasper and Forbes, 1979
S/LD	Fish River Lake Formation	cf. *Dawsonites arcuatus*	Kasper et al., 1974
		cf. *Psilophyton dapsile*	Kasper and Forbes, 1983
S	Frenchville Formation	*Eohostimella heathana*	Schopf et al., 1966 Strother and Lenk, 1983

Brook locality shown on the map (Fig. 4, # 3). It was immediately evident that here was the kind of lead that paleontologists long for. Indeed, it initiated several summers of extensive field work, chiefly by Forbes, Andrews, Kasper, and Gensel; the latter two were graduate students with Andrews at the time.

Most of our exploration and collecting has been done in the rock exposures along the banks of Trout Brook, and we introduce this with the following background information. Dorf and Rankin (1962, p. 1001) state that:

> The maximum exposed thickness of the Trout Valley formation is about 1500 feet, and the formation crops out over an area of about 1.5 by 8 miles.

The Trout Valley Formation unconformably overlies the Traveler Rhyolite which, in turn, rests conformably on the Matagamon Sandstone of Becraft-Oriskany (= Siegenian) age (Rankin, 1965). An absolute date for the Traveler Rhyolite has been determined by Bottino et al. (1966) as 360 ± 10 Ma. This allows a range of 350 to 370 Ma for the rhyolite, which places the oldest possible date for the Trout Valley Formation between earliest Middle Devonian and latest Late Devonian using the geologic time scale of Harland et al. (1964). This is not a very precise determination for the age of the formation. Based on the nature of the flora itself, it was our early opinion that the strata were of late Early Devonian age, but as more information has accrued, it is possible that the formation may be earliest Middle Devonian in age. Further discussion of this will be given following a description of the plants.

The Trout Valley Formation is somewhat enigmatic. In their initial account Dorf and Rankin (1962, p. 1001) note that: "The formation is remarkably undeformed. . . ." How do we account for the fact that this rather small patch of sedimentary rocks has escaped distortion, if not complete destruction, over the

hundreds of millions of years since the original deposition?

Whatever the reasons may be for the preservation of the Trout Valley Formation, it has proven to be a rich source of plant remains. Our information comes from only a very small part of the total area. About 200 meters upstream from The Crossing lies the most extensive outcrop yet found (Fig. 4, # 1); it has yielded large quantities of fossil plants which will be described below. Other localities along the banks of Trout Brook consist of rather small lenses where, in some cases, fossils may be found one year and are gone the next because of erosion during the spring. At the westernmost (upstream) point, considerable fossiliferous rock has been found in the stream bed and has been traced to an outcrop along the south bank (Fig. 4, # 7). Some of the fossil plants are sufficiently well-preserved to provide us with a clear understanding of their appearance in life. Others are known from more fragmentary remains, or else do not reveal the reproductive organs needed for precise identification and classification. It is tantalizing to speculate on what fossils lie hidden in the surrounding heavily wooded areas.

Below will follow brief descriptions of the better known plants of the formation, dealing with The Crossing locality in some detail. As shown in Figure 5, the beds here are slightly tilted; the entire exposure contains fossil plant material in varying amounts. One of the better preserved plants, confined to a narrow band several cm thick, is *Psilophyton dapsile* (Fig. 6). This small, simple plant is characteristic, generally, of the earliest vascular land plants and of more primitive trimerophytes. It is a decimeter or two tall, consisting of a dichotomous (Y-shaped) branch system (Fig. 7a), many of the branchlets being terminated by pairs of sporangia about 2 mm long and 0.5-0.9 mm wide (Fig. 7b). It was found in abundance and in association with no other plant remains, suggesting that it existed in life as a pure stand.

We have found several distinct species referable to *Psilophyton*. Therefore, it is appropriate to explain the importance of this genus and to point out some of the features of a primitive land plant. The genus has a long nomenclatural history that was initiated in 1859 by the Canadian geologist, Sir J. W. Dawson, with his description of material found along the north shore of Gaspé Bay named *Psilophyton princeps*. Specimens assigned to *Psilophyton* reveal considerable information about the characteristics of the earliest vascular land plants and about the stages in the developing complexity of plants in the Early Devonian. Several species of this genus have been found in the Trout Valley Formation. The first plant that we studied in the flora, *Psilophyton forbesii*, was found at the South Branch Ponds Brook locality (Fig. 4, # 3). *P. forbesii* was a much larger plant than *P. dapsile*. The strongly developed central stem of the former attained a diameter of nearly 1 cm and bore dichotomous side branches, some of which terminated in pairs of sporangia 3.5 to 5.0 mm long. It is likely that the plant attained a height of a meter of more. Specimens were also found along Trout Brook at locality # 5 (Fig. 4).

A third species, *Psilophyton microspinosum* (Fig. 8), is intermediate in size between the other two and is distinguished by the presence of sparsely distributed slender spines (emergences) about 2 mm long. The plants occurred in a small lens about 1 meter above water level, there being perhaps a cubic meter of rock containing well preserved specimens.

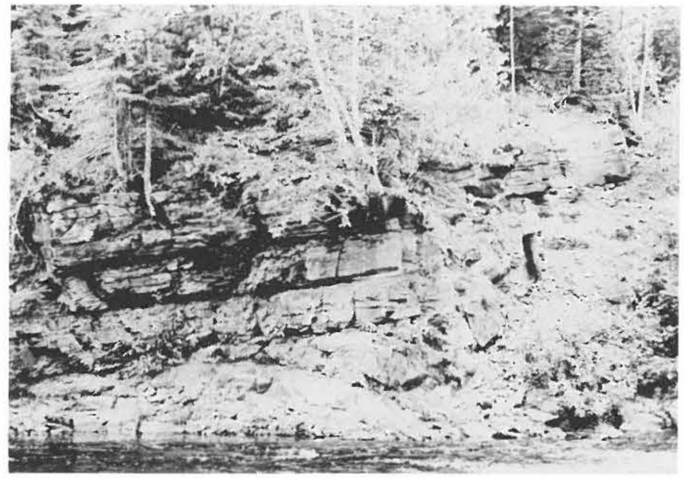

Figure 5. Strata at "The Crossing" locality, Trout Valley Formation.

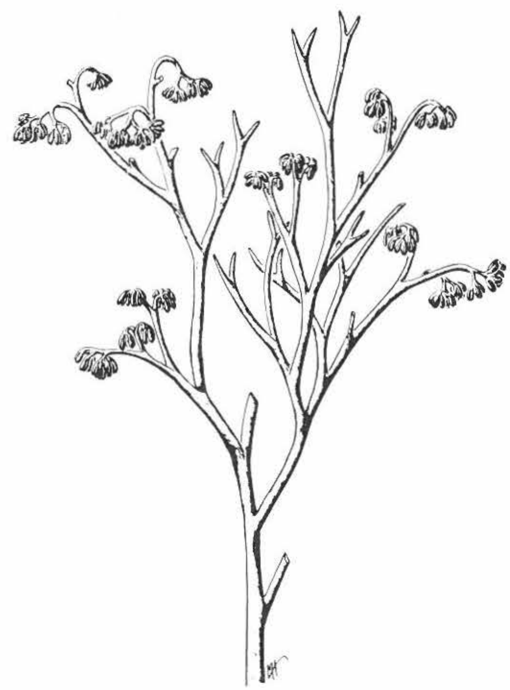

Figure 6. Restoration of the Early Devonian trimerophyte, *Psilophyton dapsile*. Scale bar equals 1 cm.

Specimens of a fourth species, *Psilophyton princeps* (Dawson) Hueber, were found in an outcrop about 7-8 meters north of where *P. forbesii* was first discovered. The stems of *P. princeps* are distinguished by their abundant, peg-like emergences which are clearly shown in the photo (Fig. 7c). *P. princeps* was the first Early Devonian species described by Dawson in 1859. This plant has only rarely been reported outside of its original

Figure 7. Some plants typical of the Lower Devonian Trout Valley flora. X 2.7. (a) Sterile branchlets of the trimerophyte *Psilophyton dapsile*. (b) Fertile branchlets of *P. dapsile*. X 2.8. (c) Two axes (stems) of the Early Devonian trimerophyte *Psilophyton princeps*. Note peg-like emergences. X 1.4. (d) Dichotomizing smooth axes occurring abundantly in several localities. No reproductive structures are attached so affinities are unknown. X 0.5.

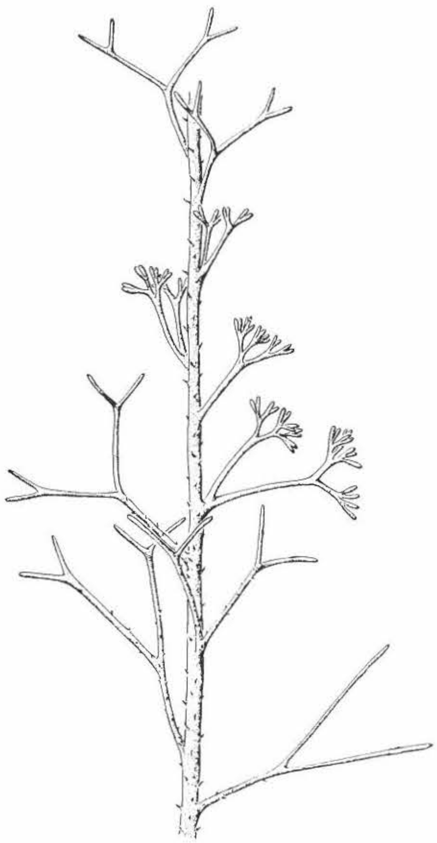

Figure 8. Restoration of the Early Devonian trimerophyte *Psilophyton microspinosum.* Axes are covered with tiny spines.

locality in the Gaspé Peninsula. A few specimens with sporangia were recovered at the Maine locality (Kasper et al., 1974). The material from the Trout Valley Formation resembles that from the type locality (Gaspé Bay), differing only slightly in the form and size of the emergences.

It is appropriate to reflect at this point on the significance of the information afforded by these plants. Numerous species of *Psilophyton* have been described from scattered Lower and Middle Devonian localities worldwide. Its significance as a genus may be problematical, but the species assigned to it seem to offer very significant data on the evolution of the more primitive early vascular land plants. We have not found petrified material in the Trout Valley flora, but *Psilophyton* specimens from other localities reveal a slender, cylindrical core of xylem (water-conducting tissue) in the axes. The different species show an increase in size, the development of strong central stems, a diversity in the kinds of emergences, and considerable variation in the abundance and size of the sporangia.

Several other plants have been found at The Crossing locality which supply varying amounts of information. Representative specimens of what is the most abundant component of the flora are shown in Figure 7d. The main axes of the plant are 3 mm in diameter and these bear dichotomously forking branches. The plants probably attained a meter in height and are found as a "pure stand" in great abundance through several decimeters of the cliff. It is clear that it was a dominant element of the flora, probably for a considerable period of time. The material is thought to represent a *Psilophyton* species, but we have never encountered the fertile parts, the sporangia, which are critical for identification.

In contrast to large quantities of material, one of the most interesting plants that we have found is known from only a few specimens, but it affords a great deal of information. *Kaulangiophyton akantha* (Fig. 9) is probably closely related to the lycopods. It is known from axes that branch to form H- or K-type patterns, and these bear sparsely distributed, stout spines up to 2 mm long which do not seem to be oriented in any regular pattern. Interspersed among these spines, on some of the branches, are large sporangia (6-8 mm in their greatest diameter) borne terminally on short stalks (Fig. 10a). Spores have not been isolated from these structures, but comparisons with similar organs in other fossil plants leave no reasonable doubt as to their identity. The sporangia probably contained very large numbers of spores – a characteristic of many of the early vascular land plants. At several places along Trout Brook we have found spiny stems up to 2 cm in diameter which suggest either larger plants of *Kaulangiophyton* or a completely different plant.

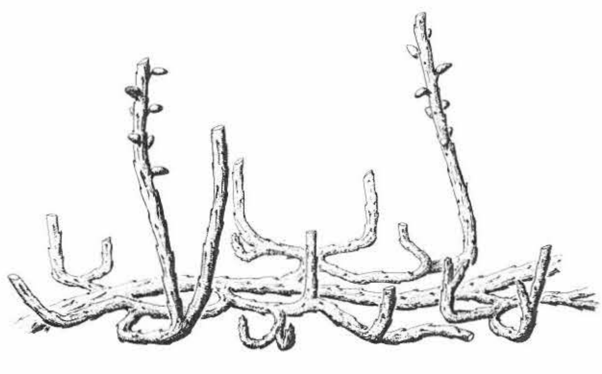

Figure 9. Restoration of the pre-lycopod *Kaulangiophyton akantha.*

A particularly vexing plant that we have found at The Crossing is shown in Figure 10b. Numerous specimens were recovered from a band of rock about 10 cm thick; the axes are 5-6 mm in diameter and are densely covered with delicate, needle-like emergences 2 mm long. It is very different from anything else that we have found but, so far, we have not encountered fertile specimens. Fossil plants of this general nature have been reported from several other Early Devonian horizons and are assigned to the "form genus" (the affinities of which are unknown), *Thursophyton*. This generic name means the "Thurso plant" and refers to specimens found many years ago in Middle Devonian rocks at Thurso in northern Scotland.

A kilometer upstream from The Crossing, we (Kasper and Andrews) encountered plant remains in a low, nearly flat ledge

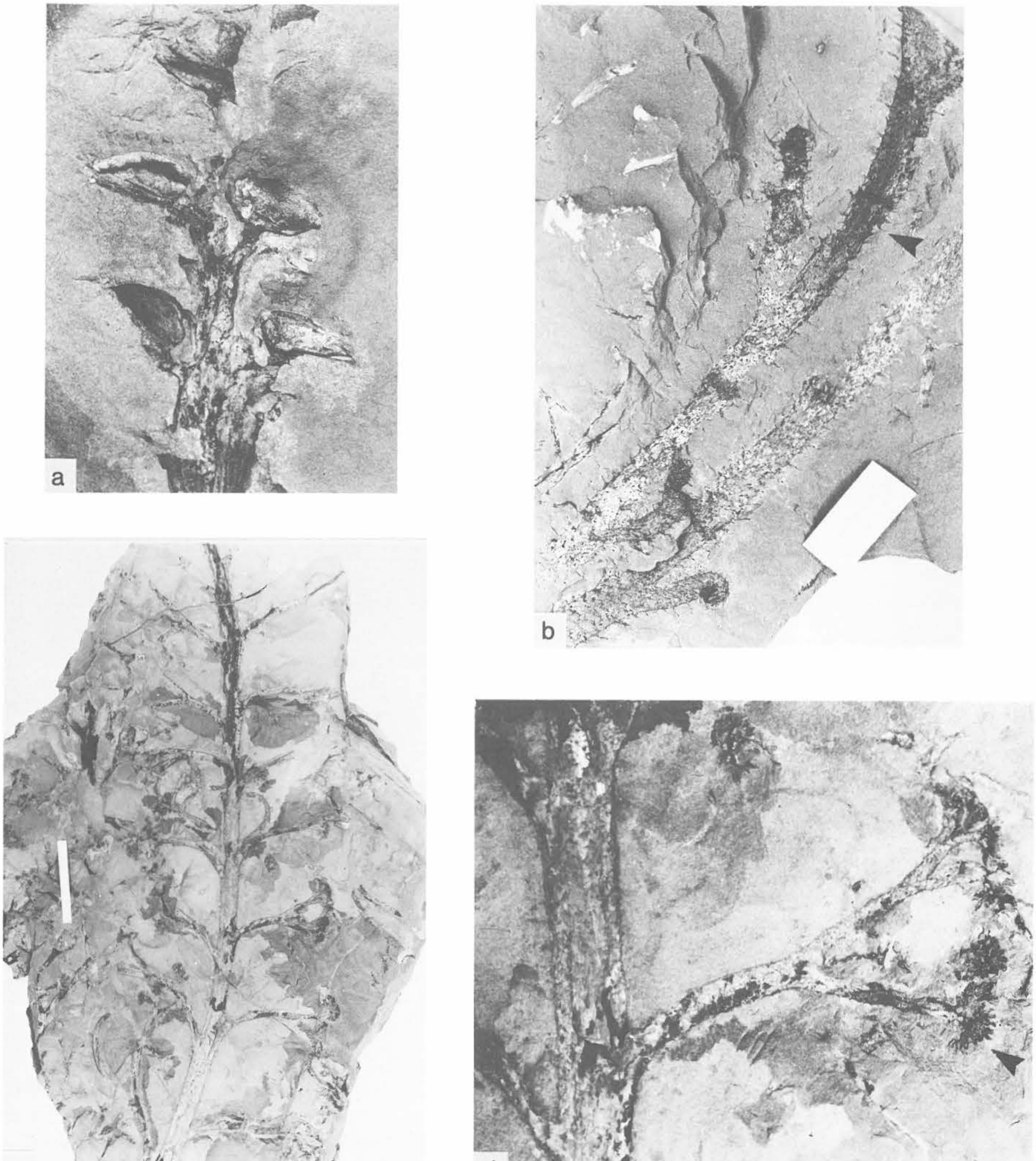

Figure 10. Plants of the Trout Valley flora. (a) Distal region of fertile stem of *Kaulangiophyton akantha*; note large, stalked sporangia. X 1.3. (b) Spiny axes similar to the genus *Thursophyton*; arrow indicates spine. X 1. (c) Large central axis of the trimerophyte *Pertica quadrifaria* with several lateral branches terminated in sporangia. X 0.2. (d) Detail of a fertile lateral branch of *P. quadrifaria*; arrow indicates sporangia. X 0.8.

along the north side of the brook (Fig. 4, # 2)—an outcrop quite different from the vertical cliff at The Crossing locality. Both the plant fossils and the lithology of the sedimentary rocks were quite different from anything that we had found previously.

The plant was given the name *Pertica quadrifaria* (Figs. 10c, 10d, 11). The generic name is from the Latin meaning a "long pole or rod" and refers to the stout upright stems. The specific name refers to the four-rowed arrangement of the primary branches. The latter were three-dimensional in their branching pattern, some being sterile and presumably photosynthetic, while others terminated in very dense clusters of sporangia from which spores were isolated (Figs. 10c, 10d). *Pertica* represents a more complex and advanced trimerophyte than *Psilophyton*, showing some early stages in the differentiation of branch systems leading to the origin of megaphyllous (broad-leaf type) leaves.

Pertica also presents a good example of the mode of preservation of many Devonian plants. When the rock is first split open, only a small part of the plant may be revealed. It is evident that at the time of deposition the plant or plant parts were buried in the enclosing sediment without significant distortion and thus were preserved in three dimensions.

In 1985 the Maine Legislature selected *Pertica quadrifaria* as the state fossil.

One of the more recently discovered plants in the Trout Valley Formation is *Leclercqia complexa* found in small quantities about 6 kilometers upstream from The Crossing (Fig. 4, # 7). It is of special interest relative to the age of the formation. *Leclercqia*, named for the distinguished Belgian paleobotanist, Suzanne Leclercq, has been reported previously from several widely separated Middle Devonian localities including New York State, Belgium, and Queensland (Australia). It also occurs in late Lower Devonian strata in northern New Brunswick. The Trout Valley specimens are not especially well preserved, but are sufficiently so as to leave no doubt as to their generic identity. The stems range in size up to 1.5 cm in diameter and are densely clothed with leaves that are 5-parted and about 5 mm long. Some of the leaves bear a single sporangium on the upper surface. The internal anatomy of the stems is known from the New York specimens where they reveal a well preserved, slightly fluted cylinder of xylem.

Numerous specimens of the curious fossil plant *Prototaxites* have been found in the conglomerate that is exposed along the upper reaches of South Branch Ponds Brook. This is a widely distributed Devonian plant, both geographically and stratigraphically. Its silicified trunks attain a maximum diameter of nearly a meter. These trunks consist of large and small tubes closely intertwined; the former are aligned longitudinally while the latter are randomly oriented. Although problematical as to its taxonomic affinities, *Prototaxites* is generally regarded as the stem of a giant alga or fungus. The Maine specimens found thus far are poorly preserved and do little more than reveal the presence of this plant in the flora.

Future field work may be expected to reveal still more new plants and to add to our knowledge of some presently known only from fragmentary remains. As one final example, we have

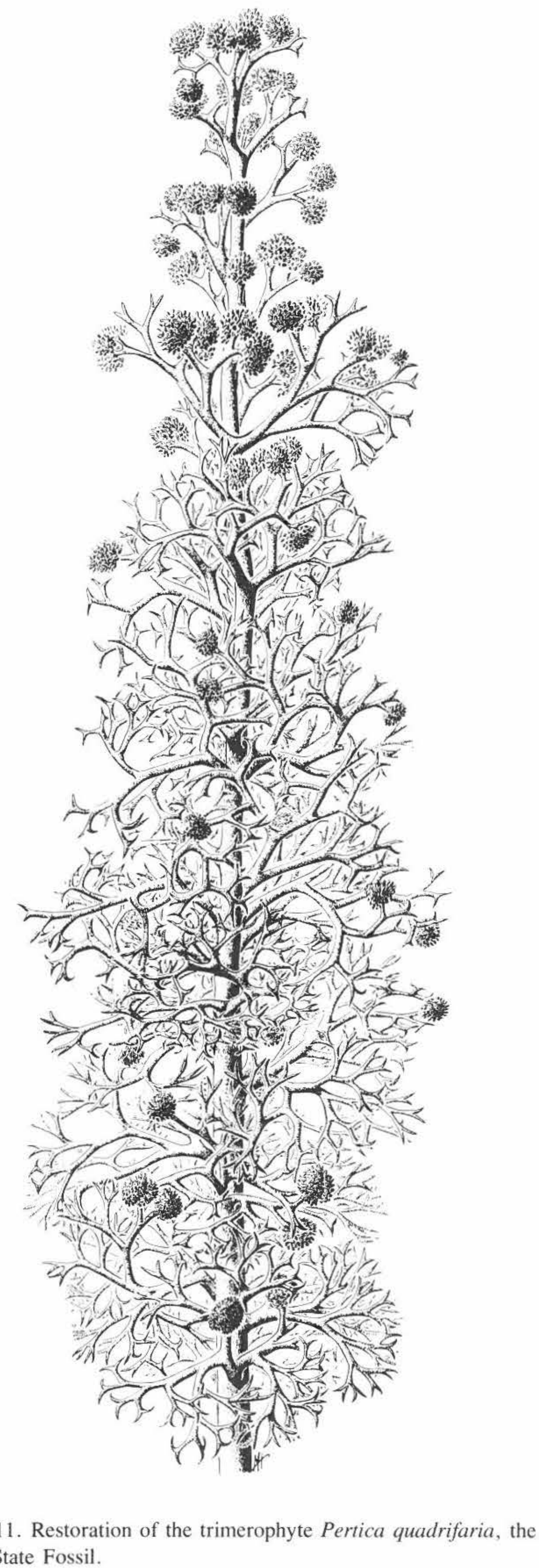

Figure 11. Restoration of the trimerophyte *Pertica quadrifaria*, the Maine State Fossil.

found distinctively spiny stems along Dry Brook about 200 meters upstream from where the road crosses the brook (Fig. 4, # 4). We tentatively attribute them to the lycopod genus *Drepanophycus*. The stems differ from any others known in the flora, but thus far no fertile specimens have been found. The area is heavily forested, but it is evident that significant and abundant fossils are present.

We are indebted to William G. Chaloner of Royal Holloway and Bedford Colleges, Surrey, England, for undertaking a study of the spores present in the flora. For the most part they are not well preserved, but the diversity of types suggests the presence of plants that have not thus far been found as macrofossils. Descriptions and illustrations of the spores are given in Andrews et al. (1977).

Flora of the Fish River Lake Formation

In their Perry Basin paper, Smith and White (1905) discuss various reports of coal from other regions in the state of Maine. One of these reports is from the Fish River extension of the Bangor and Aroostook Railroad in northernmost Maine. The purported "coal" was simply a seam of carbonaceous shale; however, this and other associated horizons contained fossil plants. White (Smith and White, 1905) assigned the plant material tentatively to two species of *Psilophyton* and suggested a Devonian age for the strata.

In 1970 Boone described the geology of this area of northern Aroostook County and established the Fish River Lake Formation for the bedrock. The formation is dated ". . . chiefly as late Silurian, but with the upper siltstones and volcanics extending across the systematic boundary to the New Scotland time-stratigraphic horizon" (Boone, 1970). Of possible significance is the presence of a marine fauna (brachiopods) in the same formation as the plant fossils. Boone assigned the plant material to *Psilophyton* and *Hostinella*, the latter a form genus for branching unornamented axes.

Kasper et al. (1974) briefly described and illustrated plants from the above locality (Nadeau Thoroughfare, Locality # 9) and another site (Red River, Locality # 8) in the Fish River Lake Formation (Table 2). The specimens are fragmentary but fertile, i.e., bearing sporangia. The smooth stems are longitudinally ribbed, branch dichotomously, and are 1.5-2.0 mm in diameter. Sterile axes end in recurved tips, and fertile axes end in clusters of elliptical paired sporangia 3-4 mm long and 1 mm wide. Because of the fragmentary nature of these plant remains, they are assigned to the form genus *Dawsonites*. More complete specimens may eventually permit their assignment to a species of *Psilophyton*. If the New Scotland upper age limit holds true, then the plants of the Fish River Lake Formation may eventually prove to be one of the earliest known occurrences for *Psilophyton*.

Fossils of the Frenchville Formation

Schopf et al. (1966) described carbonized dichotomizing axes from the Silurian Frenchville Formation in northern Maine (Table 2). The small axes were vertically oriented and perpendicular to the bedding plane. The specimens were proposed, at that time, to be the oldest remains of early land plants yet discovered. Recent reexamination of the material by Strother and Lenk (1983) has called into question the plant affinity of these controversial fossils.

MIDDLE DEVONIAN FLORA IN MAINE

The Mapleton Sandstone Flora

The presence of plant fossils in the Mapleton Sandstone was noted as early as 1900 by Williams who identified axis fragments as *Psilophyton princeps*. No serious study of them was made at that time, although specimens were sent to David White at the U.S. Geological Survey. Interest centered more on determining the age of the Mapleton Sandstone because of its lithological similarity to the Chapman Sandstone and because it had originally been considered as part of, or included within, the Chapman Sandstone. However, the Mapleton Sandstone unconformably overlies the Chapman Sandstone; further, the Chapman Sandstone is more deformed and indurated, and contains marine fauna and some plant remains that differ from and are stratigraphically older than those present in the Mapleton Sandstone. Boucot et al. (1964) also point out that apparently the Mapleton Sandstone, dated as Middle Devonian in age, is the oldest formation not affected by Acadian folding in that area.

Serious study of the plant fossils of the Mapleton Sandstone began when James M. Schopf received some specimens from Richard Naylor in 1961 for age determination and paleobotanical study. Schopf later obtained additional specimens from Forbes and collected more himself in 1963. He published a report on the plant megafossils from the Mapleton Sandstone in 1964. Schopf identified spiny axes as *Psilophyton princeps* var. *ornatum* (= *Sawdonia ornata*), fragmentary smooth axes of differing sizes and branching pattern as *Hostinella* and *Aphyllopteris* (form genera for smooth axes), stems with "petiolate" lateral appendages as *Barrandeina*, and stems and sporangia as a new species of *Calamophyton*, *C. forbesii* (Table 2). In an appendix to a report on the geology of the area by Boucot et al. (1964), Schopf provided an assessment of the age of the Mapleton Sandstone based on spores obtained by macerating

Figure 12. Permineralized (petrified) plants from the Middle Devonian Mapleton Sandstone, Maine. (a) A possible zosterophyll axis with exarch protostele, most comparable to *Stolbergia*. X 30. (b) A possible second type of zosterophyll axis. X 33. (c) Axis with a vascular strand typical of *Cladoxylon* and other cladoxyls. X 30. (d) Axis with deeply lobed strand reminiscent of some iridopterids. X 25. (e) Axis, presumably stem, with secondary xylem and probably of aneurophyte (progymnosperm) affinity. X 14. (f) Axis, presumably root, with secondary xylem and probably of aneurophyte affinity. X 30.

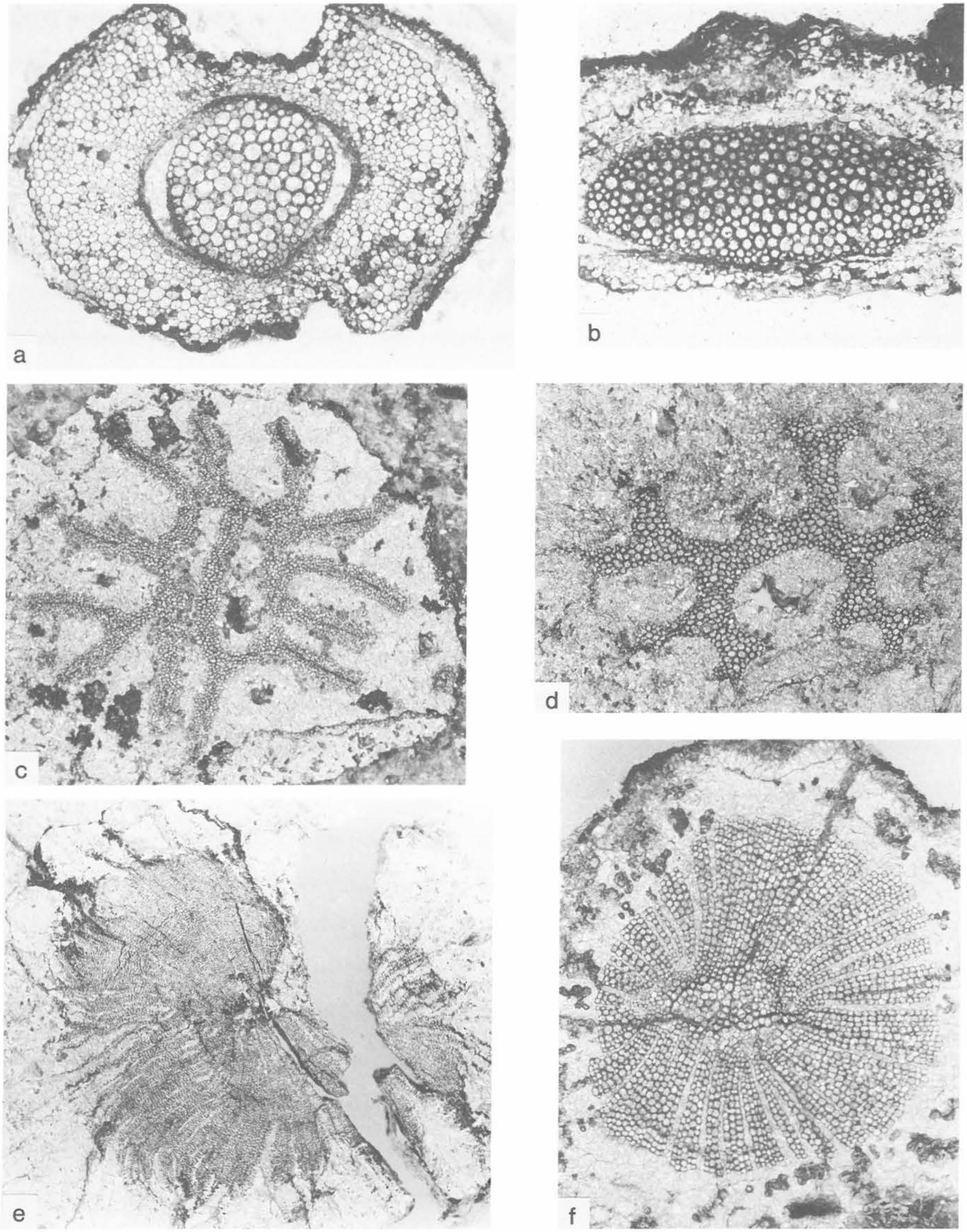

Figure 12.

some specimens of the plant-bearing rock. Both spores and plants suggested a Middle Devonian, perhaps early Givetian, age for the sediment although none of the microfossil or megafossil taxa are fully diagnostic of that age; it is possible that the strata may be somewhat older. Schopf clarified the differences in the types of plants present in the Chapman Sandstone relative to those of the Mapleton Sandstone.

Forbes was instrumental in drawing the attention of one of us (Gensel) to the potential for further study of the Mapleton Sandstone. We collected at two sites, the Winslow Farm locality and a roadside outcrop, starting in 1976. The plant fossils are preserved as fragmentary compression/impression remains and also as permineralized (pyrite and/or limonite) axes. Initial observations suggest that about six different kinds of plants are present. One or possibly two zosterophylls occur–their stems having a solid core of xylem in which the first-formed elements are located to the periphery (an exarch protostele; Figs. 12a,b). One of these (Fig. 12a) compares very closely in overall histology to *Stolbergia* described from Middle Devonian sediments in Belgium by Fairon (1967). *Stolbergia* is based on smooth axes identical to those considered to represent the distal sterile segments of a plant called *Asteroxylon elberfeldense* by earlier workers, but regarded as a separate taxon by Fairon (1967). Fertile structures of *Stolbergia* are unknown.

Other permineralized axes represent the stems of plants included in the cladoxylales, with one type appearing most similar to *Cladoxylon* or *Calamophyton* (Fig. 12c) and another apparently representing a new taxon (Fig. 13a). Other axes show anatomy consisting of a lobed, spread-apart, vascular strand (Fig. 12d) which is reminiscent of members of the Iridopteridales or *Schizopodium*. Also present are one or possibly two types of stems and perhaps one type of root with secondary xylem which may represent members of the aneurophytalean progymnosperms (Figs. 12e,f). Another appears to represent a rachis of the pre-fern *Rhacophyton*. These will be the subject of more detailed study and description in the near future.

Correlation of the permineralized axes with the morphologically preserved plant remains is possible only to a limited extent because little overlap in preservational type occurs. Schopf (1964) reported the occurrence of *Calamophyton* based on compressions/impressions from the Mapleton Sandstone, and observations by Gensel confirm his identification (see our Fig. 13b). One of the cladoxylalean stems may represent that taxon; however, none of the permineralized axes of that type conclusively demonstrate morphology typical of *Calamophyton*. Small axes covered with spines, possibly the basis for the identification of *Psilophyton princeps* var. *ornatum* by early workers and Schopf, are common in our current collections (Fig. 13c). This species has since been renamed *Sawdonia ornata* by Hueber (1971) based on other collections of spiny axes and has been assigned to the zosterophylls. Other Mapleton spiny axes produce slender, dichotomizing lateral branches which differ in morphology from those of *S. ornata* or any other zosterophyll and are more similar to specimens of possible trimerophyte or aneurophyte affinity. Schopf (Boucot et al, 1964) also mentioned a similarity in spine morphology to the poorly understood genus *Thursophyton*, but the spines seem too slender.

Fragments of fertile remains occur, including two kinds of sporangial clusters attached to short lengths of axes (Figs. 13d,e). These tend to resemble trimerophyte sporangia, but might as readily represent the fertile parts of an aneurophyte. Isolated fusiform sporangia are also present. The plants described briefly above support a Middle Devonian age for the Mapleton Sandstone. Most particularly, plants with secondary xylem such as those present in this flora are not known to occur prior to the Eifelian anywhere in the world (Banks, 1980). Cladoxyls and iridopterids are common in Middle Devonian sediments, although the former may first appear in latest Early Devonian (Lessuise and Fairon-Demaret, 1980). Zosterophylls range from Early to Late Devonian in age. Study of the microfossil assemblages is currently underway and thus far confirms Schopf's age assessment as well.

Further study of the flora of the Mapleton Sandstone will not only contribute significant information about plant evolution in that period of time, but also will allow for floristic comparison with Middle Devonian plants from other parts of eastern North America, namely New York State and Virginia, as well as with well-known Middle Devonian floras of Germany, Belgium and Czechoslovakia. The plants identified in the Mapleton Sandstone thus far are very similar to those occurring in some Givetian localities (Cairo) in New York State (Matten, 1973, 1975) and to some European Middle Devonian floras–differing mainly in paucity of aneurophytes.

LATE DEVONIAN FLORA IN MAINE

Perry Formation Flora

As mentioned earlier, the history of paleobotany in Maine begins with the reports of plant fossils from Upper Devonian strata in the vicinity of the town of Perry along Maine's easternmost coast. In three papers in the 1860's, Dawson describes and illustrates plant megafossils from strata which are eventually to be named the Perry Formation. Dawson's immediate contribution to the geology of Maine was his accurate assignment of a Devonian age to the Perry Formation. More significant, however, is the fact that these three initial papers (1861, 1862,

Figure 13. Plants from the Middle Devonian Mapleton Sandstone, Maine. (a) Permineralized axis with cladoxylalean anatomy, probably different from axis in Fig. 12c. X 28. (b) Axis showing digitate branching characteristic of *Calamophyton*, a genus considered by some in the line leading to horsetails. X 1.1. (c) Branched, spiny axes; possibly similar to specimens called *Psilophyton princeps* by earlier workers; arrow indicates spines. X 1.6. (d) Axis with short lateral branches terminating in clusters of fusiform sporangia. X 2.4. (e) Distal axis segments, probably of another plant, terminating in fusiform sporangia. X 5.

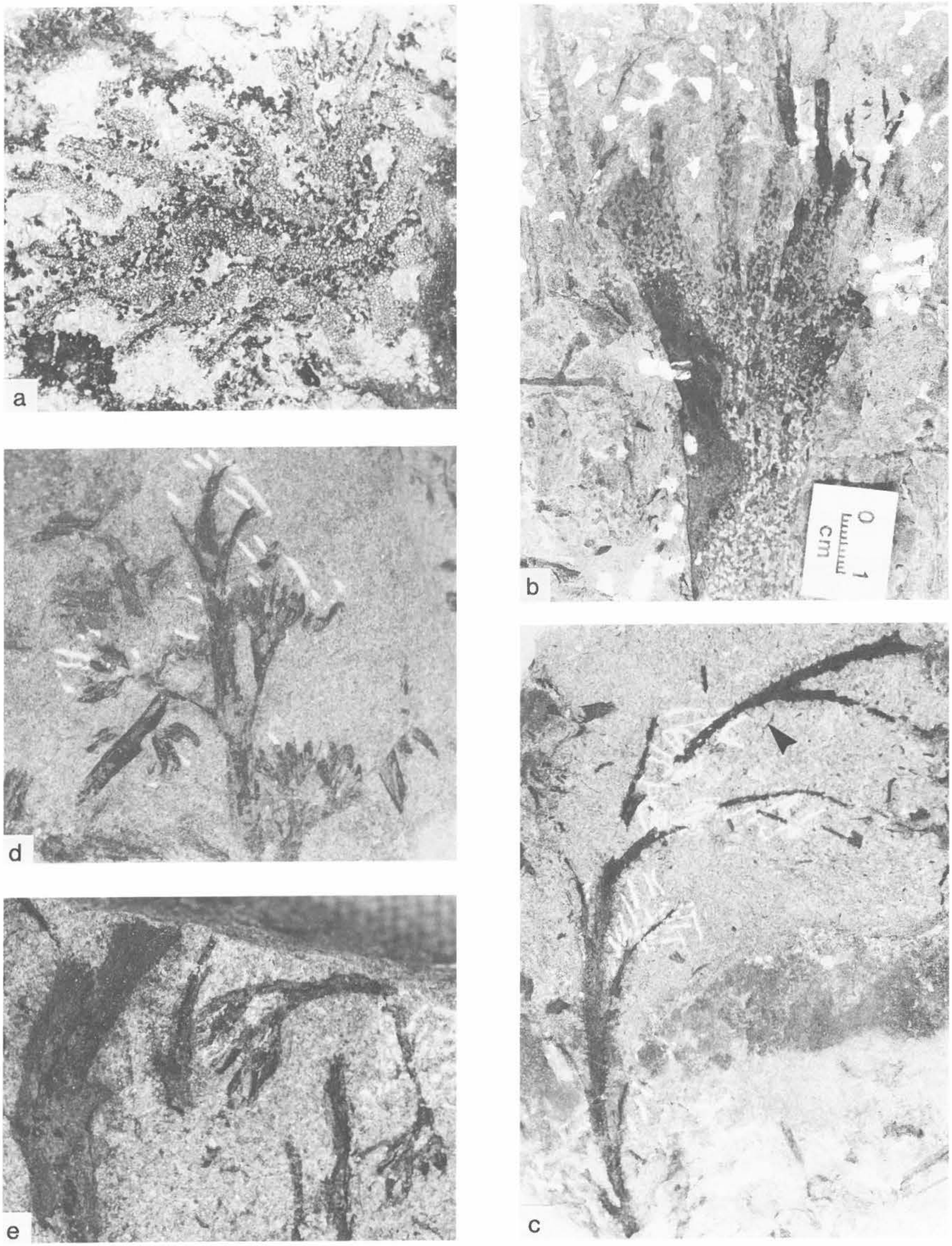

Figure 13.

1863) and his summary paper of 1871 laid the foundations for Devonian paleobotany in North America.

It is critical to point out that fossil plant specimens from the Perry Formation are rare. Dawson candidly admits that much of his material from the Perry Formation comes from other individuals. The specimens he describes in 1861 are those collected by Richardson for the Geological Survey of Canada and those deposited in the Natural History Society of Portland. Dawson did, however, have an opportunity to collect plant material from the Perry Formation himself with the help of Jethro Brown (Dawson, 1863, p. 460). The locality, as can best be pinpointed from Dawson's rambling narrative, is on the "south side of Little River" which is near Perry, Maine. Smith and White (1905), in their comprehensive treatment of the geology of the Perry Basin, provide two localities for plant fossils in the Perry Formation. One of them, from which they themselves collected material, is believed by them to be the same locality which Dawson himself worked. This is the outcrop on the "right bank of Little River, about one-half mile below the wagon-road bridge at Perry" (Smith and White, 1905, p. 35). Descriptions thereafter of fossil plant specimens from the Perry Formation have been based on museum specimens only (Kräusel and Weyland, 1941; Pettitt, 1965). In 1968 Forbes relocated the original outcrop and subsequent collections have been made by two of us (Forbes and Kasper) over several years. The site is on the south bank of Little River about 500 yards east of the bridge where U.S. Route 1 crosses the river near Perry, Maine. Because only a small peripheral portion of the fossiliferous lens is exposed, only a few specimens can be collected in any one field season. Apparently, each year's winter ice erodes the outcrop and exposes again a small peripheral segment of the lens.

The two most important species of fossil plants in the Perry Formation are *Barinophyton richardsonii* and *Archaeopteris jacksonii* (Posnick, 1982) (Table 2). Both of these were established by Dawson in his first publication on the Perry flora in 1861. The first species Dawson (1861) initially called *Lepidostrobus richardsonii* thinking that it was a lycopod strobilus (cone) and naming it after Richardson of the Canadian Geological Survey who had found the specimen. Two years later Dawson (1863) placed this species in the genus *Lycopodites* as *Lycopodites richardsonii*. White (Smith and White, 1905) in his revision of the Perry flora recognized that the specimens described by Dawson were not lycopod remains at all but completely new. As a result White established a new genus *Barinophyton* for Dawson's material, hence, *Barinophyton richardsonii*.

Prior to our recent collections, less than a dozen specimens of *B. richardsonii* had been obtained by Dawson and his colleagues or Smith and White. The species is currently being redescribed based on new fertile specimens. The complete plant is unknown, but the largest specimen (Fig. 14a) is 9.6 cm long and shows a main axis bearing two rows of strobili on either side: alternate and distichous (Posnick et al., 1983). The strobili number 12 in all. The main axis is 5 mm wide and has three of the strobili directly connected. Each strobilus consists of an adaxially curved axis bearing what are interpreted as appendages and sporangia on the adaxial surface. The strobili are broken distally, but are at least 4 cm long and 0.8-1.2 cm wide. Width depends on their orientation during preservation. In 1965 Pettitt macerated a specimen of *B. richardsonii* in the British Museum collected from the Perry Formation. He obtained megaspores 220-250 μm in diameter and microspores 48-62 μm in diameter. We have been unsuccessful, as yet, in obtaining spores from sporangia.

One unique and startling feature of *Barinophyton* is the fact that Brauer (1980) has demonstrated conclusively that *B. citrulliforme* from Upper Devonian strata in Pennsylvania has both megaspores and microspores within the *same* sporangium. This is an important discovery of what may be a major step in the transition from homospory (one size spore with bisexual gametophyte) to heterospory (two sizes of spores with male and female gametophytes). Hopefully, further work on *B. richardsonii* will shed light on this major question of reproductive biology in early vascular plants.

The second important species of the Perry flora was originally described by Dawson as two species: *Cyclopteris jacksonii* and *Cyclopteris rogersii*. Dawson established *C. jacksonii* in 1861 naming the material after Charles T. Jackson, the first state geologist. Two years later Dawson (1863) erected the other species and named it for Professor W. B. Rogers. White (Smith and White, 1905) in his floral revision placed these two species in the genus, *Archaeopteris*, as *A. jacksonii* and *A. rogersii*. For reasons to be discussed shortly, all the specimens of *Archaeopteris* have been treated by us as one species: *Archaeopteris jacksonii*.

Archaeopteris is an arborescent progymnosperm (Fig. 15) with a large trunk displaying features of gymnosperm wood anatomy. Its branch systems consist of helically arranged laminar leaves and non-laminar leaves. The latter bear numerous elliptical sporangia on their upper surface.

The *Archaeopteris* specimens from the Perry Formation are small branch fragments of sterile and fertile foliage (Figs. 14b-d). Several specimens of branchlets with sterile leaves have been collected. Sterile leaves (Fig. 14c) are generally obovate in outline, tapered proximally to the point of attachment, dichotomously veined, entire margined, and average about 16 mm long and

Figure 14. Plants from the Upper Devonian Perry Formation, Maine. (a) Specimen of *Barinophyton richardsonii* showing a portion of axis (bottom) and two rows of strobili; arrow indicates sporangia. X 0.8. (b) Penultimate branch and ultimate branchlets bearing sterile foliage of *Archaeopteris jacksonii*; the specimen (USNM 422472) is figured in Smith and White, 1905 (Pl. 3, Fig. 1) and is the counterpart to Dawson's specimen (1871, Pl. 15, Fig. 167). X 0.7. (c) Ultimate branchlet of *A. jacksonii* showing obovate sterile leaves; the specimen USNM 36460 is figured in Smith and White, 1905 (Pl. 3, Fig. 2). X 1.9. (d) A segment of a fertile branchlet presumably of *A. jacksonii*; each filiform leaf bears several sporangia on its upper surface. X 2.7.

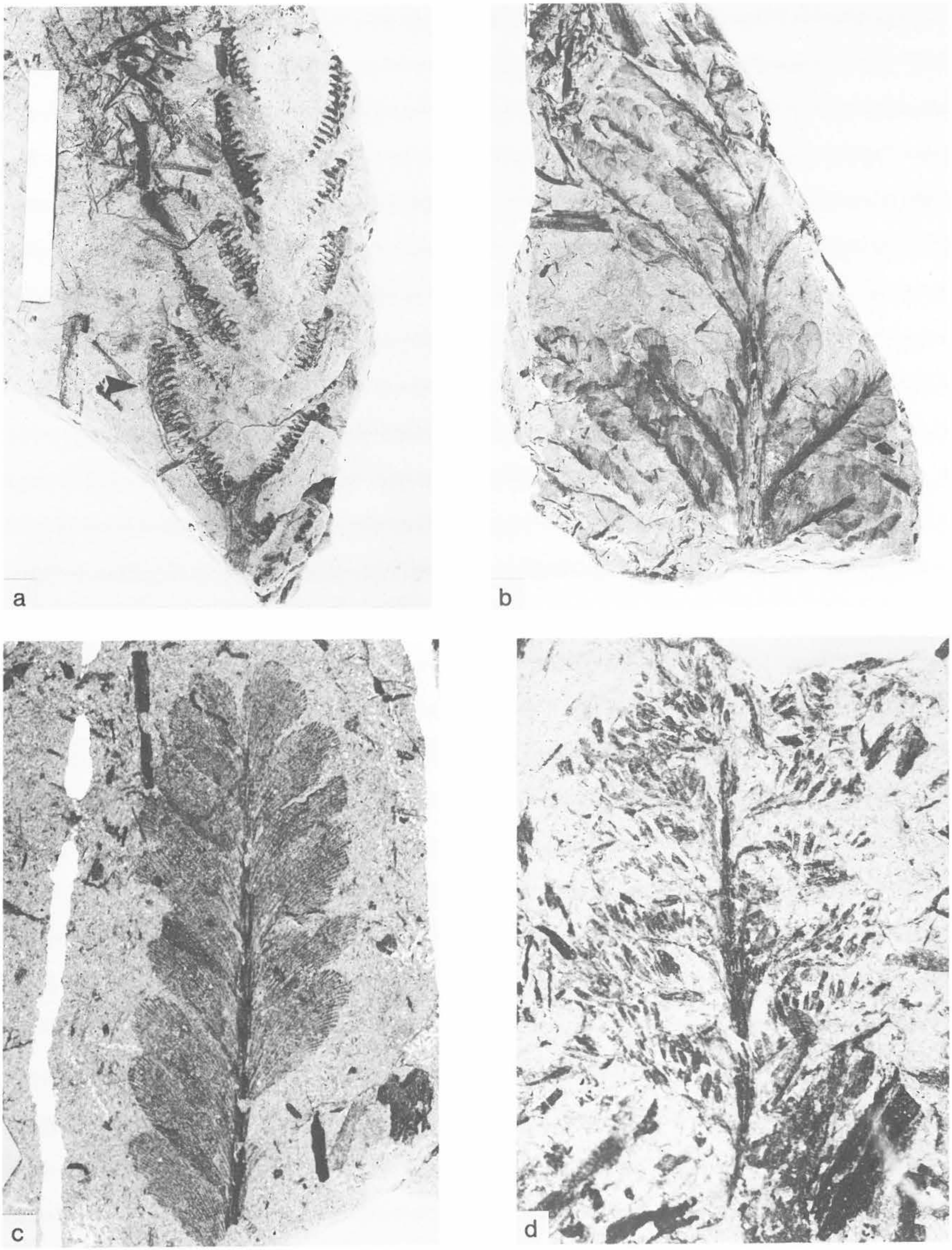

Figure 14.

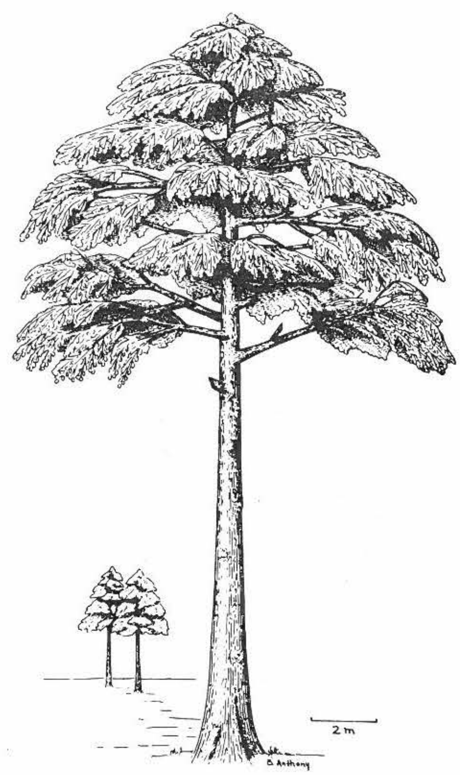

Figure 15. Restoration of the Upper Devonian progymnosperm tree, *Archaeopteris* (from Beck, 1962).

6-7 mm wide (Posnick, 1982). However, leaf size diminishes toward the end of the ultimate branchlets to 8.6 mm long and 4.2 mm wide. There is regular overlapping of adjacent leaves (Fig. 14c).

Branchlets with fertile foliage have also been recovered (Fig. 14d). These generally are less well-preserved than the sterile ones. The filiform fertile leaves are about 2 mm wide, 10-11 mm long (Posnick, 1982) and are reported to divide several times (Phillips et al., 1972). The linear to fusiform short-stalked sporangia are generally 2.0-2.5 mm long and 0.5-0.7 mm wide. Presumed microspores obtained from one fertile specimen were unornamented, trilete, and measured 45 μm in diameter (Posnick, 1982). Further collection and study will, hopefully, produce better fertile specimens and yield megaspores and microspores within the sporangia.

The genus *Archaeopteris* is very important in paleobotany. First, it is an Upper Devonian index fossil for the start of Banks' (1980) Assemblage-Zone VI, the *Archaeopteris* Zone. Second, it occurs in some abundance where found and has a widespread geographical distribution. Third, it is the basis on which an entirely new plant class, the Progymnospermopsida, was established by Beck in 1960. *Archaeopteris* is perhaps the best known of Late Devonian plants morphologically, anatomically, and reproductively. However, it presents one of the major enigmas in paleobotany: the many species of *Archaeopteris* which are based on sterile foliage all have reproductive foliage which is virtually indistinguishable from each other. This difficulty was addressed by Beck in 1969. The problem concerning *Archaeopteris* is whether we are dealing with one highly variable species, with many species or, perhaps, with different genera. Since there is often morphological variability in vegetative features and since reproductive structures in all species are indistinguishable, it is not known what actually constitutes a biological species for *Archaeopteris* (Beck, 1969). For these reasons and for the fact that Dawson's taxa are indistinguishable based on his illustrations and descriptions, the *Archaeopteris* material in the Perry Formation is treated as one species—that first described by him as *A. jacksonii* (Posnick, 1982).

PALYNOLOGY

Determining the exact age of terrestrial sediments is often problematical, and this is true of some of the plant-bearing deposits in Maine. The kinds of plants found in the Mapleton flora, for example, are not in themselves diagnostic of a particular time interval, although a general designation such as Early, Middle, or Late Devonian can be made. Other types of evidence, e.g. faunal, are not available or are also inconclusive.

The study of dispersed spores in Devonian sediments has greatly expanded in recent years. As a result, the correlation and dating of sediments are much improved and are on a much broader scale than before. The accumulated data have been summarized in the recent publication by Richardson and McGregor (1986) on spore zones for the Silurian and Devonian of the Old Red Sandstone Continent and adjacent regions. This should aid in regional and worldwide correlation.

Palynological study of rocks of both the Trout Valley Formation and Mapleton Sandstone has been carried out. In regard to the former, spores are very highly coalified and black, thus limiting identification and analyses. However, the taxa described by Chaloner in Andrews et al. (1977) include *Apiculiretusispora, Grandispora douglastownense, Emphanisporites,* and some other genera which mostly support a late Early Devonian (Emsian) age for the formation. Richardson and McGregor (1986) cite the Trout Valley Formation assemblage as representative of their late Emsian, possibly earliest Eifelian, "douglastownense-eurypterota" zone. Therefore, this flora is generally comparable in age or perhaps slightly younger than the floras of northern New Brunswick and the Gaspé Peninsula which are part of the same tectonic framework. Many plant taxa described from the Trout Valley Formation also occur at the Canadian localities which range from early to latest Emsian in age based on dispersed spore correlations (Gensel and Andrews, 1984; McGregor, pers. commun., 1979, 1981).

Schopf, in Boucot et al. (1964), recorded the occurrence of the dispersed spore genera *Cyclogranisporites, Tholisporites, Auroraspora,* and large spiny spores similar to ones called "Type H" by Lang (1925) (= *Corystosporites* or *Acinosporites macrospinosus*) from the Mapleton Sandstone. The two genera, *Auroraspora* and *Corystosporites*, are typical of Middle Devo-

nian spores and that, combined with the absence of spores with bifurcate spines and some distinctive Upper Devonian forms, led Schopf to suggest a Middle Devonian (probably early Givetian) age for the sediments. One of us (Gensel) is conducting further study of the dispersed spores from these sediments.

Andrews et al. (1977) noted that dispersed spores have been obtained from the Perry Formation sediments, but have not yet been studied. The only palynological report from the Perry Formation is that of spores obtained from sporangia of *Barinophyton* by Pettitt (1965). Two of us (Kasper and Forbes) are currently working on a redescription of *Barinophyton* from the Perry Formation and hope to add further information regarding the spores. No precise age, other than Upper Devonian, has been assigned to the Perry Formation. Analysis of the dispersed spores might allow better correlation with other Upper Devonian deposits worldwide.

In summary, then, even though plant fossils have been known from the state for better than a century, the intensive work of the past two decades has vastly increased our knowledge of plant life in the Devonian of Maine and northeastern North America. The interest and cooperation of local residents, professionals working in the region, and the Maine Geological Survey have played a major role in providing impetus, information, and encouragement for the initial discoveries and for our ongoing work in the state. It must be remembered that the results obtained over the last two decades have taken place in an area long believed to be barren of any useful plant fossils. Also evident is the fact that repeated collecting at known outcrops and continued exploration for new plant-bearing strata are likely to yield even more data. Plant fossils from the state of Maine have been a major factor in understanding the remarkable diversity of early land vascular plants of the Devonian Period.

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Transmission Electron Microscope Study of Bedding-Cleavage Relations in the Vassalboro Formation, East-Central Maine

Beth Z. Lincoln
Department of Geological Sciences
Albion College
Albion, Michigan 49224

ABSTRACT

Transmission electron microscope study of a suite of samples from the Ordovician-Silurian-aged Vassalboro Formation, east-central Maine, provides evidence bearing on the origin of cleavage. Transmission electron microscopy is essential to this study because the fine grain size of these rocks makes textural details impossible to resolve with the petrographic microscope. These rocks are characterized by a spaced cleavage consisting of alternating cleavage zones and uncleaved microlithons. The mineral assemblage in the cleavage zones differs from that in the microlithons, notably in the lack of quartz in the cleavage zones. Phyllosilicates in the cleavage zones occur in two habits: as platy grains, and as blocky grains. Typically, the blocky grains are elongated perpendicular to their (001) planes and parallel to the rock cleavage. The cleavage zones and microlithons also have different average chemical compositions. The cleavage zones are depleted strongly in most major oxides, especially SiO_2 and Na_2O, relative to the microlithons. It is proposed that the cleavage zones formed by the partial, or, in some cases, complete, dissolution and removal of minerals, especially quartz, chlorite, and white mica, and by the growth of chlorite and white mica.

INTRODUCTION

Since at least the time of Sharpe (1849) and Sorby (1853; 1856a; 1856b), geologists have been intrigued by the deformation of micaceous rocks, and especially by the formation of slaty cleavage. The relationship of cleavage to the stress and strain ellipsoids, the mechanism by which a preferred mineral orientation originates, and the relationship of slaty cleavage to other forms of cleavage are among the problems which have been examined. Many of these problems remain controversial, and fundamental questions on the origin of cleavage remain unanswered. This study documents the textures of cleaved rocks and thus presents key evidence bearing on the origin of slaty cleavage.

A suite of phyllites from the Vassalboro Formation, east-central Maine, was examined using optical and transmission electron microscopy. Transmission electron microscopy is an essential component of this work, because the grain size of these rocks is too fine to allow textural details to be resolved adequately with the petrographic microscope. The transmission electron microscope is capable of extremely high magnification, and so is a powerful tool in the study of textures of fine-grained rocks. One drawback, however, is that only very small areas (approximately several grain diameters across) can be examined. Much preliminary work with the optical microscope is necessary to ensure that representative areas of the samples are selected for the transmission electron microscope, or a biased view of the samples will result.

This suite was chosen for study because it contains a spaced cleavage defined by alternating zones of relatively uncleaved and strongly cleaved rock. Elsewhere in the Vassalboro Formation, this spaced cleavage grades into a well-developed slaty cleavage (Schwartz, 1976). Assuming that this spaced cleavage represents the early stages of the development of a penetrative slaty cleavage, then the transition from uncleaved to strongly cleaved rock is preserved in these rocks. Such rocks can provide more information on the mechanism of cleavage formation than can rocks containing one well-developed cleavage, because one can compare directly the uncleaved and cleaved zones, and so gain information on the progressive de-

velopment of cleavage.

GEOLOGIC SETTING

The Vassalboro Formation has been described by several previous workers, including Osberg (1968) and Griffin and Lindsley-Griffin (1974). It is dominantly a thick-bedded, slightly calcareous metasandstone with thin interbeds of phyllite. It is described as being similar to the Siluro-Devonian-aged Madrid Formation (Ludman and Griffin, 1974), and early work suggested that these units were correlative (Osberg et al., 1968; Griffin, 1973; Ludman and Griffin, 1974), but later work suggests that the Vassalboro Formation is Ordovician-Silurian in age (Osberg et al., 1985). It has been metamorphosed to the chlorite and biotite grades of greenschist metamorphism, and several episodes of tectonic folding and faulting have affected these rocks. Sedimentary and slump features still can be recognized, however. The cleavage described in this paper is the earliest and, in these samples, shows no evidence of overprinting by younger events.

The samples* examined in this study are from the portion of the formation informally referred to by Griffin (1973) as the "Kenduskeag formation", a name no longer in use. These samples lie in the chlorite zone of the greenschist facies of metamorphism. Several generations of cleavages are contained in this outcrop, as are original depositional features such as graded bedding, cross laminae, slump folds, and sedimentary breccias. Samples were chosen in which bedding and the earliest cleavage are both pronounced and at a high angle to each other. Here, this early cleavage is a spaced cleavage; elsewhere in the region it has been described as a penetrative slaty cleavage and as a crenulation cleavage (Schwartz, 1976). Schwartz (1976) concluded that the cleavage began as a spaced cleavage, and that in some areas the spacing between the cleavage zones decreased as the deformation progressed, resulting in a penetrative slaty cleavage. The goal of this present study is to use textural evidence to document the processes by which cleavage development progresses.

OBSERVATIONS

Compositional variation in the samples permits the study of compositional influences on cleavage formation. Some of the beds are quartz-rich or psammitic, some are phyllosilicate-rich or pelitic, and some are graded from psammitic to pelitic.

In hand sample and thin section, discrete cleavage zones cut across and disrupt bedding (Figs. 1 and 2). The spacing between these zones varies between a centimeter in psammitic beds and a fraction of a millimeter in pelitic beds. The areas in which sedimentary structures are preserved between the cleavage zones are referred to as microlithons in this study. It is believed that

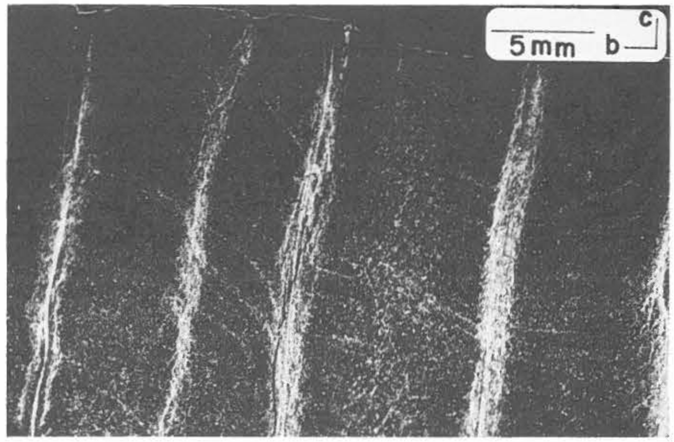

Figure 1. Photograph of a psammitic layer. Negative print directly from thin section; dark areas are rich in quartz, while light areas are rich in oxides and phyllosilicates. On all photographs, b indicates the bedding direction and c the cleavage direction. The light-colored zones, spaced 5 to 10 millimeters apart in the sample, are cleavage zones. Thin bedding laminations in the dark-colored microlithons are inclined at high angles to the cleavage plane and are rotated toward parallelism with the cleavage near several of the cleavage zones.

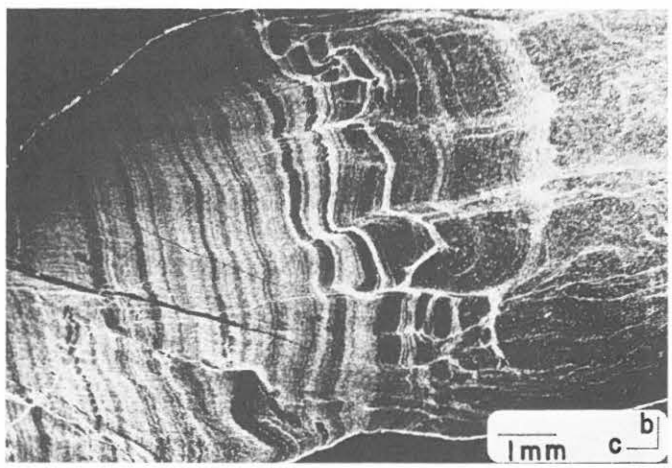

Figure 2. Negative print, from thin section, of a pelitic sample. The light-colored cleavage zones here cut the bedding lamination at high angles and produce apparent offsets in this lamination. The cleavage zones are thinner and more closely spaced than those in the psammitic layer in Figure 1.

these microlithons were not affected significantly by the formation of the cleavage zones, and so represent the original rock, prior to cleavage formation. The cleavage and bedding planes are approximately perpendicular to each other.

Optical and transmission electron microscopy reveal striking differences between the cleavage zones and microlithons, especially in the psammitic beds. Boundaries between these two zones are sharp; grains do not bend from one orientation into another. In the microlithons in the psammitic units, the mineral assemblage is quartz, white mica, plagioclase feldspar, chlorite, iron oxides, iron-titanium oxides, and minor amounts of calcite (Fig. 3). Quartz is the dominant constituent. Both quartz

*The samples were collected by H. R. Burger III at Stop #4 of field trip A-3, 1974 New England Intercollegiate Geological Conference (Griffin and Lindsley-Griffin, 1974).

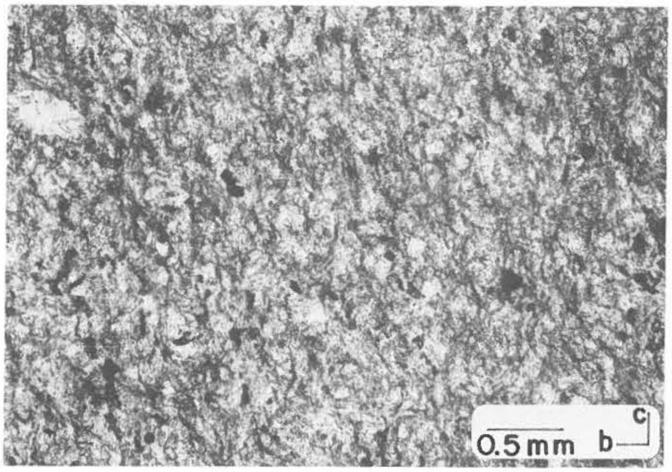

Figure 3. Detail in plane-polarized light of one microlithon in the psammitic layer shown in Figure 1. The white grains are dominantly quartz. White mica lies along quartz grain boundaries, and can be found in many orientations. Sparse stringers of opaque material lie parallel to the rock cleavage.

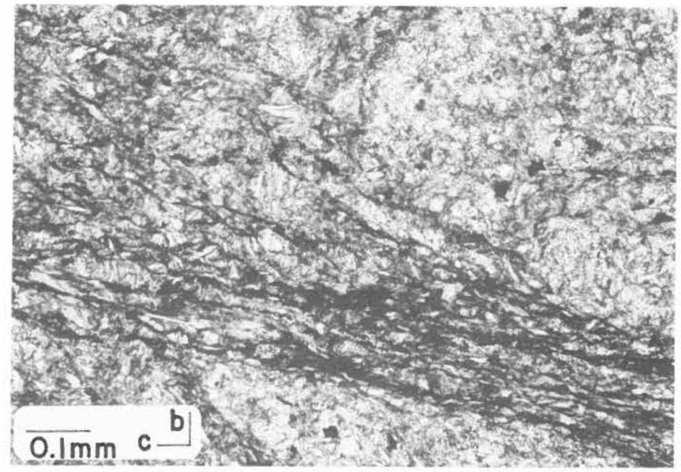

Figure 4. Detail in plane-polarized light of one cleavage zone in a psammitic layer. The cleavage zone at the bottom is dominated by the dark trails of iron oxides and iron-titanium oxides parallel to the plane of the cleavage. Both the flake-like and the blocky habits of the phyllosilicate grains occur within the cleavage zone. While some flake-like phyllosilicate grains have their long dimensions parallel to the rock cleavage, many do not. The blocky grains tend to have their (001) planes at high angles to the rock cleavage. Those that are elongated perpendicular to their (001) planes thus have their long dimensions approximately parallel to the rock cleavage. A comparison of the microlithon at the top of the photograph with the cleavage zone on the bottom reveals that quartz (which appears as equant white grains in the microlithon) is much less abundant in the cleavage zone than in the microlithon.

and feldspar occur as equidimensional grains and are subrounded to subangular in shape, and thus are interpreted to be detrital. Grain diameters in the microlithons equal 60 micrometers or less. Quartz does not exhibit undulose extinction. The phyllosilicates occur as isolated grains along quartz grain boundaries and do not form mica beards. There is a tendency for the white mica to lie parallel to the plane of the cleavage zones, but some white mica lies parallel to the bedding plane and some in all orientations between bedding and cleavage. The phyllosilicates commonly are lath-like in cross-section, with their long axes parallel to the traces of their (001) planes. The lengths of these long axes are roughly the same as the diameters of the quartz grains. The chlorite occurs as coarse grains, commonly interlayered with white mica.

In contrast to the microlithons, the cleavage zones contain very sparse quartz. Instead, cleavage zones are composed of white mica, with lesser amounts of chlorite, brown iron oxides, and iron-titanium oxides. The phyllosilicates commonly are oriented with neither their long axes nor the traces of their basal planes parallel to the rock cleavage direction (Fig. 4). The white mica exists in two habits. It occurs as flake-like grains which have the trace of their (001) planes parallel to their long dimensions. Some of these grains lie with (001) parallel to the cleavage zones, but many lie with (001) at a high angle to the zones. In a noteworthy difference from the microlithons, white mica also occurs in these zones as more nearly equant grains (Fig. 4). The traces of the (001) planes of most grains with this habit lie at a high angle to the rock cleavage, although the long dimensions of some lie roughly parallel to it. Chlorite exists only in this second habit. This observation is confirmed by an analysis of transmission electron microscope diffraction patterns; the (001) planes of white mica grains lack a strong preferred orientation, while the (001) planes of chlorite grains are oriented at a high angle to the rock cleavage. Many of the chlorite grains are elongated perpendicular to their (001) planes (Figs. 4 and 7). The phyllosilicates with this habit commonly abut iron-stained zones defining the cleavage. No phyllosilicates with this habit exist in the microlithons.

Iron oxides and iron-titanium oxides are concentrated in the cleavage zones. Many of the phyllosilicates are iron-stained, and wavy dark brown seams of oxides extend along the cleavage traces (Fig. 4). As mentioned above, the blocky phyllosilicate grains commonly abut against these oxide seams.

In the pelitic beds, the cleavage zones are thinner and more closely spaced than those in the psammitic beds (Fig. 2), but otherwise are identical. Pelitic microlithons contain less quartz and feldspar and more phyllosilicates than the psammitic microlithons, reflecting their original compositional difference (Fig. 5). The textural relationships are the same as in the psammitic layers already described. Boundaries between microlithons and cleavage zones are sharp, and grains do not bend from one orientation into another. Some isolated grains parallel to the cleavage direction are present in the microlithons. These are surrounded by grains parallel to the bedding plane (Fig. 6); perhaps these represent incipient cleavage zones.

In summary, the microlithons differ from the cleavage zones in both mineralogy and texture. The cleavage zones contain significantly less quartz and feldspar than the microlithons. It is especially noteworthy that the nearly equant phyllosilicates are found only in the cleavage zones (Fig. 7). The phyllosilicates lack a strong preferred orientation of their (001) planes in both

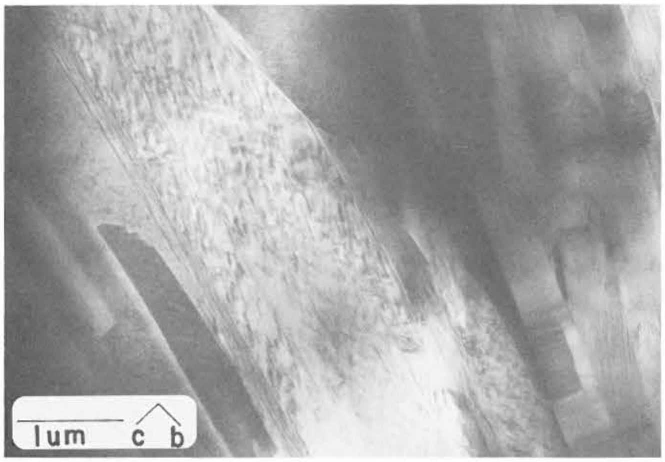

Figure 5. Transmission electron micrograph illustrating phyllosilicates in one of the microlithons in the pelitic sample shown in Figure 2. The high concentration of phyllosilicate grains is typical; in this sample quartz is scarce. In contrast to the phyllosilicates in the microlithons of the psammitic sample, the phyllosilicates in the pelitic sample tend to have their long dimensions and their (001) planes parallel to the bedding plane.

Figure 6. Transmission electron micrograph illustrating examples of isolated grains within the microlithon of a pelitic sample which lie parallel to the cleavage direction and may represent incipient cleavage zones. Note the sharp boundaries between grains of different orientations.

zones, although there is a tendency for white mica (001) planes to lie parallel to the cleavage zones within the zones themselves.

CHEMISTRY

Optical and electron microscopy reveal definite differences between the mineral assemblages in the microlithons and cleavage zones. The main difference is the scarcity of quartz, a major constituent of the microlithons, in the cleavage zones. The fine grain size of the rocks makes modal analysis impossible, so in an effort to quantify the differences between the two zones, each was analyzed for major oxides with an electron microprobe. Five areas in the microlithons and four in the

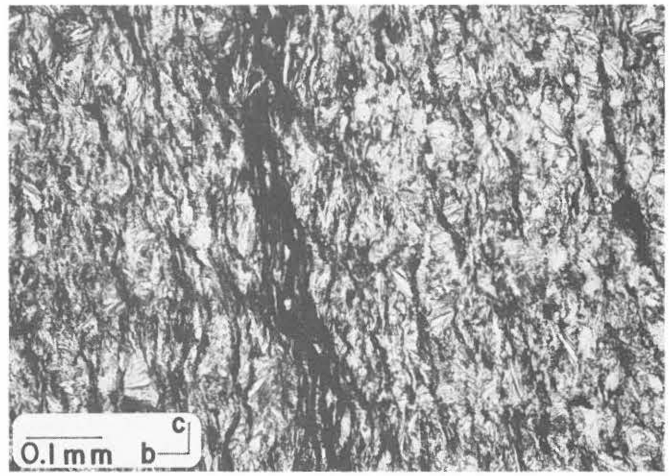

Figure 7. Optical micrograph, plane-polarized light, of a cleavage zone in a psammitic sample. Many phyllosilicates are elongated perpendicular to their (001) planes. At least one edge of these grains typically abuts one of the dark oxide traces. Solution along planes perpendicular to (001) is not enough to explain the shape of these elongate grains; grains this coarse do not exist in the microlithons. In order to have become so wide perpendicular to their (001) planes, they also must have added material along planes parallel to (001).

cleavage zones of a psammitic sample were analyzed, each at four or five spots, using a wide beam (40 micrometers in the cleavage zones and 70 micrometers in the bedding zones) to give average compositions. These are given in Table 1.

Overall averages for the cleavage and bedding zones are given in the first two columns of Table 2. Before meaningful comparisons between these two zones can be made, the composition of the original uncleaved rock must be known. In these rocks, this composition cannot be determined directly. However, it is possible to infer what it might have been. There are two extreme cases to consider. In the first, the rock behaves as a closed system and all the material removed from the cleavage zones is deposited in the areas between the cleavage zones. The original rock is differentiated into quartz-rich and phyllosilicate-rich domains; the composition of the original rock thus would be intermediate between those of the two domains. The case documented by Marlowe and Etheridge (1977) approaches this extreme. They studied a crenulation cleavage composed of alternating quartz-rich and mica-rich domains which they could trace into uncrenulated rock, and found that the composition of the uncrenulated rock indeed did lie between that of the quartz-rich and mica-rich domains for many of the components. Only MgO and FeO had been removed in significant quantities from these rocks (Marlowe and Etheridge, 1977).

In the second extreme case, the rock behaves as an open system and all the material removed from the cleavage zones leaves the rock. Some of the material that is dissolved may be reprecipitated elsewhere within the cleavage zone, but the microlithons between the cleavage zones retain their original composition. Cleavage described by Glasson and Keays (1978) exemplifies this case. This cleavage, like that described in this study, is defined by zones of phyllosilicates and titanium oxides that cut

TABLE 1. AVERAGE ANALYSES OF FIVE BEDDING AND FOUR CLEAVAGE ZONES, PSAMMITIC SAMPLE, VASSALBORO FORMATION (WEIGHT PERCENT).

	Microlithons*					Cleavage zones**			
SiO_2	80.85	80.96	70.69	67.36	82.97	46.19	42.75	41.45	42.02
K_2O	1.54	.84	2.17	2.33	1.09	6.37	7.38	7.65	8.60
CaO	0.08	0.09	0.63	0.11	0.11	1.49	0.15	0.27	0.26
TiO_2	0.38	0.51	0.50	0.43	0.25	1.77	2.75	2.67	5.21
MnO	0.03	0.02	0.06	0.07	0.03	0.07	0.07	0.07	0.13
FeO	2.62	2.99	6.88	7.13	3.09	7.38	8.56	7.75	4.53
Na_2O	2.77	2.72	1.27	2.48	2.32	1.03	0.44	0.18	0.21
MgO	1.19	1.00	2.91	3.19	1.27	3.84	4.37	4.18	2.79
Al_2O_3	9.21	7.64	12.06	14.45	8.39	24.05	27.29	26.68	27.01
Total	98.66	96.77	97.17	97.55	99.52	92.19	93.75	90.92	90.77

*Each column represents the average of five spot analyses.
**Each column represents the average of four or five spot analyses.

TABLE 2. COMPARISON OF AVERAGE CLEAVAGE ZONES AND MICROLITHONS, VASSALBORO FORMATION.

	Cleavage (Weight percent)	Microlithon	Clvg-microlithon/ microlithon	Net Change assuming immobile TiO_2	% Change assuming immobile TiO_2
SiO_2	43.10	76.57	-44	-70.67	-93
K_2O	7.50	1.59	372	-0.61	-38
CaO	0.54	0.20	170	-0.13	-65
TiO_2	3.10	0.41	656	0	0
MnO	0.09	0.04	125	-0.03	-75
FeO	7.06	4.54	56	-3.62	-80
Na_2O	0.47	2.31	-80	-2.25	-97
MgO	3.80	1.91	99	-1.42	-74
Al_2O_3	26.26	10.35	154	-6.94	-67

across microlithons rich in quartz, feldspar, white mica, and chlorite. On the basis of the compositions and textures of the cleavage zones and microlithons, they determined that dissolved material migrated out of the rocks along the cleavage planes and precipitated elsewhere (Glasson and Keays, 1978).

Intermediate cases are of course possible. Some components may be removed totally, others may be removed partially, and still others may be reprecipitated nearby.

The samples studied appear to approximate the open system case. There is no clear-cut petrographic evidence for the addition of material to the microlithons from the cleavage zones. Instead, textures indicate that most of the quartz in the microlithons is of detrital origin, and only one quartz-filled strain shadow was found in the thin sections examined. White mica seems to have recrystallized in these zones, but this most likely represents the response to metamorphism of mica already present in these zones. It is not likely that a major amount of material has been removed from the microlithons; this would result in a disruption of relict sedimentary structures, which are actually well-preserved.

The following calculations are based on the assumption that these rocks are examples of the open system case discussed above. Some dissolved components totally migrated out of the rock, and some precipitated within the cleavage zones themselves. However, material was neither added nor removed from the microlithons. This may not be strictly true; however, for reasons outlined above, it is thought that any additions must have been minor. In the event that material in fact was transferred from the cleavage zones to the bedding zones, the net changes would be less than the maximum values calculated here.

Column 3 in Table 2 shows apparent enrichments and depletions of components in the cleavage zones relative to the microlithons. SiO_2 and Na_2O show an apparent depletion in the cleavage zones; all other components show an apparent enrichment, with TiO_2 showing the greatest apparent enrichment. It is unlikely that TiO_2 actually was added to the cleavage zones, and previous workers have demonstrated that it is immobile under a wide range of conditions (e.g. Glasson and Keays, 1978). Thus TiO_2 is assumed to have been immobile, and its apparent enrichment is attributed to passive concentration upon removal of more mobile constituents. Assuming immobility of TiO_2, net changes for the other constituents were calculated. These changes are given in column 4 of Table 2, and percent changes are given in column 5. Marked depletion of all components except TiO_2 occurred in the cleavage zones. This is consistent with the petrographic observation that quartz and feldspar, abundant phases in the microlithons, are lacking in the cleavage zones.

DISCUSSION

The petrographic and chemical observations cited above support a solution-precipitation model for the formation of the cleavage zones. The textures in these zones appear to have resulted from three processes: 1) The dissolution of quartz, feldspar, and to a lesser extent, chlorite and white mica resulted in the concentration of the remaining phyllosilicates, iron oxides, and iron-titanium oxides. This dissolution is the process of pressure solution, in which under a differential stress, there is an enhanced tendency for material from the higher stressed sides of grains to dissolve in the thin film of water found along grain boundaries (Rutter, 1976, 1978; Schmid, 1983). The dissolution is evidenced by the lack of quartz and feldspar in the cleavage zones, by the nearly equant phyllosilicate grains which abut the brown zones parallel to the rock cleavage, and by the marked depletion of all components except TiO_2 in the cleavage zones. 2) Although some of this dissolved material was removed from the rock, some remained to form new flake-like white mica grains and chlorite and chlorite-white mica grains which are elongate perpendicular to their (001) planes. This is evidenced by the different habits of the phyllosilicate grains in the cleavage zones compared to the microlithons. The blocky shape of the grains in the cleavage zones results from a combination of shortening perpendicular to the trace of the cleavage zone caused by solution and grain growth parallel to the trace of the cleavage zone caused by precipitation. 3) Recrystallization of preexisting micas occurred in both the cleavage zones and microlithons as a response to the temperatures and pressures of regional metamorphism. Evidence for this is the absence of any bent or deformed grains.

Evidence for any rotation of grains may have been masked by the metamorphic recrystallization. Certainly bedding has been transposed toward the cleavage direction near the cleavage zones (Fig. 1), and there must have been collapse of sedimentary structures upon dissolution and removal of large volumes of material from the cleavage zones. Additional disruption would have occurred if unequal volumes of material were removed from adjacent pelitic and psammitic layers, as seems likely (Steuer and Platt, 1980). This collapse could have caused the rotation of platy phyllosilicate grains into the cleavage direction. However, the existence of relatively coarse, early-formed chlorite crystals lying at high angles to the cleavage zones and which were affected by the cleavage-forming processes, argues against rotation as a significant factor in these rocks.

Several other workers (e.g. Bell, 1978; Woodland, 1982) have invoked a combination of dissolution and new grain growth to explain the formation of discrete cleavage zones. In particular, the works of Glasson and Keays (1978) and Stephens et al. (1979) are relevant to this study. They postulate a similar sequence of chemical reactions to explain the development of cleavage zones in a sequence of sandstones, siltstones, and graywackes. Marlowe and Etheridge (1977) also have called on a combination of modification of existing grains and crystallization of new ones to explain the differentiation into quartz-rich and mica-rich domains which define a crenulation cleavage.

The cleavage zones described in this study differ in the degree of phyllosilicate preferred orientation from many of the spaced cleavages found elsewhere. The white mica grains in the cleavage zones described here have a tendency to lie with their (001) planes parallel to the rock cleavage, but white mica also lies in off-cleavage orientations. This is in contrast to cleavage zones in, for example, the Martinsburg Formation of Pennsylvania. Transmission electron microscopy confirms that the white mica grains in the cleavage zones of the Martinsburg Formation have a strong preferred orientation parallel to the rock cleavage (Lincoln, 1980, 1985). In other respects, these two cleavages are similar and are believed to have formed by the same processes. These processes are driven by stress and by such chemical factors as partial pressure of fluids and temperature gradients (Marlowe and Etheridge, 1977; Stephens et al., 1979; Piqué, 1982). It is possible that in these samples from the Vassalboro Formation the chemical factors played a greater role, while in the Martinsburg Formation the stress was more important, resulting in the differences in preferred orientation.

The original composition of the rock influenced the development of the cleavage. As noted earlier, the cleavage zones in the psammitic layers are farther apart than those in the pelitic layers, although individual zones are wider in the psammitic layers than are their counterparts in the pelitic layers. Robin (1979) states that micas promote silica diffusion. This is in accord with the above observation that the cleavage zones are better developed in mica-rich than in mica-poor layers. It is possible that the cleavage zones began to develop first in portions of the pelitic layers with higher concentrations of white mica.

CONCLUSIONS

The textures in this suite from the Vassalboro Formation provide evidence bearing on the origin of spaced cleavage. The cleavage zones contain a different mineral assemblage than do the intervening zones. The cleavage zones consist of phyllosilicate grains and opaque material and lack quartz, which is an important constituent of these rocks in their uncleaved state. It is assumed in this study that the microlithons are close in composition to the whole rock prior to cleavage development. The process of cleavage development created the differences between the cleavage zones and the microlithons.

Textures observed in this study provide evidence bearing on the nature of this process. The lack of quartz in the cleavage zones and the presence of truncated white mica and chlorite grains evidence the complete or partial dissolution of these minerals. Dissolution occurred along planes parallel to the cleavage zones, and also perpendicular to the maximum shortening direction in these rocks. As a consequence of this dissolution, relatively insoluble material, such as iron oxides, was concentrated.

Along with dissolution, crystallization of new grains occurred.

The shapes of the chlorite and chlorite-white mica grains in the cleavage zones suggest that these grains underwent what might be termed classic pressure solution, dissolving along planes parallel to the cleavage zones and growing along planes perpendicular to these zones. There was also the growth of some new platy white mica grains. Studies of similar zones in other rocks have shown this new white mica to have a different composition than the preexisting white mica (e.g. Stephens et al., 1979), whereas studies of still other rocks have shown no difference in composition between the two populations (e.g. Gray, 1977). Accompanying this crystallization was metamorphic recrystallization due to the temperatures and pressures of regional metamorphism.

The cleavage is better developed and more penetrative in the more pelitic portions of the samples than in the more psammitic. This suggests that original composition is important in cleavage development.

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Seismic Structure of the Earth's Crust Underlying the State of Maine

Alan L. Kafka
John E. Ebel
Weston Observatory
Department of Geology and Geophysics
Boston College
Weston, Massachusetts 02193

ABSTRACT

As part of the Maine seismic refraction experiment conducted in 1984, the U.S. Geological Survey (USGS) detonated a number of explosions in Maine and southern Quebec. In addition to the recordings made by the USGS, these explosions were also recorded by stations of the New England Seismic Network operated by Weston Observatory and by portable instruments installed by Observatory personnel at various locations in northern New England. The seismic structure of the upper crust was studied using an analysis of group velocity dispersion of Rg waves, a tomographic time-term analysis of Pg waves, and an analysis of velocities of direct P waves. Crustal thickness estimates were made using PmP phases. Three regions of differing upper crustal seismic structure were found: (1) a region between the Maine-Quebec border and the central Maine gravity gradient; (2) a region lying to the southeast of the central Maine gravity gradient, extending to and including the area around Penobscot Bay; and (3) the coastal region east of Penobscot Bay. The first region is characterized by mostly small magnitude time-term residuals scattered around zero and by some seismic anisotropy in the northwest. The second region shows strong anisotropy in both the surface waves and the body waves, and generally has larger, positive time-term residuals. The third region has faster upper crustal velocities than those observed in the second region, variable time-term residuals of both positive and negative values, and no observed anisotropy. The thickness of the crust underlying Maine shows some variability, but generally tends to increase from about 33 km along the eastern coast to about 38 km in the northwestern part of the study area. The lateral variation in upper crustal structure delineated by this seismic analysis may elucidate the three-dimensional configuration of lithotectonic terrane boundaries in Maine.

INTRODUCTION

The scale of lateral and vertical variations in earth structure that can be resolved using seismic methods depends upon the wavelength of the seismic waves. Shorter-wavelength energy can be used to resolve smaller features. Determining the details of earth structure using seismic waves requires a high density of sources and receivers as well as the use of the shortest wavelengths (highest frequencies) possible. Where short-wavelength energy is available but dense source/receiver sampling is not, more general information on crustal structure can be achieved with proper data analysis.

The analysis of travel times of body waves and of dispersion of surface waves has classically been used to study the variation of the seismic velocities in the earth as a function of depth (e.g. Bullen, 1963). Recently, experiments utilizing large numbers of sources and receivers in body-wave and surface-wave experiments have led to improved models of the seismic structure of the earth's crust with resolution of lateral as well as vertical velocity variations (e.g. McMechan and Mooney, 1980; Nakanishi, 1985; Suetsugu and Nakanishi, 1985). One important seismic method which is used to investigate the structure of the earth is called tomography (Fawcett and Clayton, 1984). The tomographic method involves using data from a number of crisscrossing paths of seismic waves in a given region to delimit lateral and vertical variations in seismic velocity. Since

tomography is a ray method, it is used with data where the typical wavelength is short compared to the size of the structures being studied. It can be applied to both surface waves (e.g. Nakanishi and Anderson, 1983; Woodhouse and Dziewonski, 1984) and body waves (e.g. Hearn and Clayton, 1986a, 1986b). In this paper, we discuss the analysis of short-period surface-waves ($1.6 \geq T \geq 0.4$ sec) and body-waves ($1.0 > T > 0.1$ sec) recorded in Maine. From the data analysis presented in this paper, we are able to discern variations in the structure of the earth's crust underlying Maine.

In 1984 a seismic refraction experiment was conducted in Maine by the United States Geological Survey (USGS), and this experiment included a large number of explosion sources with known origin times and locations (Murphy and Luetgert, 1986, in press). This refraction experiment was part of a major geophysical investigation of the northern Appalachians in Maine using seismic reflection, seismic refraction, gravity, and magnetic methods (Stewart et al., 1986). Forty-eight refraction blasts were detonated for this experiment. These shots ranged in size from 1600 to 4000 pounds, and they were located at various sites in Maine and southern Quebec. In addition to the recordings made by the USGS, the shots were also recorded by the regional New England Seismic Network operated by Weston Observatory, and by portable stations installed by Observatory personnel (Fig. 1). The Weston Observatory data set complements the USGS data set because it includes many ray paths not sampled by the USGS data set. In this paper, we analyze the Weston Observatory data set recorded from the USGS experiment.

Prior to the 1984 experiment, relatively little work had been done on analyzing the details of the seismic structure of the crust beneath Maine. The crustal structure of southeastern Maine had been studied by Suzuki (1964) using data from the 1961 Gulf of Maine seismic refraction experiment. A crustal model for nearby central New Hampshire had been determined by Taylor and Toksoz (1979). Eichorn (1980) used quarry blasts and portable seismic stations in central and northern Maine to investigate the crustal structure in that part of the state. The crustal model of Chiburis and Ahner (1980), determined for southern New England, has been used successfully at Weston Observatory to locate earthquakes in northern New England, including Maine (Ebel, 1986). Several of the earlier models of the New England crust are shown in Table 1. The models determined from these earlier studies only give information on the average vertical seismic structure of the area and do not elucidate details of lateral variations in velocity or layer thickness. Thus, the 1984 experiment provided an opportunity to greatly expand the knowledge of crustal structure beneath the state of Maine. While a full reduction of the USGS data set has not yet been completed, some results from the refraction and the reflection experiments have already been published (Doll et al., 1986; Green et al., 1986; Klemperer and Luetgert, 1987; Luetgert, 1985a, 1985b; Luetgert and Bottcher, 1987; Luetgert et al., 1986, 1987; Mann and Luetgert, 1985; Spencer et al., 1987; Unger et al., 1987). These results will be discussed below in conjunction with the results of our analysis of the Weston Observatory data set.

METHODS

The Rg Method

The dispersive properties of surface waves can be used to map lateral heterogeneity in the structure of the earth's crust. Kafka and Reiter (1987) analyzed short-period Rayleigh waves (Rg) recorded from the USGS refraction blasts detonated in southern and central Maine. The method used in that study was also used by Kafka and Dollin (1985) to analyze Rg waves recorded from quarry blasts in southern New England, and it is a short-period analogue of methods used to study lateral variation of dispersion of long-period surface waves propagating in the crust and mantle (e.g. Santo, 1965; Nakanishi and Anderson, 1983). These dispersion methods involve measuring the phase and/or group velocity of surface waves propagating over numerous paths from source to receiver or from station to station. The variation in group or phase velocity dispersion from one path to the next implies a variation in earth structure.

The depth of structures resolved by fundamental mode surface waves is directly related to the periods (and hence wavelengths) observed. The USGS blasts studied by Kafka and Reiter (1987) were all 2000 lb blasts, and they did not generate Rg signals beyond about 1.3 sec period in most cases, although occasionally Rg appeared to be present out to almost 1.6 sec. By comparison, the seismograms of quarry blasts in southern New England studied by Kafka and Dollin (1985) and Kafka and McTigue (1985) have Rg signals at periods as great as 2.5 sec with sufficient energy present that group velocity could be measured. The longer-period Rg waves recorded from the quarry blasts are sensitive to variations in crustal structure to depths of about 2.0 to 3.0 km. Because of the more limited period range observed from the Maine refraction blasts, the Rg data discussed

TABLE 1: EXAMPLES OF CRUSTAL MODELS FOR THE NORTHEASTERN UNITED STATES

V_p (km/sec)	Depth to top (km)	Thickness (km)	Region	Reference
5.31	0.00	0.88	Connecticut	Chiburis and Ahner (1980)
6.06	0.88	12.21		
6.59	13.09	21.51		
8.10	34.60			
6.03	0.00	12.0	Maine	Eichorn (1980)
6.60	12.00	21.0		
6.73	34.00	5.0		
7.20	39.00			
5.70	0.00	7.3	Central New Hampshire	Taylor and Toksoz (1979)
6.30	7.30	18.8		
7.30	26.10	12.9		
8.13	39.00			

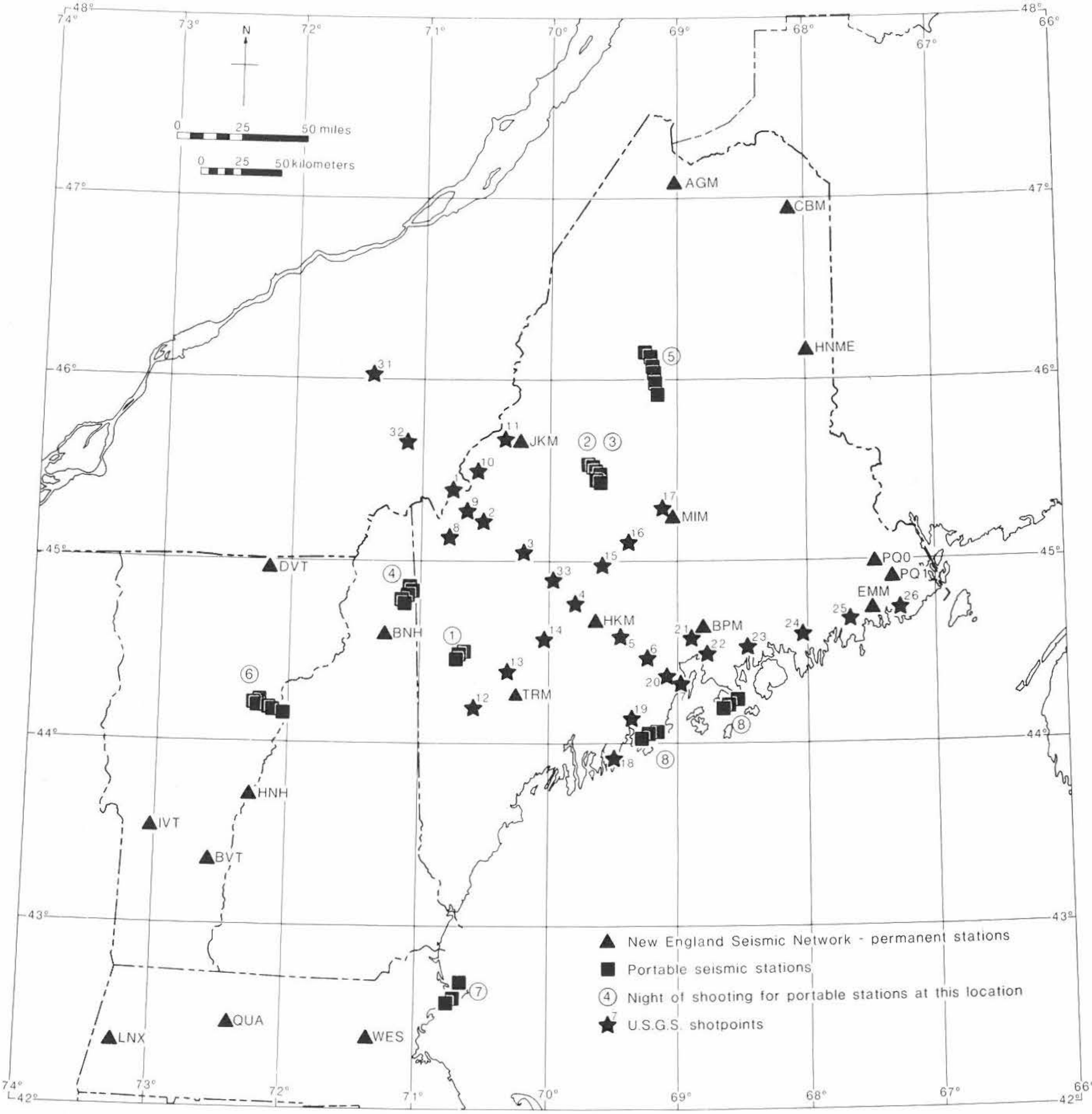

Figure 1. Map of northeastern United States showing stations and shotpoints from this study. Permanent stations of the New England Seismic Network are shown as solid triangles; stars indicate locations of USGS shotpoints (uncircled numbers); and solid squares indicate locations of portable seismic stations deployed for this study. Numbers in circles indicate the night of shooting when portable stations were deployed at a particular location.

here are not very sensitive to variations in crustal structure deeper than about 1.5 to 2.0 km. Within that very shallow portion of the crust, however, Rg dispersion studies reveal details of the seismic velocity structure. Thus, these Rg studies complement the body-wave studies discussed here, since the body-wave studies yield models of deeper portions of the crust.

The Tomographic Time-term Method

Travel times from the first arriving P waves recorded from the USGS refraction blasts were analyzed using the method of Hearn and Clayton (1986a). This method, which we refer to as the tomographic time-term method, is illustrated in Figure

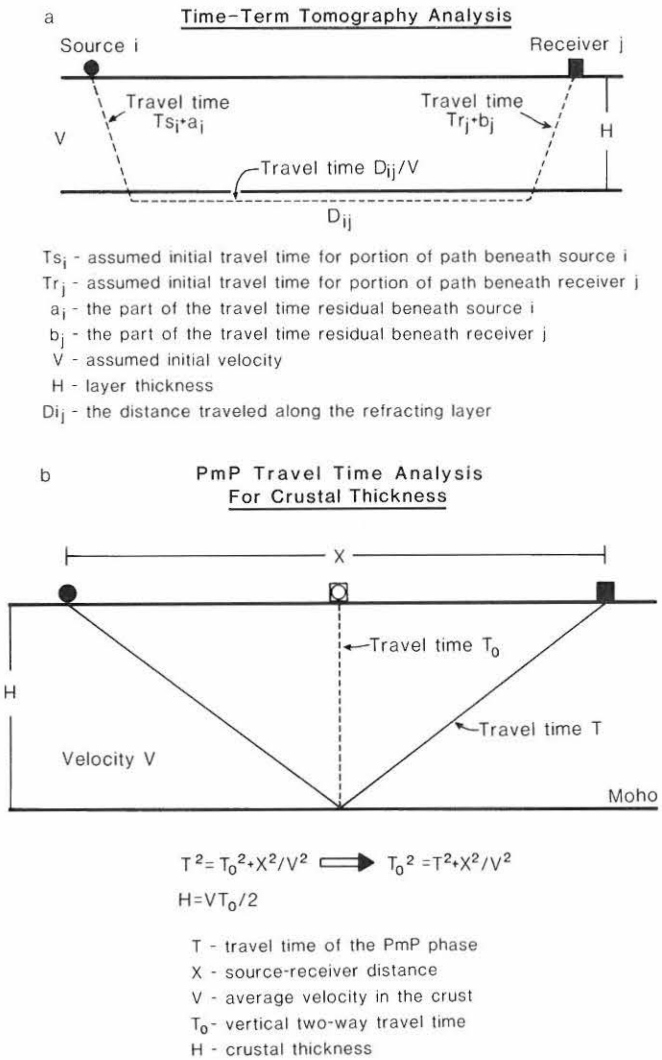

Figure 2. Schematic diagrams of (a) time-term tomography analysis and (b) PmP travel-time analysis.

2a. The travel times are measured for all first arriving P waves which refract as head waves along the top of a particular layer in the crust. The expected travel time calculated from a preliminary layered velocity model of the crust is then subtracted from the observed travel time for each ray path, yielding a travel-time residual. Negative and positive residuals are early and late arrivals, respectively, relative to the preliminary crustal model. The absolute values of the time-term residuals are a function of the initial velocity model used in the analysis.

The refracting layer is divided into a number of blocks, and for each path the resulting travel-time residual is distributed into the crust above the refracting layer beneath the source, the crust above the refracting layer beneath the receiver, and each refractor block. For a large number of source-receiver paths, this leads to the problem of finding a crustal structure that best fits the average residuals for each source position, receiver position, and refractor block. This is an overdetermined problem that can be solved by the least-squares method. The average source and receiver residuals, also called time-term residuals, can be interpreted as resulting from variations of crustal velocities relative to those of the initial model, variations of the depth to the top of the refracting layer, or a combination of both of these effects. The average residual values for the refractor block can be caused by changes in the velocity of the refracting layer, by topography along the top of the refracting layer, or by both effects together. The method is most effective when there are a large number of observations which sample each source, receiver, and refractor block with a number of different source-receiver geometries.

The highest frequency P waves used for the tomographic time-term analysis were about 5 to 10 Hz. This frequency range corresponds to wavelengths as short as 0.6 km. In principle, objects as small as this shortest wavelength can be resolved, but tomographic methods in general act as spatial smoothing filters (Fawcett and Clayton, 1984), and the practical limit on the lateral resolution depends upon the density of source-receiver pairs. In the vertical direction, the depth resolution depends upon the thickness of the crust above the refracting layer as well as upon the thickness and vertical velocity structure within the refracting layer. A positive vertical velocity gradient with depth in the refracting layer will cause longer travel paths to have deeper bottoming points within that layer, and this will affect both the final refractor block residuals and, to some lesser extent, the final time-term residuals computed with the method. Thus, the time-term residuals reflect seismic velocity and layer thickness variations from the assumed model at depths primarily above the refracting layer, while the average block residuals are mainly due to seismic velocity variations within the refracting layer itself.

The PmP Method

The thickness of the crust at different points in Maine was also sampled using the travel times of PmP waves. The PmP phase is a P wave that is reflected from the Moho discontinuity. In Figure 2b we show how the thickness of the crust is sampled by the PmP wave. For surface sources and receivers, the PmP reflection point is approximately halfway between the source and receiver, and the crustal thickness at that point can be estimated from the PmP travel time and an assumed average crustal velocity (Sheriff and Geldart, 1982; Luetgert et al., 1987). This method is approximate since it assumes that the crust can be modeled as a single layer.

The ability of the PmP method to resolve crustal thickness depends upon the accuracy of the average crustal velocity that was assumed in the initial model, the source-receiver distance, and the accuracy of the measured PmP travel times. In this experiment, the source-receiver distances and the PmP travel times are known quite well (assuming, of course, that the PmP phase has been correctly identified), so the accuracy of the crustal thickness estimates hinge primarily upon the accuracy of the average crustal velocity assumed in the initial model. The er-

ror in the crustal thickness values as a function of a given error in the average crustal velocity increases with increasing source-receiver distance. For example, an error of 0.1 km/sec in the average crustal velocity yields crustal thickness errors of 1.7 km and 4.6 km at source-receiver distances of 100 km and 200 km respectively. An assumed average velocity which is slower than that along the ray path will yield crustal depth estimates which are less than the correct depths.

DATA ANALYSIS

Seismic Stations Used for this Study

The locations of the permanent stations of the New England Seismic Network and the portable seismic stations used in this study are shown in Figure 1 along with the USGS refraction shot locations. Each of the New England Seismic Network stations has a 1-Hz vertical seismometer. The data are transmitted via telephone telemetry to Weston Observatory where they are recorded in both analog and digital form. The digital system has a sampling rate of 50 samples/sec for each station (0.02 sec reading accuracy) and a system displacement response that peaks at approximately 10 Hz. The analog data were recorded on two 16 mm photographic recorders. Each photographic unit records 18 stations, and the photographic records can be read to a precision of 0.05 sec. Absolute time is provided by a satellite receiver connected to both the analog and digital systems. A more complete description of the New England Seismic Network instrumentation is given in Ebel (1985).

The portable instrumentation included five smoked-paper recorders and two three-component digital recorders. The smoked-paper recorders were connected to 1-Hz vertical seismometers and were operated at recording speeds of 240 mm/min or 120 mm/min, yielding reading accuracies of 0.02 or 0.04 sec respectively with proper magnification. The digital units recorded all three components of ground motion from 1-Hz seismometers. The sampling rates for the digital units were set to 100 samples/sec. The clocks of the portable instruments were calibrated against the satellite time standard of the New England Seismic Network just prior to their deployment.

Instrumentation failures somewhat reduced the yield of usable data. During the 8 nights of shooting, the digital recording system failed on the fourth night and one of the photographic recorders failed on the first night of shooting. On a given night, only 6 of the shots could be recorded by the New England Seismic Network digital system due to data storage limitations. Digital data were successfully recorded at the New England Seismic Network stations for 44 of the shots. Of the 48 portable field stations deployed, 30 produced records that could be analyzed.

Analysis of Rg Waves

Since the USGS refraction blasts were located at essentially zero depth, they generated fundamental mode Rayleigh waves (Rg) that were recorded at some of the New England Seismic Network stations. Rg data from two New England Seismic Network stations in central Maine (HKM and MIM) were analyzed by Kafka and Reiter (1987), and their data analysis and results are summarized below.

All traces selected for the Rg analysis were first inspected visually for prominent Rg arrivals, and an amplitude spectrum was calculated for each trace. The highest signal-to-noise ratio for Rg waves recorded from the refraction blasts was generally in the period range of about 0.3 to 1.3 sec. Seismograms that had strong signals in that period range were examined, and for those seismograms the background noise was also analyzed to ensure that it did not overlap severely with the Rg signal. Based on this preliminary screening of the data, twelve paths were chosen for the dispersion analysis.

To illustrate the Rg methodology used by Kafka and Reiter (1987), we show their analysis for one of the better examples of Rg data recorded from the Maine refraction shots (Fig. 3). The upper trace in Figure 3a is the seismogram recorded at station HKM from shotpoint #6, and the lower trace is the same seismogram after being low-pass filtered. The prominent dispersed wave train arriving between about 7 and 12 sec is the Rg wave. Figure 3b shows the amplitude spectrum of the original recorded trace. In that figure, the prominent spectral peak centered at about 2 Hz corresponds to the Rg wave. This peak is characteristic of Rg waves recorded by the New England Seismic Network from surface blasts and very shallow earthquakes.

Group velocities of Rg waves from the Maine refraction blasts studied by Kafka and Reiter (1987) were measured using the technique of narrow-band pass filtering (Dziewonski et al., 1969). The result of applying the narrow-band pass filter analysis to the seismogram from shot #6 recorded at station HKM is shown in Figure 3c. The seismogram is Fourier transformed and narrow-band pass filtered about a series of center frequencies. For each center frequency, the group arrival time is approximated by measuring the arrival time of the maximum amplitude of the envelope of the filtered signal. This process is repeated for a series of center frequencies, and group velocities are estimated at each center frequency. Group velocities, determined using the narrow-band pass filter analysis, are shown in Figure 4 for all twelve paths analyzed.

The group velocity dispersion curves shown in Figure 4 were inverted by Kafka and Reiter (1987) using a linearized least squares method (Backus and Gilbert, 1970; Der et al., 1970; Franklin, 1972) to determine the velocity structure of the shallow crust underlying the twelve paths analyzed. Similar methods have been used extensively to invert longer-period surface waves for deeper structures (e.g. Mitchell and Herrmann, 1979; Taylor and Toksoz, 1982). Since Rayleigh waves are most sensitive to shear wave velocity, the shallow crustal models determined from this inversion are essentially shear wave velocity models. To compare the Rg inversion results with seismic velocities determined from the body wave analyses, it is necessary to estimate V_p. Kafka and Reiter used a V_p/V_s ratio of 1.78 to convert S-wave velocities to P-wave velocity estimates. Later in the report we discuss the compatibility of the

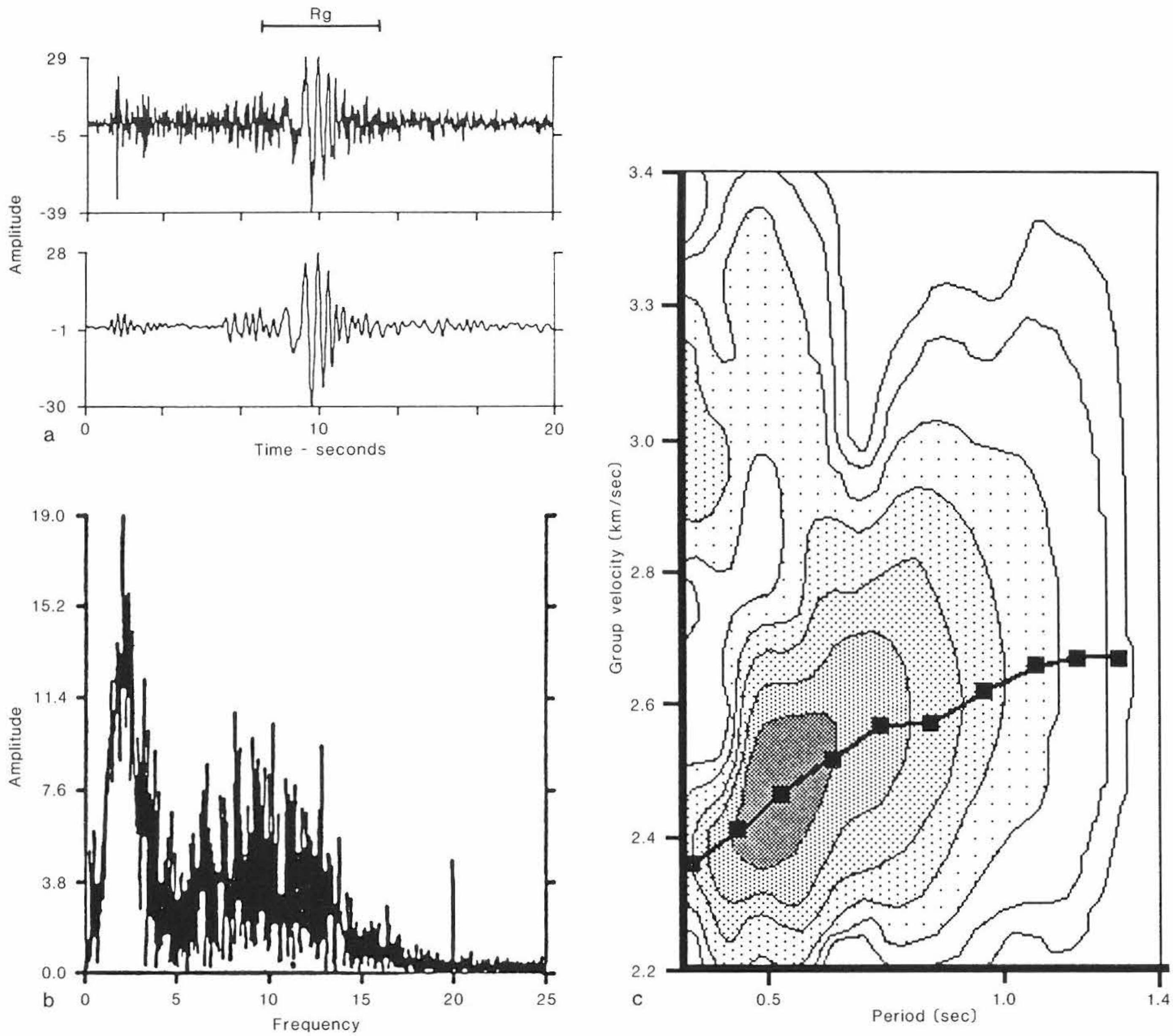

Figure 3. Example of Rg waves recorded from one of the USGS blasts and illustration of Rg methodology. (a) Upper trace is the seismogram recorded at a distance of 39 km from a 2000 lb blast. Below is a low-pass filtered trace with a high-frequency cutoff of 4 Hz. (b) Amplitude spectrum of the raw seismogram (upper trace) shown in (a). The prominent peak at about 2 Hz is characteristic of Rg waves recorded by the New England Seismic Network. (c) Narrow-band pass filter analysis of the seismogram shown in (a). Each contour interval represents a change of 2 db in the amplitude of the filtered signal envelope at a given center frequency.

surface wave and body wave results for a range of V_p/V_s values.

Analysis of Body Waves

Tomographic Time-term Analysis. The tomographic time-term method was applied to two separate data sets read from the New England Seismic Network records: first arrivals representing the Pg phase (a head wave in the upper crust with an apparent velocity of about 6.1 km/sec) and those representing the Pn phase (a head wave in the uppermost mantle with an apparent velocity of about 8.1 km/sec). All first arrival times for source-receiver distances of 30 km to 160 km were assumed to be the Pg phase, and first arrivals at distances beyond 190 km were assumed to be the Pn phase. These distances were chosen based on the calculated crossover distances for the Chiburis and Ahner (1980) crustal model as well as on a preliminary analysis of the USGS refraction observations (J. Luetgert, pers.

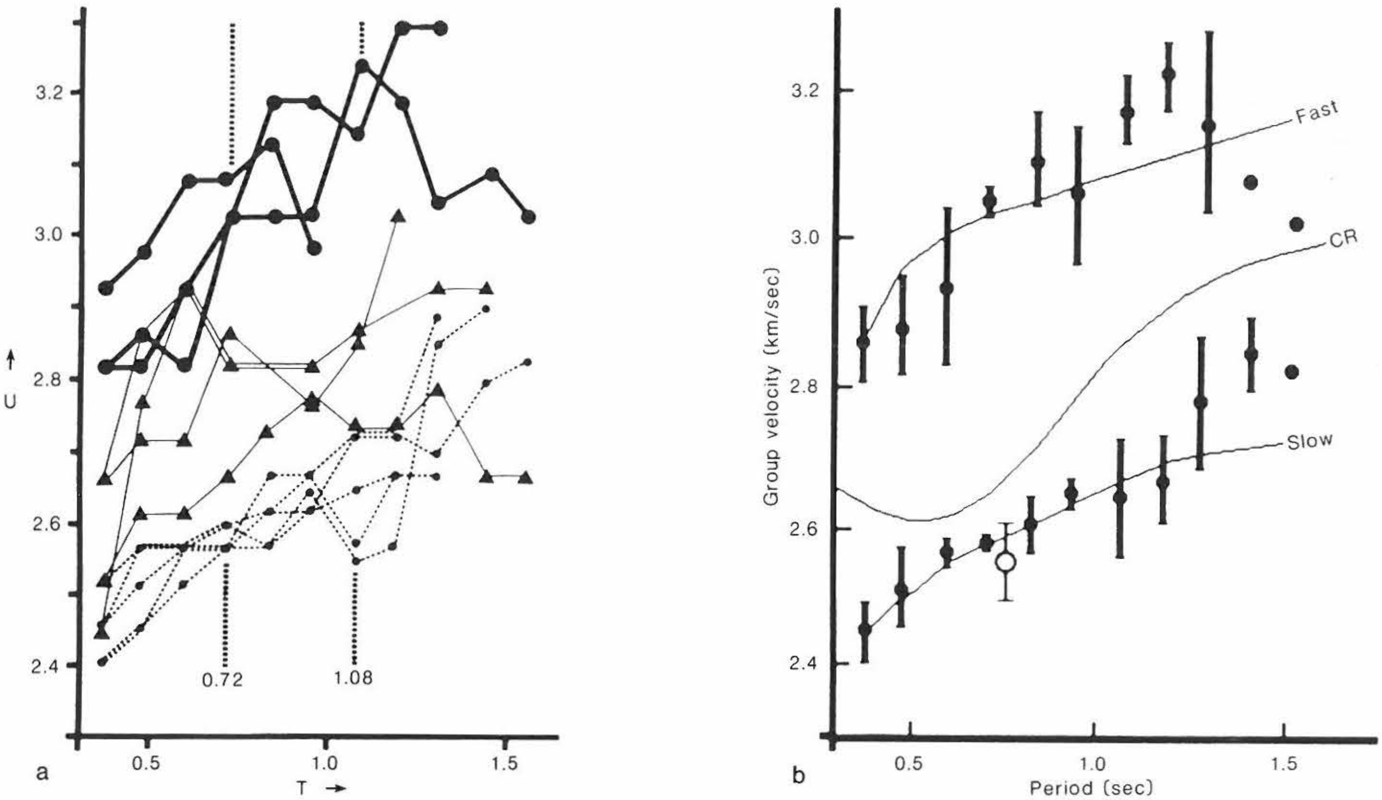

Figure 4. (a) Group velocity data for Rg waves recorded by stations HKM and MIM. Thick solid lines, narrow solid lines and dotted lines represent fast, intermediate and slow clusters respectively. U is group velocity in km/sec, and T is period in sec. (b) Average and standard deviation of Rg group velocities for slow and fast clusters shown in (a). The open circle corresponds to a peak-and-trough measurement for blast #20 to HKM. Also shown are dispersion curves for shallow crustal models of the region surrounding stations HKM and MIM along with the dispersion curve corresponding to the Chiburis and Ahner (1980) refraction model (curve labelled CR). The shallow crustal models corresponding to the theoretical dispersion curves are shown in Figure 6.

commun., 1984). No portable instrument data were used in this particular analysis.

The travel times for all shot-receiver pairs with clear first arrivals were read from both the analog and the digital New England Seismic Network records, and travel-time residuals for each path were then computed relative to the Chiburis and Ahner (1980) model. The set of residuals at each station was then examined. Some shotpoints had been reused as many as 6 times, and for each of these shotpoints the travel-time residuals for different paths could be compared and the most appropriate residual selected. In general, the largest shot at a given shotpoint gave the most reliable travel-time readings. Next, the travel-time residuals from adjacent shotpoints to a given station were examined. Residuals that differed significantly from those corresponding to other nearby shotpoints were discarded. In all cases where it existed, the readings from the digital data were given preference over the analog readings. In the last step, each selected travel-time residual was assigned a weighting factor based on the reading precision and quality and on the shot size. The final data sets for the Pg and Pn analyses had 153 and 57 different travel paths respectively.

The refractor layers were divided into a number of different blocks. The blocks were chosen to be sufficiently large to collect a good sampling of rays, but small enough to resolve some lateral heterogeneity. For the Pg refractor, the blocks were about 52 km by 54 km in lateral extent, while for the Pn refractor they were about 75 km by 78 km. Travel distances for each ray in each block were computed by hand and stored along with the path identification, travel-time residual, and path weight in a computer file. The data sets were then inverted using the least-squares procedure of Hearn and Clayton (1986a) to get the average residuals at each shot and receiver point for each refractor block. The refractor block residuals were converted to refractor velocity perturbations (Hearn and Clayton, 1986a), and new velocitites were found for each refractor block.

PmP Analysis. For the PmP crustal thickness analysis, the relatively wide spacing between individual receivers as well as between individual shotpoints made the recognition of the true PmP phase on most records difficult. However, Nutting (1984) showed that, for the Chiburis and Ahner (1980) crustal model, the PmP phase has a very large amplitude beyond 90 km (a distance where it becomes post-critical). Between about 90 km and

120 km, PmP is noticeably separated in time from other major crustal reflections and refractions. From 120 km to about 200 km, PmP arrives very close in time to other P phases that reflect and refract in the middle of the crust. Beyond 200 km, PmP again becomes separated in time and easier to identify. All of the records (New England Seismic Network and portable) were scanned for possible PmP phases, with special attention being paid to the distance range of 90 km to 120 km and to distances beyond 200 km. All phases with amplitudes noticeably above the surrounding signal and with travel times near that expected for PmP were selected and read. Only 39 readings out of 129 seismograms were considered reliable. Crustal thickness values were then computed from the final set of PmP travel times, and the resulting value for each path was assigned to a point midway between the source and receiver.

RESULTS

Rg Wave Results

In the period range analyzed (0.4 to 1.6 sec), Kafka and Reiter (1987) observed Rg group velocities ranging from about 2.4 to 3.3 km/sec (Fig. 4). As in other studies of Rg dispersion, they observed normal dispersion at shorter periods. This normal dispersion indicates that seismic velocities are relatively low in the upper 0.5 to 1.0 km of the crust beneath the region surrounding stations HKM and MIM. The results of their study indicate that in the region investigated, the average V_p increases from about 4.9 km/sec very near the surface, to about 6.2 km/sec at a depth of about 2 km. The total range of observed group velocities (about ± 0.3 km/sec) is similar to that observed by Kafka and Dollin (1985) for Rg waves in southern New England (about ± 0.4 km/sec). The Rg dispersion results shown in Figure 4 were divided into three clusters based on the velocities observed in the period range of 0.72 to 1.08 sec. These three clusters are characterized by *fast, intermediate,* and *slow* group velocities.

Figure 5a shows the paths of Rg waves from Kafka and Reiter's study superposed on a geologic map of southeastern Maine. A simplified model of the structural grain in that region is shown in Figure 5b where the arrows indicate the orientation of the structural grain. Because the grain changes orientation across the region, our simplified model consists of two subregions with different orientations. The observed group velocities appear to depend primarily on the azimuthal orientation of the path relative to the trend of the structural grain. Paths that are transverse to the grain tend to be in the slow cluster, and the three paths that are approximately parallel to the grain are in the fast cluster. Also, paths that are in the intermediate cluster tend to be oblique to the grain. Because of the distribution of paths sampled and the complexity of the geology, it is difficult to unambiguously distinguish between this apparent lateral anisotropy and lateral inhomogeneity in the shallow crust beneath this region. Nonetheless, lateral anisotropy in the shallow crust provides a simple explanation of the observed pattern of Rg dispersion.

The Rg inversion results of Kafka and Reiter (1987) are shown in Figure 6. The average shallow crustal model in Figure 6a is the result of inverting the mean of Rg group velocities for all paths and all periods analyzed by Kafka and Reiter (1987). The data corresponding to that model are shown in Figure 6b. The shallow crustal model resulting from the inversion of the Rg data from the slow cluster is also shown in Figure 6a. Using the V_p/V_s ratio of 1.78 (which was assumed by Kafka and Reiter, 1987), the upper 1.2 km of the slow cluster model is characterized by V_p ranging from 4.8 km/sec very near the surface to 5.4 km/sec at 1.2 km depth. Below 1.2 km, the slow cluster model is characterized by V_p of 5.6 km/sec. In Figure 6c, we show how variation in the assumed V_p/V_s ratio affects the V_p estimates, and we compare those estimates with the reflection and refraction results of Klemperer and Luetgert (1987) for essentially the same path (shotpoints 4, 5, 6, 20, and 7).

The data quality, range of periods observed, and number of paths for the intermediate and fast group velocity clusters were insufficient for a formal inversion. It is clear, however, that to match the group velocities observed for the fast paths, much higher seismic velocities are necessary in the upper 1 to 2 km. Average values for the fast paths are shown in Figure 4b, and a model corresponding to the fast dispersion curves is shown in Figure 6a. The "fast" model shown in Figure 6a was obtained by increasing the velocities in the layers of the "average" model until the calculated dispersion curve approximated the group velocities observed for the fast cluster.

Body Wave Results

Results of Time-term Analysis. The tomographic time-term analysis was applied to the Pg data a number of times, each with a different assumption about the station weights and/or with all refractor block velocities frozen at 6.06 km/sec (the upper crustal velocity of the Chiburis and Ahner (1980) model, which was the only model tested). In these runs, the relative pattern of the time-term residuals was stable, although there was variation in the absolute time-term residuals. Because of the timing limitations of the recording system, time-term variations on the order of 0.05 sec or less are considered unresolvable. For many of the shot or receiver locations, the variations in the time-term residuals among all the runs was 0.05 sec or less. For some of the locations with the largest residuals (either positive or negative) the variations of the time-term residuals calculated among the different runs were as large as 0.4 sec. The run with the originally assigned weights and with the variable refractor block velocities was considered the most reliable. All time-term residuals from that run were near the mean of those values found for all the different test assumptions. It should be noted that this tomographic time-term analysis does not take into account any anisotropy in the upper crust. At those shot and receiver points where significant anisotropy may exist (see

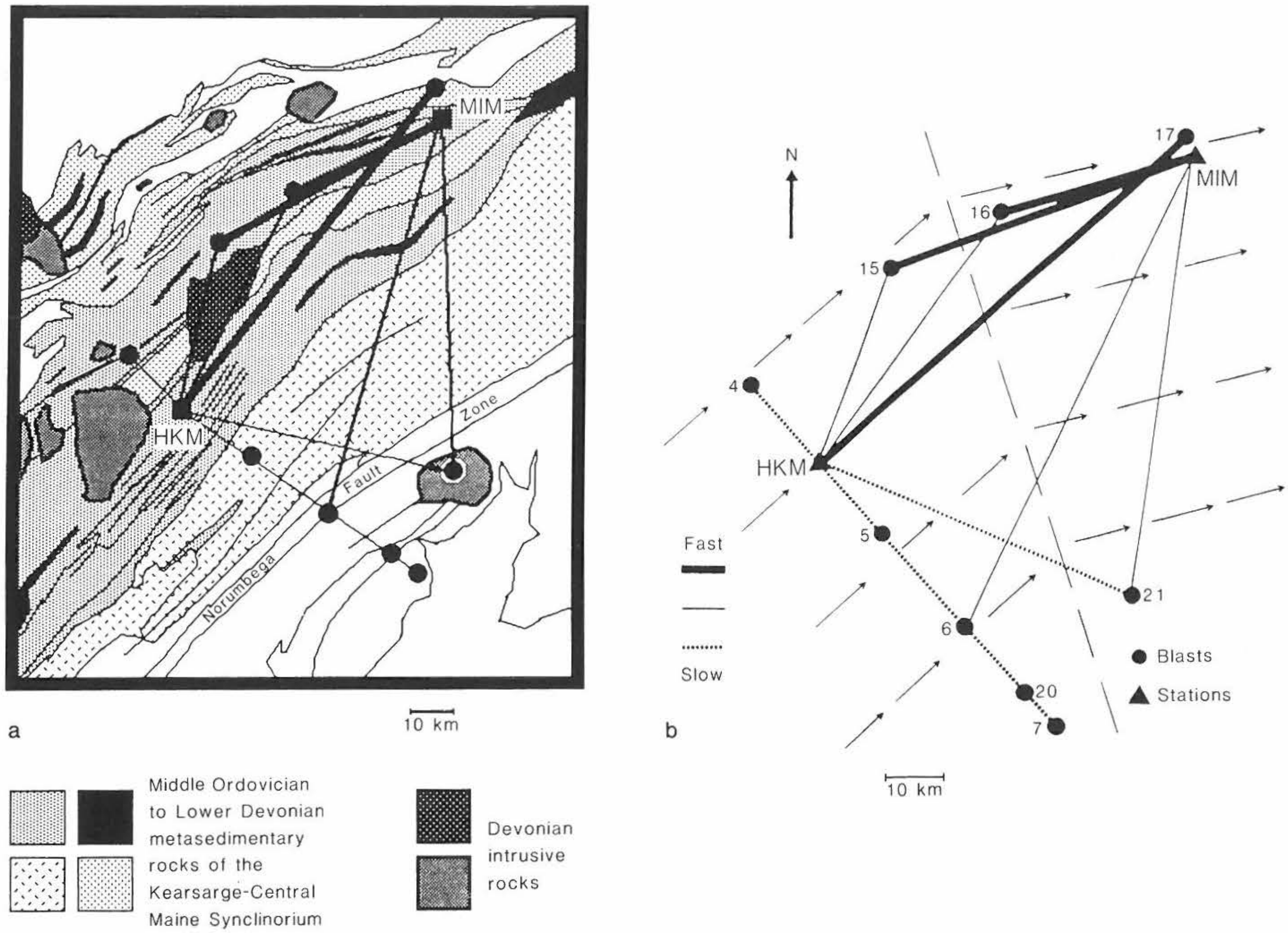

Figure 5. (a) Paths of Rg waves discussed in the text, superposed on a map of geologic structures adapted from the Bedrock Geologic Map of Maine (Osberg et al., 1985). The Norumbega fault zone is labeled NFZ. Closed circles represent refraction blasts, and closed squares represent stations MIM and HKM. (b) Simplified model of the pattern of Rg wave group velocities and the structural grain in the vicinity of stations HKM and MIM. Thick solid lines, narrow solid lines, and dotted lines represent fast, intermediate, and slow paths respectively. For these paths Rg group velocities range from 2.6 to 3.2 km/sec at periods near 1 sec. Arrows represent a simplified model of the orientation of the structural grain.

above and below as well as Doll et al., 1986; Klemperer and Luetgert, 1987), the time-term residuals represent an average travel-time residual which is weighted by the orientations of the observed ray paths relative to the direction of the fastest and slowest velocities. It is because of the variability and uncertainty of the spatial extent of the anisotropy in Maine that it was not included in the tomographic time-term analysis.

The results of the tomographic time-term analysis for the Pg phase are illustrated in Figure 7. All of the shotpoint and receiver time-term residuals are shown, along with a contoured interpretation of those values. Only two refractor blocks were found to have slowness values resolvably different from that of the initial crustal model, and those two blocks along with their calculated velocities are also indicated in Figure 7. In general, the shotpoints in Canada have negative time-term residuals. For the area of northwestern Maine and northern New Hampshire, the time-term residuals are relatively small, ranging from -0.08 sec to 0.16 sec. Between shotpoint 4 and station HKM there is an abrupt change in the time-term residuals to quite large positive values, ranging up to 0.36 sec. This region of large, positive time-term residuals appears to surround Penobscot Bay, extending from HKM southeast to the coast and from shotpoint 18 to shotpoint 22. In this area, the one resolvably slow refractor block, with a velocity of 6.00 km/sec, was found. To the east, the time-term residuals again become negative in the area where the one resolvably fast block (velocity of 6.13 km/sec) is located.

The scatter in time-term residuals between adjacent points is in some cases quite large. In particular, shotpoint 13 was found to have a time-term residual of -0.41 sec, a much more negative value than that for nearby shotpoints or station TRM. An examination of the data used for shotpoint 13 revealed that, while most of the residuals scatter just above or below 0.0, there was

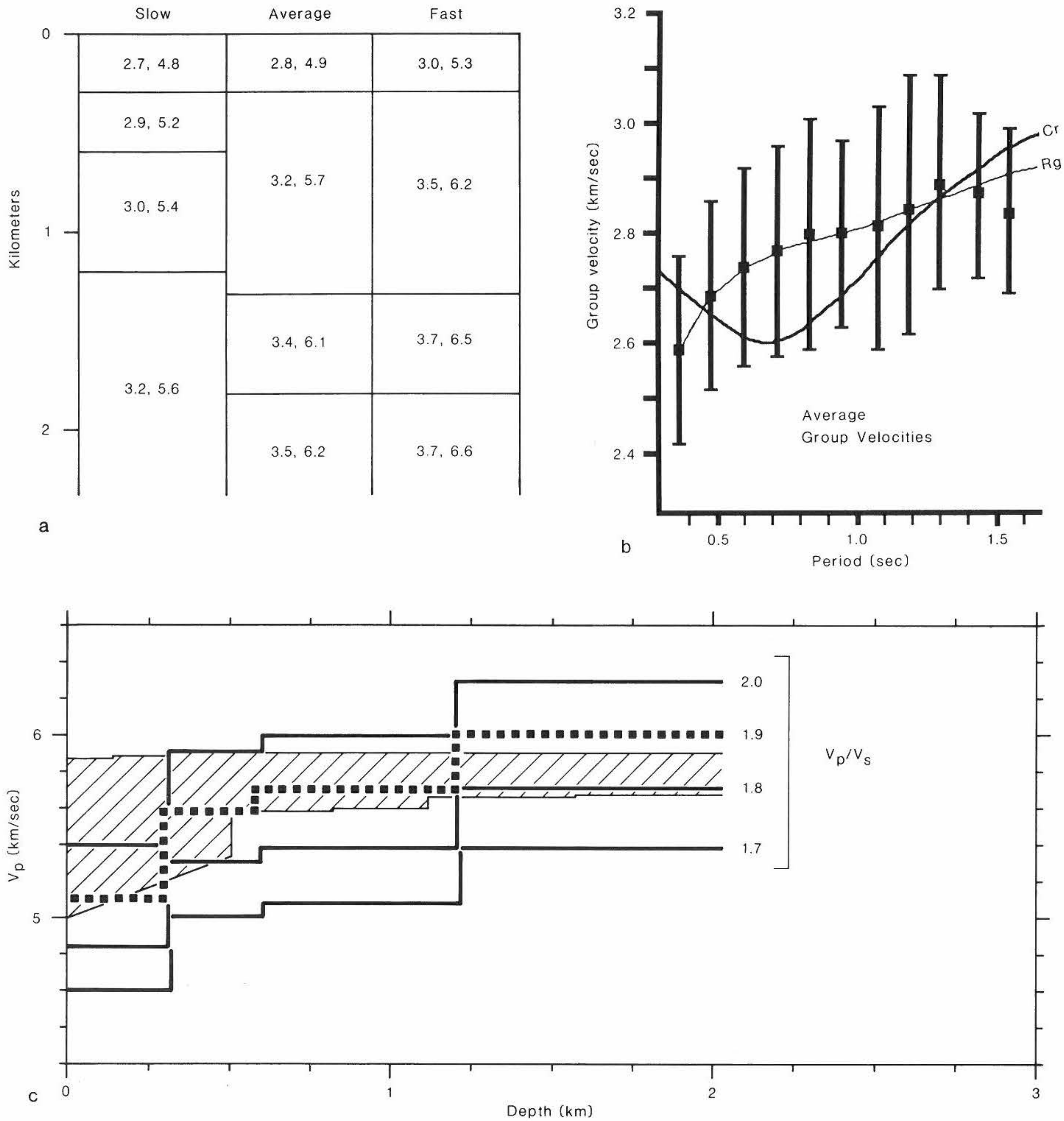

Figure 6. (a) Shallow crustal models determined from Rg dispersion data (from Kafka and Reiter, 1987). Numbers in layers represent V_s and V_p respectively in km/sec. (b) Average and standard deviation for all Rg group velocities shown in Figure 4. Also shown are theoretical dispersion curves corresponding to Chiburis and Ahner (1980) refraction model (CR) and the model resulting from inversion of the Rg data shown here [curve labelled Rg – average model shown in (a)]. (c) Comparison of Rg wave results for the slow paths (solid and dashed lines) with body wave refraction results of Klemperer and Luetgert (1987) (cross-hatched area) for a range of V_p/V_s values.

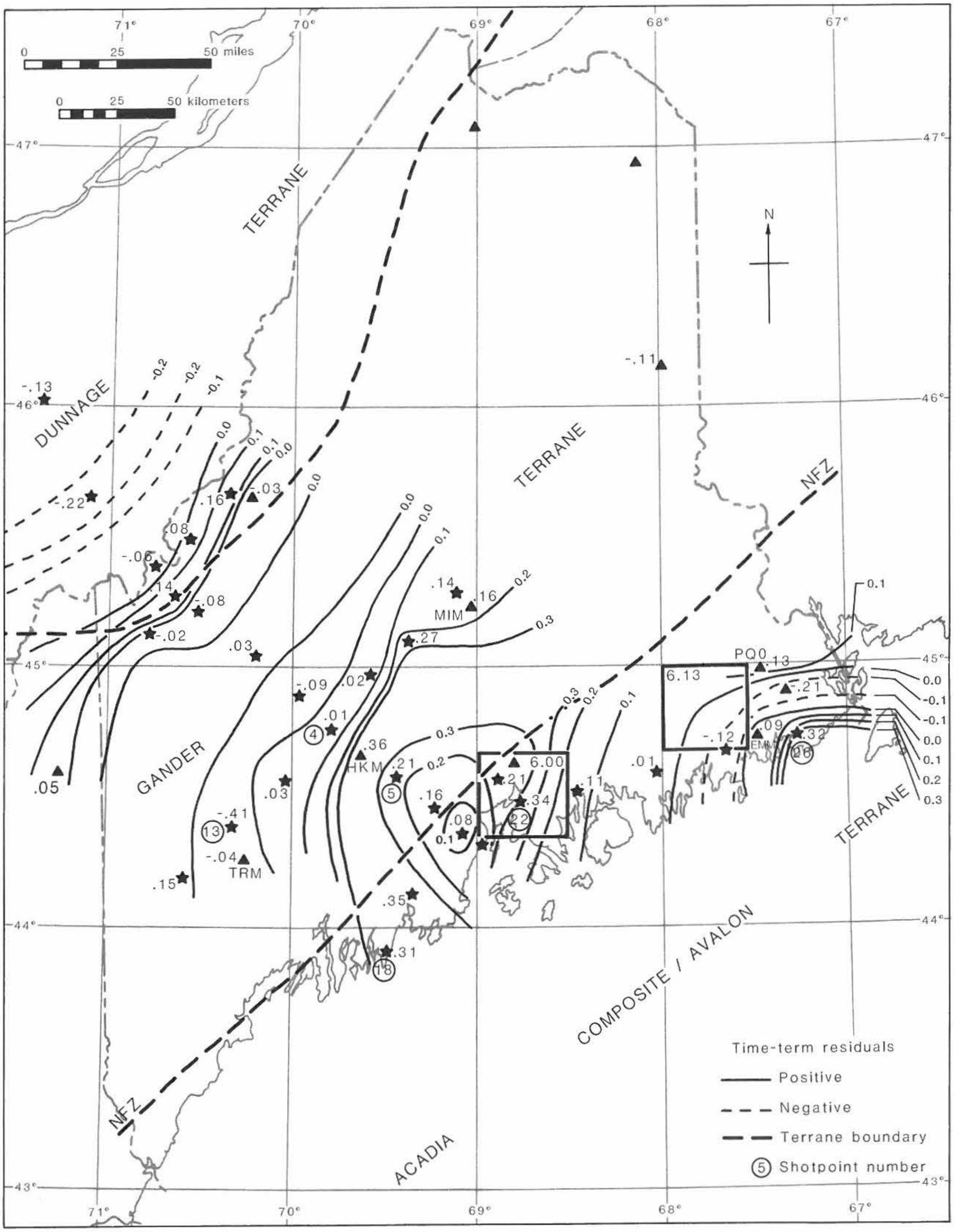

Figure 7. Results of tomographic time-term analysis of Pg waves recorded from the USGS refraction blasts. Numbers next to shotpoints and stations are time-term residuals in sec. Also shown are the boundaries (heavy dashed lines) of the Dunnage, Gander, and Acadia Composite/Avalon terranes discussed in the text. Outlined areas are refractor blocks with resolvably different slowness values.

one travel-time residual on the path to HKM which was very negative (-0.56 sec) and controlled the absolute value of the time-term residual for that point. Shotpoint 26 also appears to have an anomalous time-term residual (0.32 sec) compared to the values found at EMM and PQ0 (0.09 sec and -0.21 sec respectively). However, that time-term residual is consistent with the input data and probably is caused by local structure under the shotpoint, as will be discussed later. The strong change in the time-term residuals between shotpoint 5 and HKM is also considered real from both an examination of the input data and from the consistency of the time-term residuals to the northwest and southeast of this locality.

The application of the tomographic time-term method to the Pn data did not yield reliable results. Time-term residuals ranging from -0.89 sec to 2.04 sec and Pn refractor velocities between 7.61 km/sec and 8.34 km/sec were obtained from the analysis. Adjacent shotpoints or receiver points had radically different time-term residuals in several cases. The low redundancy of the data (most shotpoints and receivers had only 1 or 2 usable data points) was most likely the cause of the problems in this part of the analysis. Even with the poor results of the Pn analysis, however, some of the patterns in the time-terms found in the Pg analysis (for example the region of positive residuals southeast of HKM) appear to have an expression in the Pn results.

Results of PmP Analysis. The PmP analysis yielded a widely-spaced sampling of crustal thickness measurements across the study area (Fig. 8). The same average velocity of 6.4 km/sec was used in a single layer crustal model (Luetgert et al., 1987) for the calculation of all crustal depths shown in Figure 8. It is instructive to compare the results shown in Figure 8 with the results of Luetgert et al. (1987) who were able to use a dense sampling of PmP phases to estimate crustal thickness throughout much of the study area. Along the coast of eastern Maine, a crustal thickness of about 33 km was found in both studies. In northwestern Maine the values found in this study appear to scatter around those reported by Luetgert et al. (1987) for that region. Both studies show a depression in the Moho depth beneath southwestern Maine, although Luetgert et al. (1987) report PmP travel times consistent with a crustal thickness of about 35-36 km beneath that area, whereas thickness values up to 44 km were found in this study. Measurements made in the northern and northeastern part of the area show the greatest discrepancies between the two studies. Luetgert et al. (1987) find thin crust east of shotpoint 17, while Figure 8 shows a quite thick crust there. West of shotpoint 17 we find crustal thickness values around 35 km, while the analysis of Luetgert et al. (1987) is more consistent with 38 km in that area.

Some of the discrepancies between our results and those of Luetgert et al. (1987) can be qualitatively explained. First, most of the crustal thickness estimates that were made from phases recorded at distances of 180 km or more (the underlined values in Fig. 8) yield values which are greater than those expected from the Luetgert et al. (1987) results. Since the PmP phase at this distance range travels a larger percentage of its path in the lower crust than does PmP at distances of 80 km to 120 km, the average crustal velocity necessary to reduce these large distance measurements may need to be greater than 6.4 km/sec. As was noted earlier, small changes in this velocity parameter can affect crustal thickness estimates by several kilometers at distances greater than 200 km. It is also possible that the PmP phase was misidentified at these large distances, since all signals from these blasts had rather low amplitudes on the Weston Observatory records. Second, Luetgert et al. (1987) show that a sizable reflection from a velocity discontinuity in the mid-crust precedes PmP by about 2 sec in the distance range from 89 to 120 km. Thus, some of our crustal thickness values which are less than those expected from the results of Luetgert et al. (1987) may be due to a misidentification of the mid-crustal reflection as PmP. Synthetics calculated by Luetgert et al. (1987) show no significant arrivals within several seconds after the PmP phase in the 89 to 120 km distance range. This makes it problematical to argue that our crustal thickness values that are greater than those of Luetgert et al. (1987) at distances less than 120 km are due to PmP misidentification.

In general, the crustal thickness results from this study support the following principal conclusions of Luetgert et al. (1987): (1) the crust is thinnest under coastal Maine and thickens inland, and (2) there appears to be a depression of the Moho under southwestern Maine. One data point under the White Mountains of central New Hampshire suggests the possibility of thicker crust in that area.

Compatibility of Surface Wave and Body Wave Results

The Rg surface-wave method constrains the seismic structure in very shallow portions of the crust, and the body-wave methods used here generally constrain the structure in deeper parts of the crust. We can test the compatibility of the surface-wave results with the body-wave results by using travel times of the initial P waves to calculate the average velocity for short travel paths which sample shallow portions of the crust. This is done by taking a given shot-receiver distance and dividing that distance by the travel time of the initial P wave. The average velocities for all paths less than about 30 km long are shown in Figure 9. It must be kept in mind that if the velocity is increasing with depth, then the longer the source-receiver distance, the deeper the energy bottoms in the earth. For the paths illustrated in Figure 9, it is likely that the energy bottoms within the upper 2 or 3 km and that the average velocity most closely matches the velocity at and just above the bottoming point of the ray.

Several points can be noted by examining Figure 9. First, the anisotropy pattern noted in the surface wave analysis and by Klemperer and Luetgert (1987) appears to have some expression in the average P wave velocities in the vicinity of stations MIM and HKM. The fast, intermediate, and slow surface wave paths have essentially the same pattern as the P wave average velocities. Second, the average velocities calculated from

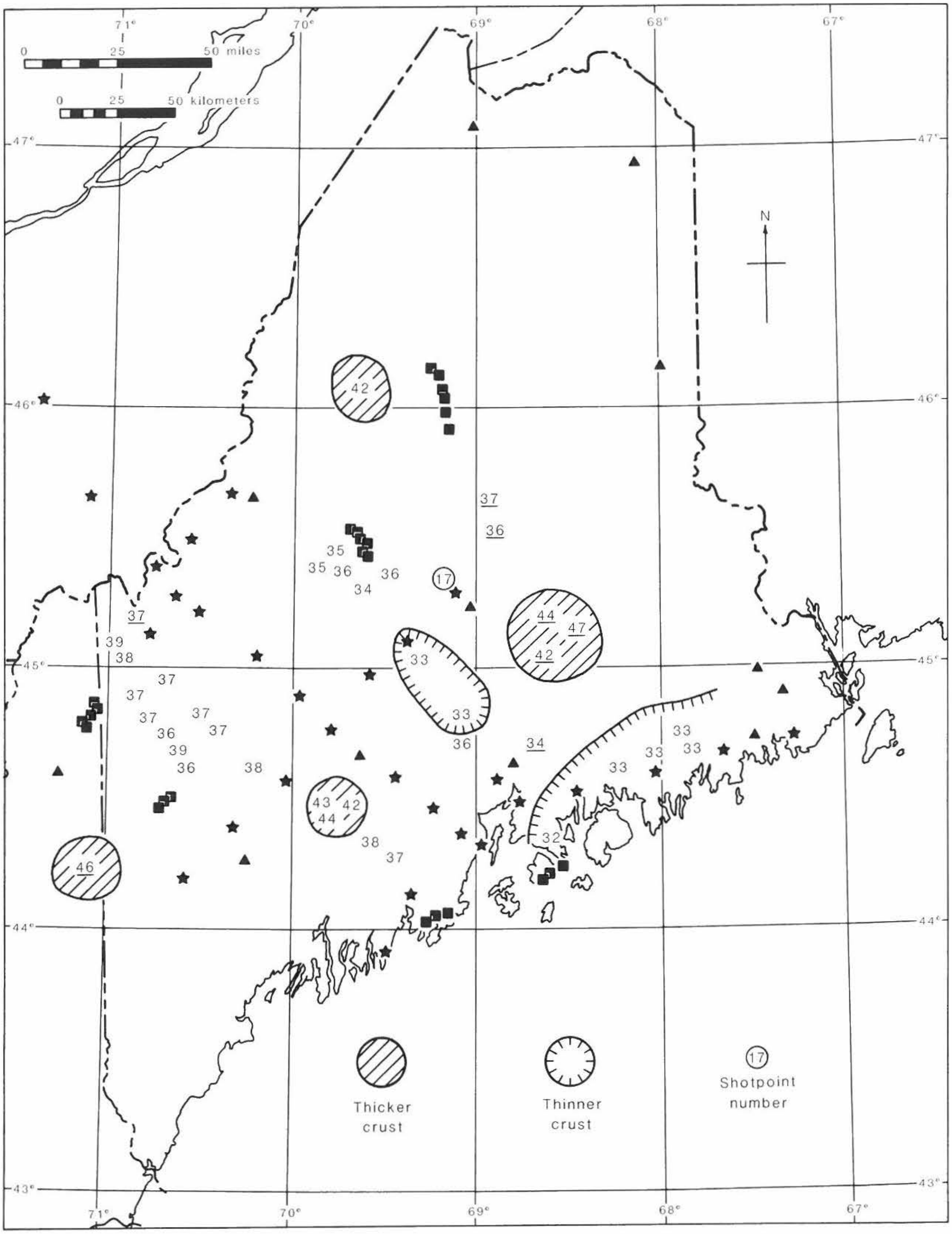

Figure 8. Results of PmP crustal thickness analysis. The crustal thickness estimated for each path was assigned to a point midway between the source and receiver. The underlined values are from shot-receiver distances greater than 180 km.

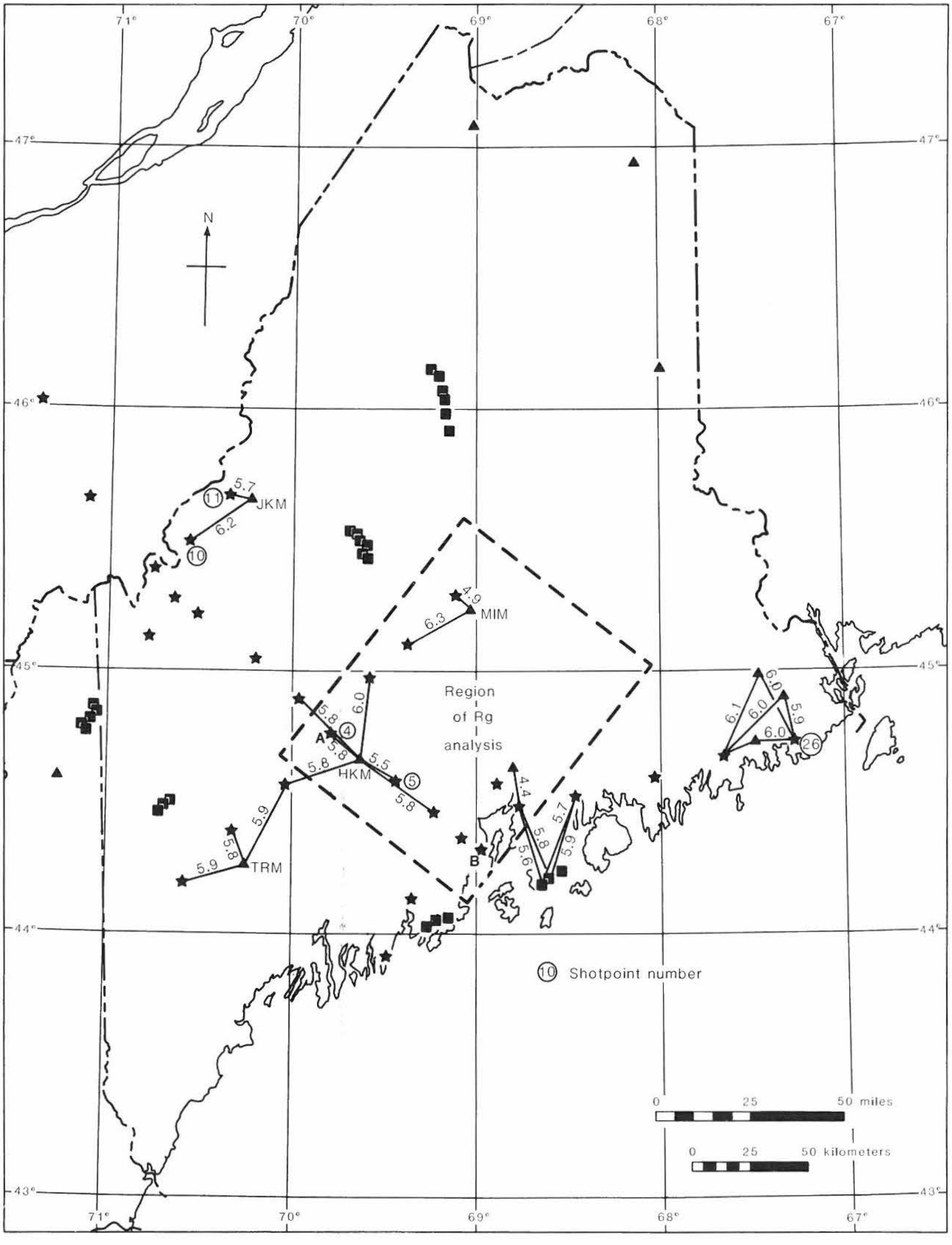

Figure 9a. Average P-wave velocity determined from the travel times of first arrivals of P waves divided into the source-receiver distance for short paths (< about 30 km).

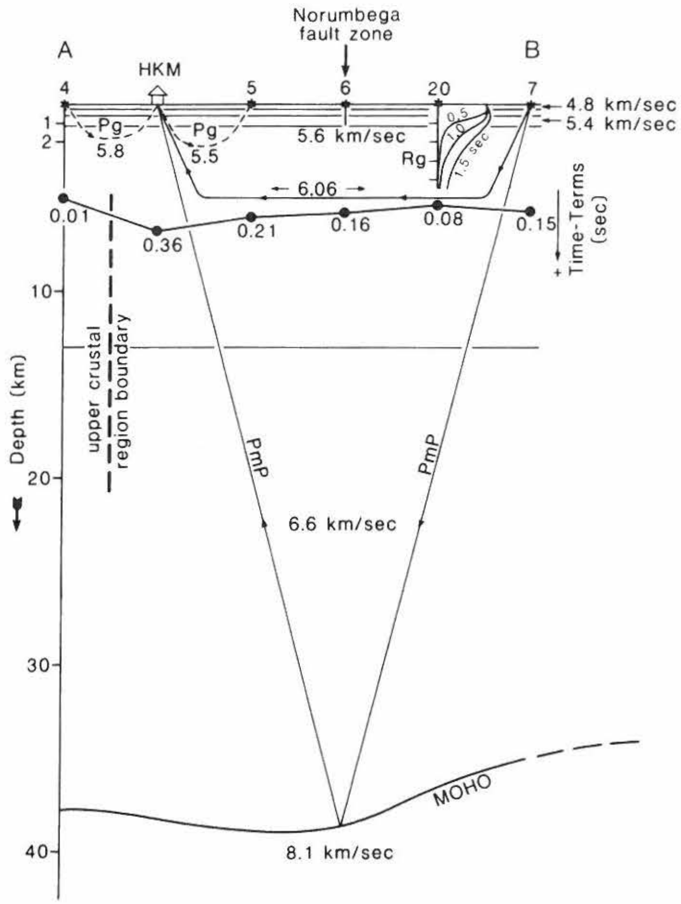

Figure 9b. Schematic diagram of cross-section AB in figure 9a illustrating the relationship between the various wave types and methods used in this study. Vertical exaggeration is approximately 3 to 1. Time-term residuals are plotted in upper crustal layer, and they indicate lower V_p and/or a thicker layer. The curves labelled Rg represent displacement versus depth for Rg waves at periods of 0.5, 1.0 and 1.5 sec.

the P waves generally match well with the velocities found from the surface waves at about the expected bottoming depths of the body waves. The average V_p for one very short path near MIM which bottomed at very shallow depth was 4.9 km/sec. That result compares well with values obtained for the top layers of the slow and average Rg models. The rest of the average V_p measurements indicate a range of 5.5 to 6.3 km/sec for V_p at about 2 or 3 km depth, and the surface wave results range from 5.6 to 6.6 km/sec for V_p at a depth of about 2 km. The assumed V_p/V_s ratio (discussed below) and the assumed bottoming point of the body waves are not accurate enough to conclude whether the differences in these two ranges are significant. Third, anisotropy does not appear to exist near station TRM nor in coastal Maine east of Penobscot Bay. However, Doll et al. (1986) argued that there may be anisotropy in the vicinity of shotpoints 10 and 11, and the fast average velocity parallel to the Appalachians to station JKM could support that argument. Fourth, the velocities in easternmost Maine are, on average, greater than those in any of the areas to the west and northwest. This is consistent with the Pg time-term results, even though different data were used in the two analyses. It is curious that the velocities found for paths from shotpoint 26 are not significantly slower than those for other nearby shotpoints. This suggests that the source of the slow anomaly found in the Pg time-term analysis lies deeper than 2 or 3 km in the crust or is directly beneath the shotpoint. Finally, the average velocity from shotpoint 4 to HKM is significantly faster than that from shotpoint 5 to HKM, evidence that the sharp change in the time-term residuals in this area has some expression in very shallow parts of the crust.

A more detailed comparison of body wave and surface wave results can be obtained for the slow Rg paths by comparing the results of Kafka and Reiter (1987) with those of Klemperer and Luetgert (1987) for essentially the same path (Fig. 6c here and Profile 1 of their study). Based on that comparison, it appears that the V_p/V_s ratio used by Kafka and Reiter (1987) is too low and that a value of about 1.9 may be more appropriate.

DISCUSSION OF RESULTS

Relationship Between Lithotectonic Terranes and Regions with Distinct Seismic Structure

In this section, we discuss the relationship between our seismic results and locations of distinct lithotectonic terranes in the study area. The concept of lithotectonic terranes has been applied to the northern Appalachians by a number of authors, notably Williams and Hatcher (1982) and Keppie (1985). Lithotectonic terranes are crustal blocks of distinct stratigraphy, structure, petrology, metamorphism, and paleomagnetics with a geologic history different from that of adjacent terranes until some juxtaposing or suturing event occurred. In Maine, Williams and Hatcher (1982) defined three major terranes, from northwest to southeast: the Dunnage, the Gander, and the Avalon. Keppie (1985) argued that each of these are in fact accumulations of several terranes and should be considered superterranes. In Maine, he defined a number of terranes including the Boundary composite terrane, itself an accumulation of several terranes, in the northwestern part of the Gander superterrane. South of the Fredericton-Norumbega fault he identified a number of individual terranes within the Avalon superterrane, and he grouped this with the Meguma superterrane of Williams and Hatcher (1982) into the Acadia composite terrane (called here the Acadia composite/Avalon terrane, see Fig. 7). The Boundary composite terrane probably assembled in late Cambrian or early Ordovician and is thought to have accreted to the North American margin along with the rest of the Gander superterrane during Ordovician time (Keppie, 1985). According to paleomagnetic evidence, the Acadia composite/Avalon terrane had amalgamated by early Cambrian time and then underwent 1500 km of sinistral movement relative to the North American craton (Keppie, 1985; Kent and Opdyke, 1978; Kent, 1982). This movement was proposed to have taken

place during mid-Cretaceous time (Kent, 1982), but reexamination of the paleomagnetic evidence has moved this date back to at least the early Permian and significantly reduced the earlier-calculated offset (Kent and Opdyke, 1984).

Four regions of differing seismic characteristics can be generally delineated from our seismic results (see Fig. 11). The first region is in Canada where the two shotpoints show negative time-term residuals. These two shotpoints, both of which occur outside of the primary source/receiver area, are poorly sampled, each having only a few rays and a restricted range of approach azimuths. Thus, it is not clear whether these negative values represent real travel-time variations under the sites or whether they merely demonstrate an inadequecy of the tomographic time-term method for sites at the periphery of the experiment area. The second region extends from the Maine-Quebec border southeast to the central cross line of the USGS experiment (shotpoints 12 through 17). In this region, the time-term residuals generally scatter around zero. A few more positive time-term residuals occur in the area of the Boundary composite terrane in northwestern Maine. Generally, small time-term residuals would be expected for northwestern Maine from the upper-crustal velocity model along the northern cross line (shotpoints 8 through 11) reported by Doll et al. (1986). The more positive time-term residuals may represent real velocity or structure variations, or they could be an artifact of the anisotropy proposed for this area (Doll et al., 1986). The third region is that of generally larger, positive time-term residuals between HKM and the coast and between shotpoints 18 and 22. From the surface-wave and body-wave results presented in this study and those of Klemperer and Luetgert (1987), there appears to be a strong upper-crustal anisotropy (as much as 10%) in this region with the fast velocity direction being parallel to the regional trend of the geologic structure. The positive time-term residuals and the slow refractor block in this region may be due to the rather large angles with which a number of the Pg ray paths cross the regional structure. This region crosses the Norumbega fault zone from the Gander superterrane into the Acadia composite/Avalon terrane. The fourth region lies east of shotpoint 22 along coastal Maine. The Pg time-term residuals show a great variation but are generally consistent with the direct P-wave velocity measurements which show faster crust on average than that immediately to the west. Such a relationship between the two regions has also been noted by Klemperer and Luetgert (1987). There does not appear to be any strong anisotropy in this last region, which corresponds roughly to the St. Stephen terrane of Keppie (1985).

In general, the sharpest differences in seismic structure occur at the edge of and within the Acadia composite/Avalon terrane. Since the seismically slow crust in the Penobscot Bay region crosses the Norumbega fault, it seems unlikely that the Norumbega fault accounts for the strike-slip offset required by the paleomagnetic data. Such a fault would be expected to have an expression at the crustal depths sampled in this seismic analysis. Ludman (1986) has also argued for this interpretation of the Norumbega fault zone based on geologic evidence.

Correlation of Time-term Results with Gravity and Magnetic Anomalies

It is interesting to compare the Pg tomographic time-term results with the regional gravity and magnetic fields. There appears to be a correspondence between the tomographic time-term results and the Bouguer gravity field in Figure 10. More negative Bouguer anomaly values generally correlate with more negative time-term residuals, while more positive anomalies occur where there are more positive time-term residuals. The positive time-term residuals in northwestern Maine follow a ridge in the gravity field, a feature which is associated with the Boundary composite terrane. The sharp change in the time-term residuals to larger positive values southeast of shotpoint 4 occurs precisely at a step in the gravity field. This feature, known as the central Maine gravity gradient, stretches northeast into Canada where it is exposed at the surface in the Kingman fault zone (Ludman, 1986). Strong variations in the time-term residuals in easternmost Maine are coincident with a zone of rapidly changing gravity anomalies and of numerous plutonic bodies. Shotpoint 26 with its large positive time-term lies atop the western edge of a strong gravity gradient. Both the seismic and the gravity observations are due to contrasts located within the upper third of the crust.

The aeromagnetic field in Maine (Zietz et al., 1980) does not match the seismic results nearly as well as the gravity field does. This is not surprising since rocks of different density and elastic properties need not have different magnetic properties.

Implications for Models of Tectonic Evolution of the Crust Underlying Maine

The similarity of the seismic properties of the upper crust across the Norumbega fault zone as determined from the body-wave and surface wave analysis supports the arguments of Kent and Opdyke (1984) and Ludman (1986) that the Norumbega fault zone is not a major, post-Acadian megashear. On the other hand, the differences in the seismic properties of the upper crust between shotpoint 4 and HKM and between shotpoints 22 and 23 indicate the possibility of terrane boundaries at these localities. In the first case, the change in seismic properties occurs at the central Maine gravity gradient. This is also the location of the Kingman fault zone, a feature argued by Ludman (1986) to be a pre-Silurian suture. In the second case, the change occurs at the western edge of the coastal volcanic belt, a locality which has been suggested to be a terrane boundary (Klemperer and Luetgert, 1987).

The present-day evolution of the crust underlying Maine is expressed primarily by the seismicity of the region (Fig. 11). There is widespread activity in the southwestern part of the state, while more concentrated pockets of activity occur in the central and eastern areas. Except for a few small earthquakes which have been observed, northern Maine is the most seismically quiet part of the state. In central Maine there may be a tendency for earthquakes to occur in the vicinities of the boundaries

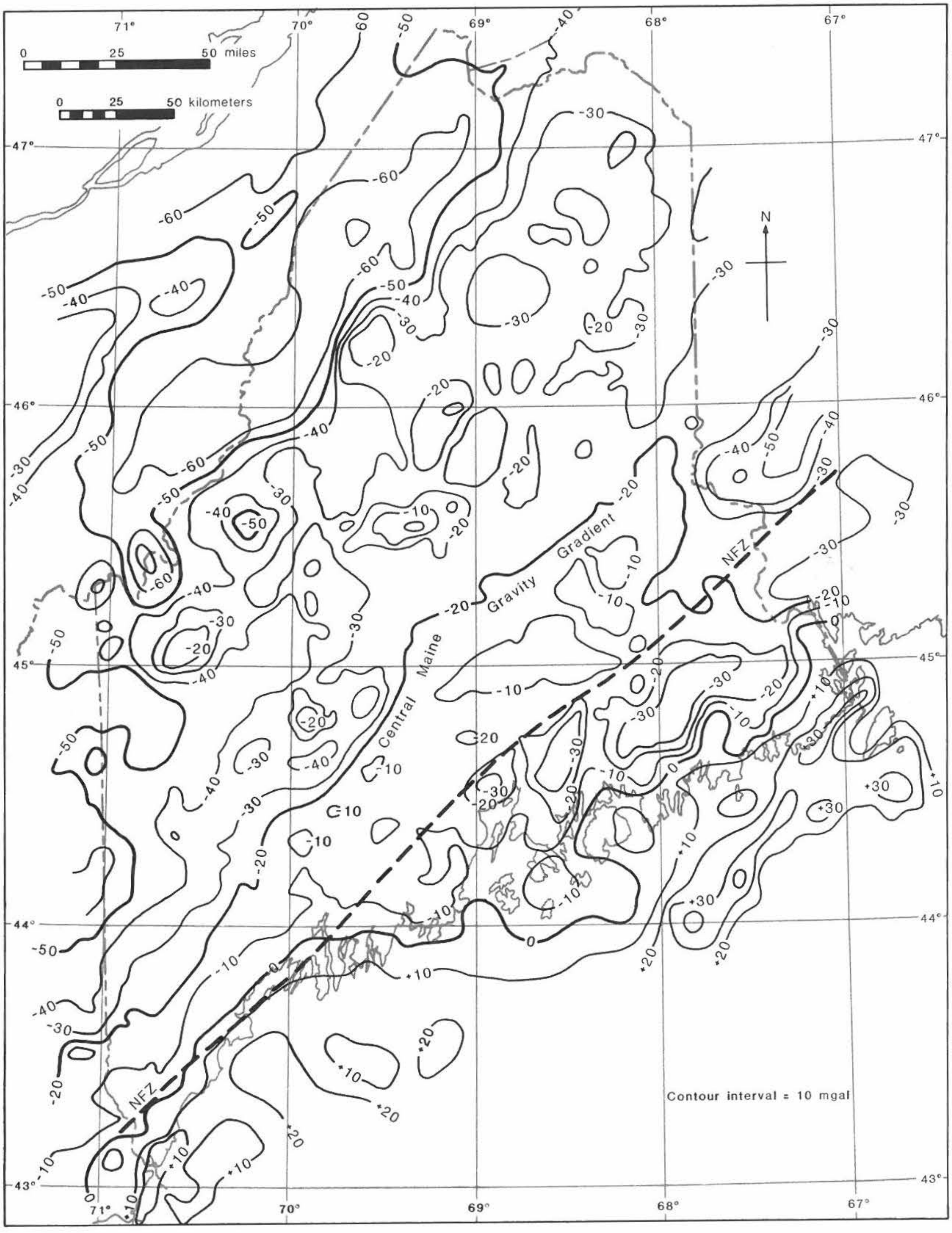

Figure 10. Bouguer gravity anomalies in the state of Maine (adapted from Haworth et al., 1980).

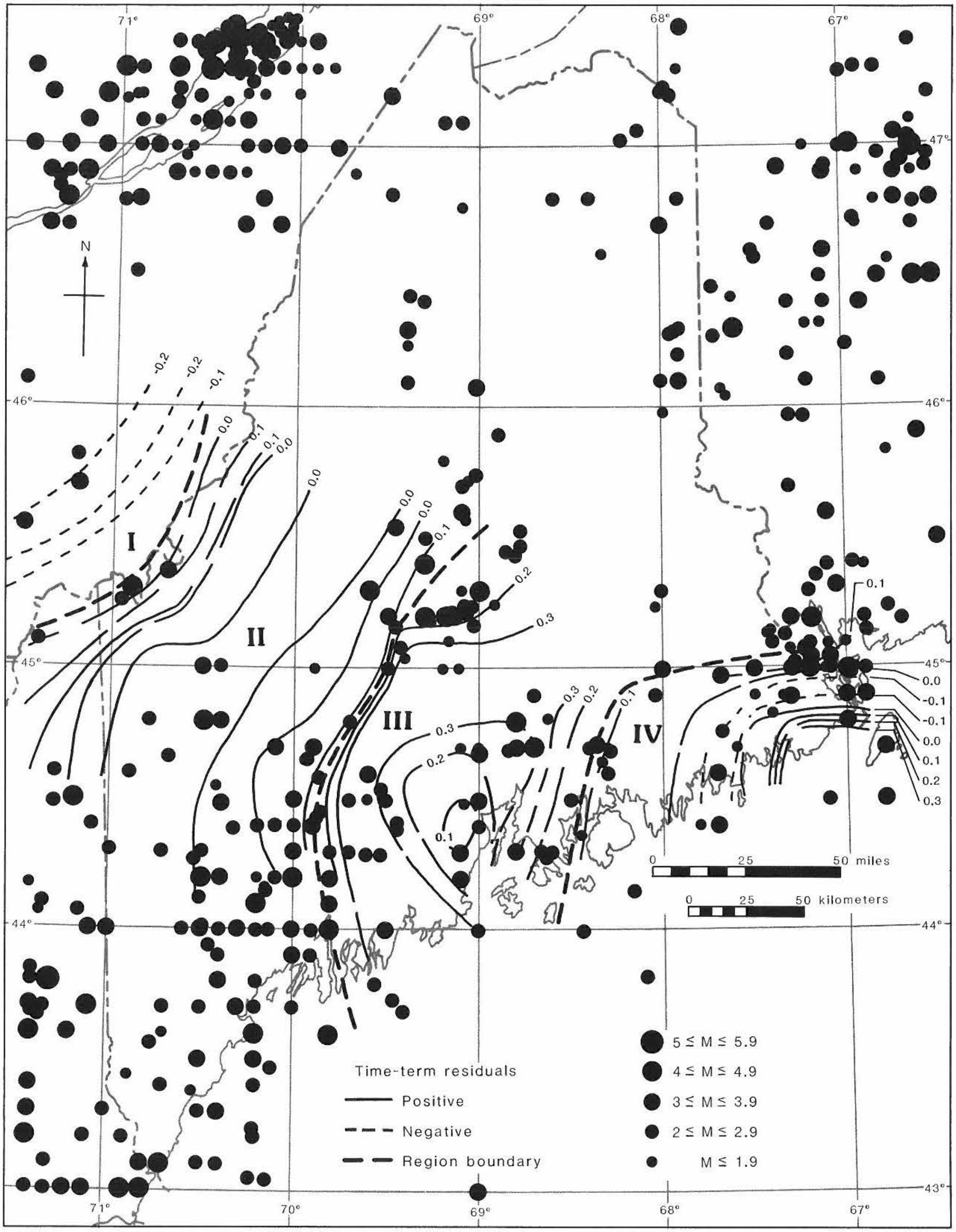

Figure 11. Geophysical basement terranes determined from the Pg time-term analysis. Also plotted are epicenters of earthquakes located in Maine for the period 1534-1985. Dashed lines indicate boundaries of four regions of different upper crustal seismic structure found in this study (I, II, III, and IV).

of the geophysically different regions found in this study (Fig. 11). For example, there is some indication of a band of seismicity that trends northeast from the southwest corner of Maine to the international border in east-central Maine. Such a trend would approximately follow the trend of the central Maine gravity gradient. Thus, some of the present-day earthquake activity could be due to the reactivation of a terrane boundary in the basement rocks by the modern tectonic stress field. The inability to identify presently active faults in the surface geology (Ebel, 1983; Ebel and McCaffrey, 1984; Ebel, 1986) may be an indication that it is in the basements blocks beneath the Paleozoic cover rocks that the now-active zones of weakness associated with the earthquakes can be found.

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