

MAINE GEOLOGICAL SURVEY

Robert G. Doyle, State Geologist

Shorter Contributions
to
Maine Geology

BULLETIN 23

Department of Economic Development

Augusta, Maine

1970

Appropriation Number 1230

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Plant Fossils of the Trout Valley Formation

Henry N. Andrews and Andrew E. Kasper¹

Introduction to Maine Paleobotany

Within the last decade Maine's contribution to paleobotany has markedly increased. Northern Maine, an area previously thought of as lacking significant plant fossils, has in recent years, yielded some exceptionally interesting localities with diverse plant assemblages. All localities described here lie within the Trout Valley Formation in Baxter State Park. (Fig. 1). There are several other fossil plant deposits to the north that are also being investigated by us but those in the Trout Valley Formation have proven to be the most productive thus far. It is the aim of this report to summarize our findings to date, to illustrate the diversity of forms and to describe in some detail the better known plants.

The history of paleobotany in Maine begins early but is not extensive. One of the first reports of fossil plants appeared in 1863 by the

Canadian geologist John W. Dawson; he described and illustrated fragmentary plant fossils from the Perry Basin of southeastern Maine. In this and numerous other works Dawson brought to light new genera and species of plants which laid much of the groundwork for Devonian paleobotany. The names of many Devonian plants can be traced back to this distinguished naturalist.

Much later, as a result of an investigation of the possible existence of coal suitable for mining in the Perry area, David White (Smith and White, 1905) described fragmentary plant fossils from the Upper Devonian of the Perry Basin. In 1962 Dorf and Rankin presented a short account of fossil plants from the Trout Valley Formation; it was, in fact, in that publication that the formation was first recognized and named. Schopf in 1964 described a new species of *Calamophyton* from the Mapleton Sandstone and in 1966 Schopf et al. reported finding erect plants in Early Silurian beds at Stockholm, Maine.

More recently two papers have appeared from our own laboratory that deal with new plants found in the Trout Valley Formation (Andrews et al., 1968; Gensel et al., 1969).

Early Land Vegetation

For reasons that will be given later we believe the fossil plants of the Trout Valley Formation to be of Siegenian or Emsian age although it is possible that the horizon may be as young as early Middle Devonian. In any event the flora includes vascular (woody) plants that are among the simplest and most primitive that have been found. It is difficult to compare them with modern plants and for this reason we have included some introductory comments on their habit, mode of preservation, and significant characters which can be used in identification. We hope that this may be of special interest to geologists who are not versed in Devonian paleobotany but who may find such information useful.

Recognizable and well preserved fossil algae and fungi have been recovered from chert dating back some 2 billion years (Barghoorn and

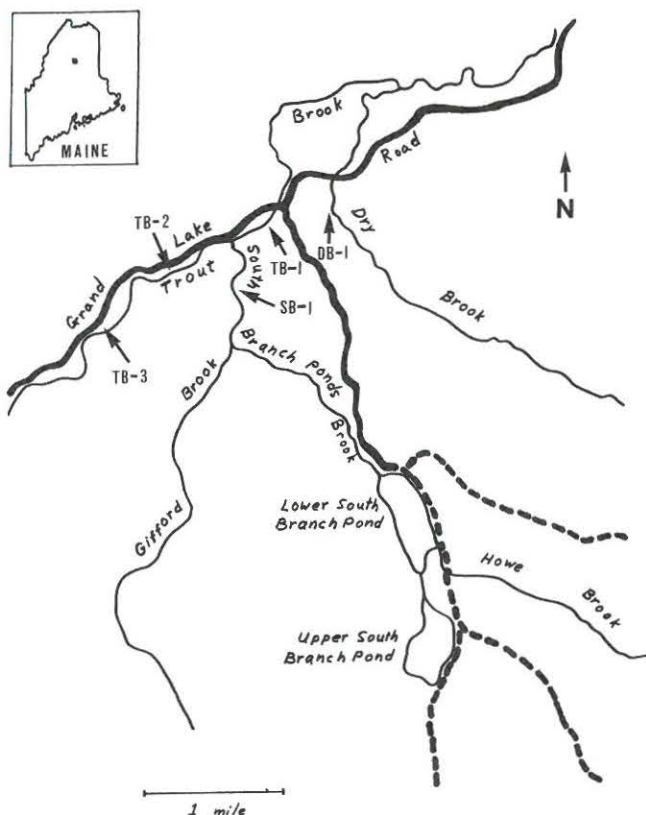


Fig. 1. A part of the Traveler Mountain Quadrangle, Maine, showing the area from which our fossil plant collections have been made in the Trout Valley Formation.

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Tyler 1965). From still older chert—approximately 3 billion years—fossil bacteria have been described and illustrated (Barghoorn and Schopf, 1966). Other than these preserved microorganisms we have very little significant and dependable information on plant life prior to the Devonian. We do not know as yet the specific factors which required such a long time lapse between the origin of plant life and the first vascular land plants.

The vascular plants that we find in early Devonian horizons are remnants of a vegetation that had just become adapted to a land environment. As such they possessed characteristics of both their algal ancestors and their recent emergence on land. In fact if certain features are not preserved it may be difficult to determine whether the fossils are algae or land plants. When the specimens are sufficiently well preserved the presence of the following land plant characteristics will distinguish the two: water conducting tissue, termed wood or xylem; a waxy coating (cuticle) over the aerial portions of the plant which served to prevent excess water loss; pores (stomata) on the axes and appendages which allowed for gas exchange. Also, the manner in which the sporangia were borne as well as spore and sporangial structure differ in the two groups; this is taken up in further detail below.

The bulk of the plant body of these early land organisms consisted of branching stems. Leaves as we know them today (large laminar organs with conducting tissue) had not evolved; therefore the stems themselves functioned as leaves in carrying on photosynthesis. The plant body presumably increased in size much as plants do today by means of multiplication and elongation of cells in the branch tips. Branching was accomplished by the division of this growing area into two. Such divisions may have been equal or unequal. When equal, a uniformly *dichotomous* branch system (Figs. 3A, 8) resulted—as in present-day *Fucus*, a common seaweed found along the Maine coast. When unequal, that is, if one “overtops” the other, it is referred to as *pseudomonopodial* (Figures 4, 9.) Some of the early land plants (as will be demonstrated in our Maine flora) had a pseudomonopodial pattern in which the main axis was perfectly straight and upright; in such a case there had been complete overtopping in the development of the plant. The term pseudo-

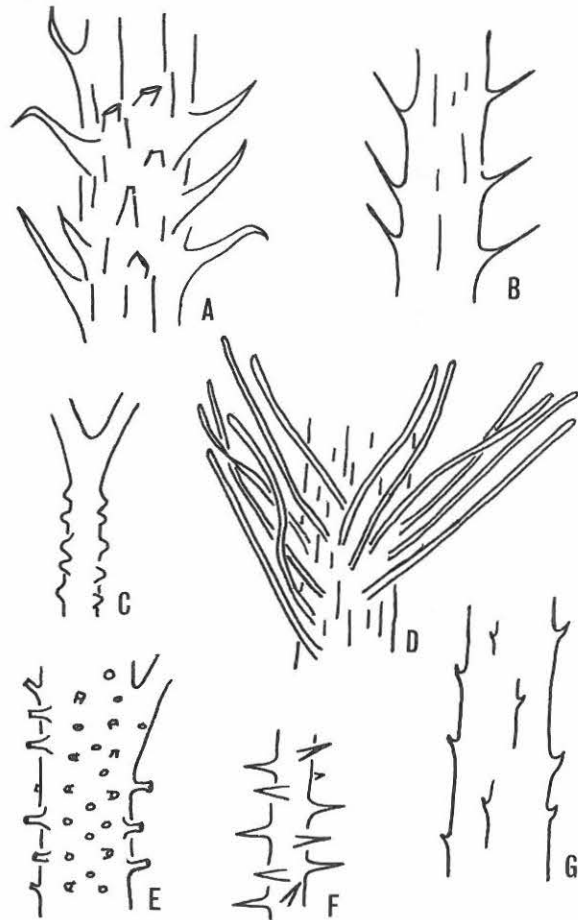


Fig. 2. Some representative early Devonian plants showing variations in the form and size of the enations, approximately natural size. A, *Drepanophycus spinaeformis* from the Gaspé. B, plant from the Dry Brook locality (DB-1, Fig. 1), possibly referable to *Drepanophycus*. C, *Crenatacaulis verruculosus* based on specimens illustrated by Banks and Davis (1969). D, *Baragwanathia longifolia* drawn from Australian specimens. E, *Psilophyton princeps* from specimens collected at locality SB-1, Fig. 1. F, “*Psilophyton princeps* var. *ornatum*” from the Gaspé. G, *Kaulangiophyton akantha* from locality TB-1, Fig. 1.

monopodial is used, rather than *monopodial*, to distinguish the morphology of these early plants from that of modern “higher” plants, such as a pine or sunflower. In these living seed plants the side branches originate from *buds* in the axils of leaves rather than a dichotomy of the branch tip. A more detailed discussion of this subject is not appropriate here. However, it may be added that the two modes of branching, dichotomous and pseudomonopodial, may be combined in a single plant which will have an “apparent main stem” and equally forking side branches.

Since fossil plants are usually recovered in numerous fragments the problem of whether a fragment is a main stem or side branch is a real one. To avoid a commitment either way when they are uncertain, paleobotanists refer to such fragments as *axes*.

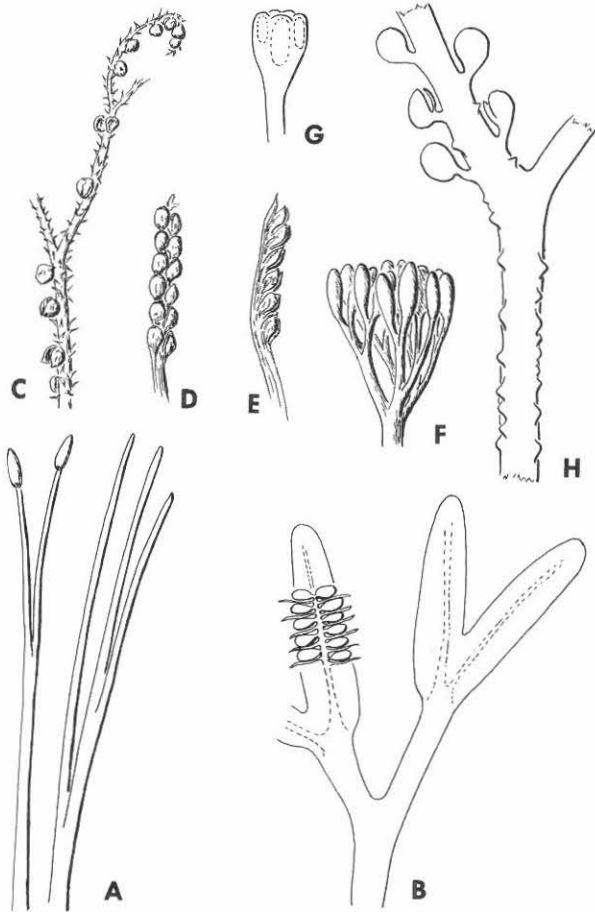


Fig. 3. Some representative early Devonian plants showing variations in the arrangement of sporangia. The measurement after each name is the maximum dimension of the sporangia. A, *Rhynia major* from Kidston and Lang (1921), (12 mm). B, *Krithodeophyton croftii* based on data given in Edwards (1968) and additional information supplied by Dr. Edwards; the exact relationship between sporangia and appendages is uncertain, (1.5mm). C, "*Psilophyton princeps* var. *ornatum*" from Ananiev and Stepanov (1968); see also Hueber (1964, 1967), (3 mm). D, E, *Bucheria ovata* from Dorf (1934), (about 3mm). F, *Hedeia* cf. *corymbosa* from Andrews, (1959), after Cookson (1949), (9 mm). G, *Yarravia oblonga* from Andrews (1959), after Lang and Cookson (1935), (7 mm). H, *Crenaticaulis verruculosus* based on photos and information in Banks and Davis (1969), (2mm).

Another broadly used and non-committal term is *enation*. As defined by Webster's this is "an outgrowth from the surface of an organ." We use the term in reference to leaf-like, spine-like, hair-like or gland-like appendages that are borne on the axes of these early plants. In Figure 2 we have selected some representative examples of enations of early Devonian plants. They are drawn approximately to scale and it will be noticed that there is considerable size difference, ranging in length from one or two mm up to several cm. They may be very delicate (almost hair-like) or very stout; they may be abundant or sparsely distributed and the numbers may vary on different parts of a plant. In many cases they are distinctive in shape and arrangement and are, therefore, regarded as important in plant identification and classification.

Of great importance are the reproductive organs of these plants. Since the plants preserved represent the sporophyte generation of the life cycle, the reproductive organs are the spore sacs (sporangia) with their contained spores. When we refer to a specimen as being *fertile* it is implied that sporangia are present. There are two general positions in which the sporangia may be located, either *terminal* on the main or lateral branches or *lateral* along the axes. There are, however, many modifications of these two patterns; a representative sampling is shown in Figure 3. Paleobotanists are particularly interested in the matter of the size, shape, mode of attachment and distribution of sporangia on a plant. How did all of the variations that we presently know — and more are revealed each year—originate? What was the starting point and what are the paths of evolution that have been traversed? How did the patterns that we find in these early Devonian plants evolve into the much more complex ones of the Middle and Upper Devonian? These are some of the botanical problems with which we are most vitally concerned.

One other introductory matter may be appropriately introduced here, that is, what constituted the root system of these early plants. Roots of Devonian plants in general and those of the Trout Valley flora in particular are apparently missing. One explanation is that the upper portions of these plants were broken off and transported to a suitable site for preservation while the roots were not. It is highly probable, how-

ever, that most of the early Devonian plants were not differentiated into roots, stems and leaves, as we find in most living higher plants. We have already noted that in some fossil plants (Figure 4) there is no differentiation into stem and leaf and in others (Fig. 2, B,F,G) the "leaves" are at best very rudimentary. A clue is also to be found in the root-stem relationship in certain modern plants. In the Lycopodiums or Clubmosses (plants of very ancient lineage) that are so common in New England, the anatomy of the structures that function as roots and stems is essentially identical; the roots, however, are smaller and do not bear leaves. The general habit of some of the early Devonian plants (Fig. 10) was probably quite like that of some of the trailing Lycopodiums. That is, the stems crept along the ground, branching quite freely; some of the branches were erect and fertile and may or may not have borne small leaf-like structures (enations); other

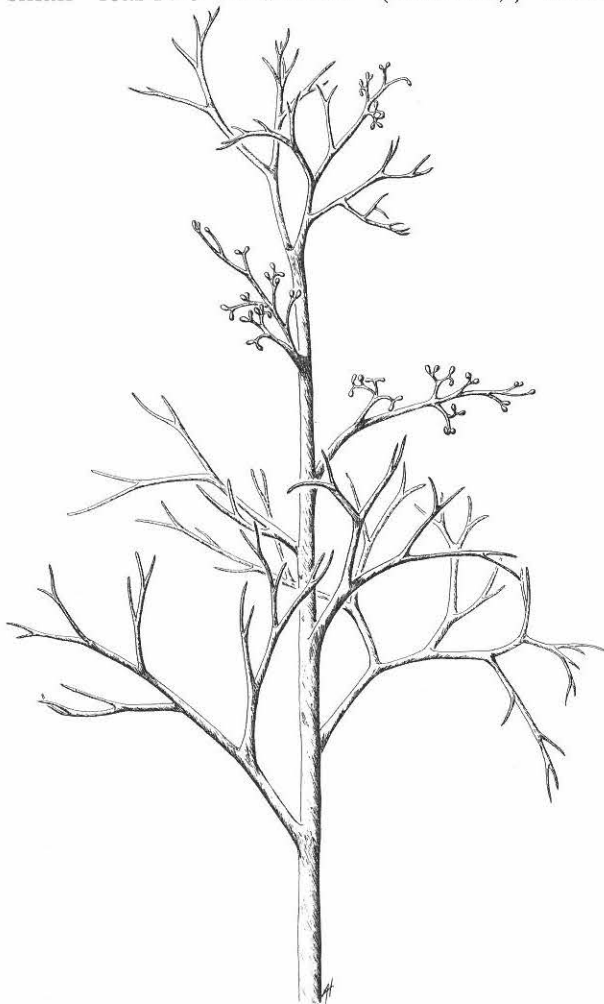


Fig. 4. *Psilophyton forbesii*. A restoration of the plant based on numerous specimens from locality SB-1. From Andrews et. al. (1968).

branches which did not bear enations penetrated the soil and functioned as roots.

Plants of the Trout Valley Formation

We have obtained abundant plants from several localities and in all cases they occur as impressions or as thin, non-cellular, carbonaceous films on the rock surfaces. It is disappointing that petrified remains (which would reveal cell structure) have not been found. However, as will be demonstrated, we have obtained a considerable wealth of information concerning the gross morphology of the plants, some of which have proven to be new species or new genera, while others seem to be identical with plants reported previously from other places.

Psilophyton forbesii. This is the first plant in the area that we studied in detail; excellent specimens were originally obtained by William Forbes and we have made several collections since. The site SB-1 (Fig. 1) is the east bank of South Branch Ponds Brook about 500 yards north of its junction with Gifford Brook. *P. forbesii* (Fig. 4) was a plant that stood two or more feet high; it had an upright (pseudomonopodial) axis with dichotomizing side branches some of which were terminated by pairs of sporangia. The stems attain a maximum width of 9 mm and the smallest branchlets are about 2 mm. The compressed stems display distinct longitudinal striations the exact significance of which we are not certain. It seems most likely that they represent fiber bundles or the arms of a lobed vascular (wood) system. In life the stems were probably smooth, the longitudinal striations being revealed only after compression and there being no evidence of enations of any kind. The ellipsoidal sporangia measure about 4.0 x 1.5 mm in their long and narrow diameters.

The vagaries of fossil plant collecting are well illustrated by this plant. We have excavated at the site noted above at least seven or eight times; the stems and branches are abundant and rather beautifully preserved as a maroon-brown compression but we have found only one specimen bearing sporangia. The day we found it was regarded as a very successful one! In general the reproductive structures of our Maine fossils seem to be either very abundant or almost completely lacking. The latter situation

is not easy to explain. Some possibilities may be that either the plants were preserved at a time of year when they were not reproducing, or the plants simply did not produce sporangia abundantly and relied on vegetative reproduction. These are little more than guesses.

From the same locality as mentioned above we have identified specimens of *Psilophyton princeps* sensu Hueber, 1967. (Fig. 5). Since this is quite distinct from *Psilophyton forbesii* a few words of explanation are in order concerning the use of this binomial. *P. princeps* was first described by J. W. Dawson in 1859 and was based on collections that he made along

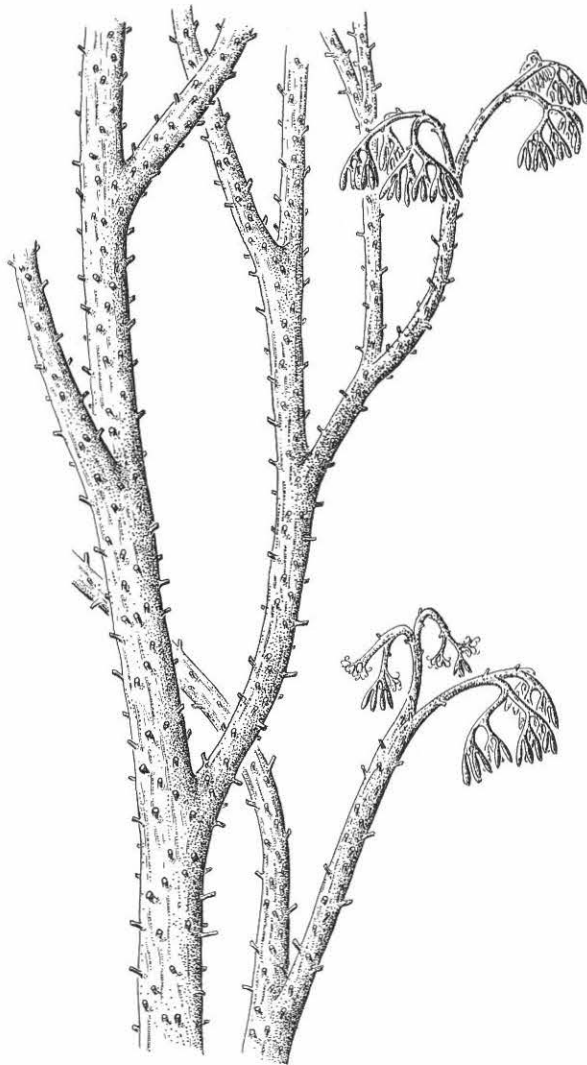


Fig. 5. *Psilophyton princeps* Dawson, a restoration prepared by Hueber (1967), fig. 1. For a detailed explanation see pages 6-8, and Hueber and Banks, 1967.

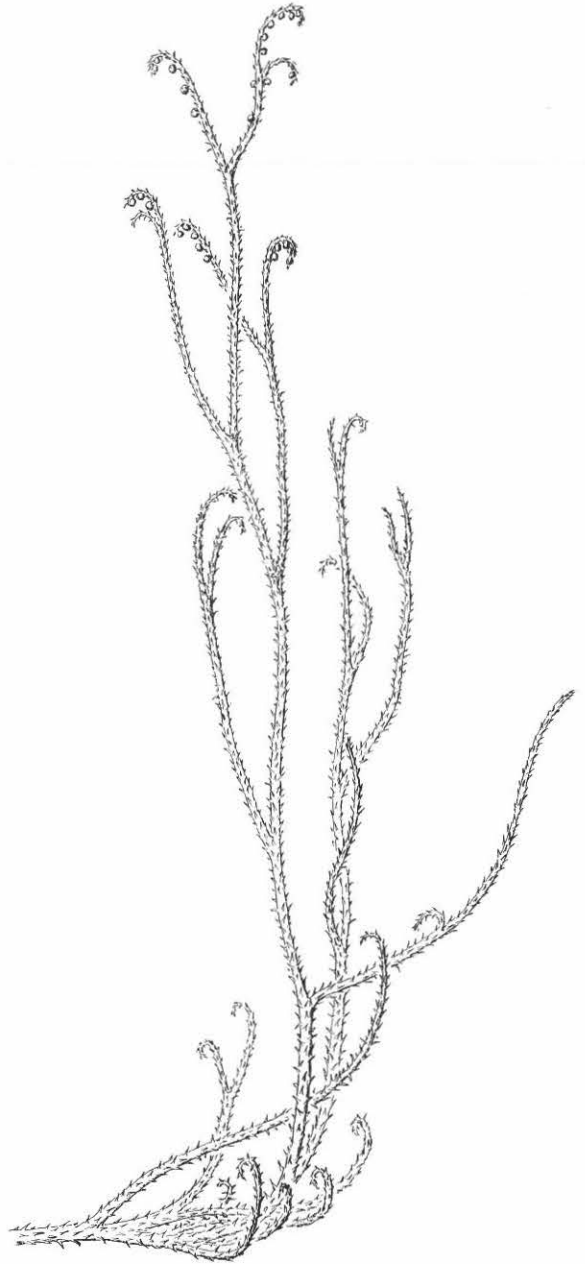


Fig. 6. Restoration of "*Psilophyton princeps* var. *ornatum*" Ananiev and Stepanov (1968), based on fossils from the Lower Devonian of western Siberia.

the Gaspé coast. He published additional data a few years later and his restoration of the plant became a classic that was copied by many later writers of botany and geology texts. The history of our knowledge of this plant has become a rather long and complicated one. It is well to remember that Dawson's field work was carried out under conditions that were anything but easy. There were few roads on the Gaspé in those days, travel being mostly by boat. It is not known exactly where Dawson made all

of his collections and some of them have been lost. Recent careful reinvestigations by Hueber and Banks (1967) indicate that Dawson combined fragments of three distinct plants under the name *P. princeps*. However, as noted above, Dawson's restoration of the plant attracted many later workers who were not familiar with all of the technical details involved and the initial errors have been perpetuated. Several later workers have described new species of *Psilophyton* based on very fragmentary fossils and thus added confusion to an already questionable genus. In 1916 the Swedish botanist Halle pointed out that Dawson had not proven actual connection between the spiny axes of *P. princeps* and its sporangia. He therefore proposed the generic name *Dawsonites* for the sporangia and the rather delicate, smooth branchlets to which they were attached. Some years later Croft and Lang (1942), and more conclusively Hueber (1964), showed that the spiny axes which Dawson had described as *Psilophyton princeps* variety *ornatum* actually bore lateral sporangia. Even more recently the Soviet paleobotanists Ananiev and Stepanov (1968) have published photos and a fine restoration of "*P. princeps* var *ornatum*" (Fig. 6) showing the spiny stems with lateral sporangia; this account is based on fossils from the Lower Devonian of western Siberia.

Also, Hueber and Banks (1967) have stated that the supposed root system of Dawson's reconstruction is actually referable to the genus *Taeniocrada*. The latter occurs in our Maine flora and will be considered later.

What then remains of Dawson's original



Fig. 7. Locality TB-1 on the south bank of Trout Brook about 200 yards upstream from The Crossing.

plant? Hueber and Banks (1967) after searching through Dawson's remaining collections, and after collecting extensively from localities that Dawson visited, selected a new type specimen and have redefined the genus. Later, Hueber (1967) clearly described and illustrated the type species *Psilophyton princeps* (Fig. 5) —markedly dissimilar from *P. princeps* var. *ornatum* (Fig. 6) for which a new generic name is needed. The distinctive features of the genus *Psilophyton* then, are as follows: the axes are naked or spiny and may be pseudomonopodial or dichotomous or a combination of both; the sporangia which are elliptical and dehisced longitudinally are pendulous and borne in pairs on the ends of lateral branches; the wood is a slender solid strand. It is the opinion of the present writers that this revised description may prove to be too broad and will require changes as we continue to learn more about these early land plants. However, it is a step in the right direction and clears the air around this key name in the history of Devonian paleobotany.

The most productive fossil plant locality (TB-1, Fig. 1) that we have found in the Trout Valley Formation is a conspicuous outcrop (Fig. 7) along the south bank of Trout Brook about 200 yards upstream from The Crossing. Two plants from here probably fall within the genus *Psilophyton*. The first, referred to here as plant-A as a matter of reference, is known from axes that are 2mm in width, smooth (lacking any ribbing or spines) and are dichotomously branched (Fig. 8). The ultimate branchlets terminate in pairs of small, elliptical sporangia that are 2.0 mm long and 0.8 mm in diameter. In striking contrast to our specimens of *P. forbesii*, this Trout Brook plant is abundantly fertile. We first encountered it in the summer of 1967 and were delighted with the numerous sporangia-bearing specimens that we found. It is, however, worth noting that this, like other Devonian fossil plants, is a three-dimensional structure and when it is first exposed by splitting the rock matrix only a portion of the plant is immediately revealed. We then work with fine needles in order to expose the entire branch system that dips down into the rock. Frequently, branches are broken and cannot be shown connected with the stem or we are unsure as to whether we have chipped out a representative portion of the plant. Thus a great many hours of work may go into the preparation of a single specimen which we feel is



Fig. 8. A specimen of "plant-A" from locality TB-1 showing dichotomous branching and paired terminal sporangia (2mm long).

representative of the gross morphology of the plant.

We have another plant from locality TB-1 that is likewise probably referable to the genus *Psilophyton* and it is designated here as plant-B. Since specimens bearing sporangia have not been found as yet there is some uncertainty as to its classification. The plant is typically pseudomonopodial (Fig. 9) in habit with long slender axes up to 3 mm in width; these bear lateral branches that dichotomize quite profusely, terminating in very fine branchlets. The main axes are smooth and longitudinally ribbed — the latter effect probably the result of compression and exposure of either internal fiber bundles or a lobed vascular strand. This plant occurs in great abundance through several inches of rock. It is found at the same horizon as plant-A but the two are separated by a distance of about 100 feet along the outcrop. Are they



Fig. 9. A specimen of "plant-B" from locality TB-1 showing the pseudomonopodial habit (diameter of main axis is 2mm).

parts of one and the same plant or are two species present? It probably will not be possible to resolve this question until we are successful in finding fertile specimens of plant-B.

The most interesting plant that we have found at locality TB-1 is one that we have named *Kaulangiophyton akantha* (Gensel et al., 1969). Compressed axes of this plant attain a diameter of 9 mm and bear slightly curved spines about 2 mm long. The spines are rather sparsely distributed along the axes, yet quite distinct. Branching follows a characteristic pattern (found in a few other Devonian plants) that is called H- or K-type branching. This pattern occurs when a side branch dichotomizes in a wide angle near the main axis and each of the resulting limbs parallels the main stem in opposite directions (Fig. 10). We believe that the plant may have had a trailing habit like that of some modern Lycopodiums, with short

branches that dichotomize several times. The ultimate branchlets bear numerous sporangia, each being about 3 mm long and 1 mm broad. They appear in a rather dense cluster or fan-shaped pattern on the rock.

Our study of this plant is in its initial stages. A good deal of excavation of the specimens collected remains to be done but we have been fortunate in obtaining a fine suite of specimens that show many aspects of the general habit of the plant—and numerous fertile specimens. The plant was evidently at least two feet tall and quite distinct from any others that we have encountered. In checking the literature the closest comparison that we can draw is with *Trimerophyton robustius* (Hopping, 1956). *Trimerophyton* was founded on a specimen originally collected by Dawson from the Gaspé coast and which he identified as *Psilophyton robustius*. The genus *Trimerophyton* is known from a single specimen and is characterized by side branches that trifurcate. At present we are not sure that our specimens possess exactly the same kind of branching and sporangial arrangement. In the present state of our investigation it looks as though our Maine fossils will prove to be a distinct species of *Trimerophyton* or possibly a new allied genus.

Fragmentary remains of spiny axes have been found at two additional sites (TB-3 and DB-1) in the Trout Valley Formation. The first of these outcrops is located on the north bank of Trout Brook a short distance upstream from TB-2. Here we have found long segments (up to 25 cm) of unbranched axes 1.5 to 2.0 cm broad with fairly stout spines up to 6 mm long.

The second locality (DB-1) is on the west bank of Dry Brook approximately 200 yards south (upstream) from the Grand Lake Road. The largest stem fragments that we have found here are about 12 cm long, 1.5 cm broad, and bear fairly stout spines or leaves 10 mm long which taper to a fine point (Fig. 2, B). These axes rarely display branching; some specimens show a carbonaceous strand (presumably representing wood) in the center but there is no evidence that it extended into the leaves or spines. This is a fossil that most paleobotanists would assign to the genus *Drepanophycus*. We are reluctant to do so at present and hope that continued collecting will turn up fertile specimens.

The most problematical plant in the flora by virtue of both size and structure is *Protot-*

axites. This is a unique and curious plant that has puzzled many paleobotanists—a few preliminary comments are required. It was first described by Dawson (1859) in the same publication in which he first reported *Psilophyton*. He apparently believed it to be a coniferous log whence the reference to the Yew genus (*Taxus*). Since Dawson's time it has been found in several other places scattered through the late Silurian and Devonian. The "logs" attain a diameter of nearly three feet and consist of two kinds of tubular cells aligned more or less longitudinally. The best preserved specimens reported to date are ones that Arnold (1952) described as *Prototaxites southworthii*, from the Upper Devonian Kettle Point black shale, Kettle Point, Lambton, Ontario. The cellular structure is unique; it is not a vascular plant and certainly not at all closely related to the conifers. The closest comparison seems to be with the stipes (stalks) of some of the larger brown seaweeds but specialists who deal with the living algae are reluctant to accept one with a stem three feet in diameter!

Fragments of *Prototaxites* "logs" are fairly abundant in the conglomerate along South Branch Ponds Brook. They appear as masses of black chert in the conspicuous exposure on the west bank, about 100 yards south of the junction with Gifford Brook, and in the stream bed itself. In the summer of 1969 we found a large specimen in the stream bed gravel of Dry Brook about ¼ mile upstream from the road. All of the Maine specimens that we have found to date are very poorly preserved.

The Classification of Early Devonian Plants

It is evident that the Trout Valley Formation flora includes several distinct types of early land vascular plants, some of which are quite complete and others are known from tantalizing fragments. We expect to add significantly to our present state of knowledge of the fragmentary plants during the next few years.

The numerous contributions that have been made in recent years, clarifying certain groups, and introducing new ones, have resulted in rapidly changing systems of classification. H. P. Banks (1968) has summarized the evidence clearly in a recent study "The early history of land plants." Four of the groups in his classification are of special concern here; they are subdivisions of the Division Tracheophyta

(plants with xylem or woody conducting tissue):

Rhyniophytina. These are small and rather delicate plants with naked stems that dichotomize and bear terminal sporangia. The classic and much figured *Rhynia* (Fig. 3 A) from the Lower or Middle Devonian of Rhynie, Scotland, is typical. *Taenioocrada*, in the Maine flora, is included here.

Trimerophytina. This group includes plants that have a pseudomonopodial habit in which the side branches dichotomize or trichotomize and terminate in rather dense masses of sporangia (Fig. 5). The *Psilophyton* species that we have described belong here as well as the fossil that we consider to be referable to, or closely related to, *Trimerophyton*.

Lycophytina. This is the lycopod group which includes the extant genus *Lycopodium*, so abundant in New England woodlands, and several abundant and well known arborescent plants in the Carboniferous. The axes are clothed in abundant microphylls (small leaves) and the sporangia are borne on the upper surface of a sporophyll. The sporophylls (leaves bearing sporangia) may be aggregated into distinct cones or they may be scattered among the ordinary leaves as in the living *Lycopodium lucidulum*. *Drepanophycus* (Figs. 2 A; 11 B) is included here; however, as we have pointed out, some caution is needed since most of the fossils that have been identified as *Drepanophycus* are sterile. The classic *Asteroxylon mackiei* (Lyon, 1964) from Rhynie, Scotland, is probably best placed here. *Baragwanathia longifolia* from Australia was originally reported to be of Silurian age but more recent studies (of associated graptolites) suggests a Lower Devonian age; in either case it is one of the most ancient of vascular plants and seems to be a good typical lycopod.

Zosterophyllophytina. This group includes plants that are naked or spiny and with sporangia that are attached laterally, either in terminal clusters or along the axes. Some representative examples are shown in Figure 3C, D, E, H. Our *Kaulangiophyton akantha* seems best assigned to this group but it is not a "perfect fit"! If the enations were larger it would probably be more appropriately placed in the *Lycophytina*. Banks (1968, p. 78) proposes that the plants in this group, with laterally attached sporangia, should no longer be called psilophytes.

Any classification such as this is a man-made affair. Our knowledge of early Devonian plants is expanding quite rapidly and classifications will have to change accordingly. It is the plants that do *not* fit that are most interesting to the paleobotanist who is trying to clarify the intricate paths of plant evolution.

The Age of the Trout Valley Flora

The origin and early phases of evolution of vascular plants have been studied and reported on quite extensively in recent years. On the basis of both macrofossils and spores, evidence has been put forward that vascular plants existed as far back as the Cambrian; one of us has summarized the presumed evidence (Andrews, 1966). However, careful scrutiny of these reports by several competent workers in very recent years leaves no doubt that there is little or no satisfactory evidence for the existence of vascular plants prior to late Silurian times.

Based on studies of macrofossils, Banks (1967, p. 728) has succinctly summarized the early evolution of vascular plants as follows: "... arising in Late Silurian time, expanding slowly during Early Devonian, broadening considerably during Middle Devonian, and becoming much more complex in Late Devonian as some of the earliest types receded."

Based on studies of spores, Chaloner (1967, p. 83) offers a closely comparable picture: "The spore record suggests a relatively small number of land plant types first appearing in the Silurian, with steadily increasing diversity through the Devonian. This diversity is accompanied by a divergence in spore size groups through the Devonian into two increasingly clearly defined size classes by the Carboniferous."

Although there is much work remaining to be done in our study of the Maine fossils, it is appropriate to offer a tentative opinion of the age of the flora. As a basis of comparison we are taking the summarized evidence for plant macrofossils presented by Banks (1967, 1968) and for microfossils (spores) that were presented by Chaloner (1967). This is shown in Table I in which the time divisions are taken from the recent Devonian correlation chart by Oliver, et al., 1969.

Next, we may dispense with such apologies as are necessary. It is well known that many plant genera are racially long-lived; to take two striking examples, the pines (*Pinus*) and the

Table I. Distribution of some representative plants in the Devonian. Time divisions based on Oliver et. al., 1969. (Space allocated to the several time divisions is not in proportion with reference to time in millions of years.)

Famennian	Upper Devonian	Rhacophyton Archaeopteris-Callixylon Archaeosperma arnoldii
Frasnian		Rhacophyton, Cladoxylon, Schizopodium Archaeopteris-Callixylon, Tetraxylopteris Aneurophyton
Eifelian Givetian	Middle Devonian	Schizopodium, Svalbardia, Aneurophyton Cladoxylon, Pseudoporochnus nodosus Calamophyton bicephalum
Emsian	Lower Devonian	Spores well in excess of 200 μ
Siegenian		Cooksonia, Toeniocrada, Eogaspesia, Hicklingia, Rhynia Zosterophyllum, Bucheria, Asteroxylon, Trimerophyton Hyenia, Enigmophyton, Crenaticaulis Spores slightly in excess of 200 μ
Gedinnian		Cooksonia, Drepanophycus, Bagawanathia, Yarravia, Hedia, Gosslingia, Zosterophyllum, Psilophyton princeps "Psilophyton princpls var. ornatum"
Ludlovian	Silurian	Cooksonia

Cinnamon ferns (*Osmunda*) which are among New England's most widely distributed and best known plants, have been on the scene for well over one hundred million years and are not especially good index fossils. The Devonian, too, apparently includes some long-lived genera: *Drepanophycus* is reported to be widely distributed through the Devonian (Banks, 1968; Grierson and Banks, 1963) and "*Psilophyton princeps* var. *ornatum*" ranges from the Lower Devonian to the early Upper Devonian (Hueber and Grierson, 1961). In the case of *Drepanophycus* it is the opinion of the present writers that a great deal of work remains to be done before we clearly understand the limits of this "genus."

However, leaving these apparent exceptions, and considering the Lower Devonian floras as a whole, a glance at Table I shows the kinds of plants that appeared before the close of the Emsian were small; many were leafless (*Rhynia*, *Cooksonia*, *Trimerophyton*, *Yarravia?*, *Hedra?*) but others had enations (e.g., *Asteroxylon*, "*Psilophyton princeps* var. *ornatum*"). In only a few are true microphylls present which attain a significant size (*Drepanophycus*, *Baragwanathia*). The vascular structure (stem anatomy) when known is mostly a simple or slightly lobed cylinder (in *Asteroxylon* and *Baragwanathia* they are quite deeply lobed, pointing out things to come). The sporangia are borne terminally or laterally on relatively unspecialized branches.

Two of the best known plants of the Middle Devonian (lower Givetian) are *Pseudosporochnus nodosus* and *Calamophyton bicephalum* from Goe in eastern Belgium. *Pseudosporochnus* has a main stem that is larger than that of most of the earlier plants and there is a distinct tendency to differentiate leaves. In *Calamophyton* the foliar organs are distinct three-dimensional structures and the fertile appendages considerably more complex than plants of the Lower Devonian. It is not necessary to discuss in any detail the genera that come into the picture above the Givetian. A few examples may be noted to indicate the great evolutionary advances: *Schizopodium* is a plant with a highly complex polystelic trunk (personal communication, F. M. Hueber); *Rhacophyton* has very complex fertile fronds, highly specialized sporangial morphology and secondary wood (Andrews and Phillips, 1968); *Archaeopteris* is a large arborescent plant that had achieved heterospory and strong cambial

activity; and a true seed, *Archaeosperma arnoldii*, appears in the geologic record for the first time in the Famennian (Pettitt and Beck, 1968) — the plant kingdom made fantastic strides between the early and late Devonian.

The Eifelian record seems to be relatively sparse and this is a matter of concern. However, on the basis of the information that we have available now, the Trout Valley Formation flora appears to be middle or upper Lower Devonian.

Acknowledgement

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A New Silurian Fossil Locality on Lawler Ridge, Sherman Quadrangle, Maine

by David C. Roy¹ and William H. Forbes²

During the summer of 1969 Forbes collected graptolites from an exposure along Interstate 95 on the west slope of Lawler Ridge near Little Salmon Stream Lake in the southwest corner of the Sherman quadrangle. The graptolite assemblage of Late Llandovery age comes from a sequence which is tentatively correlated with "The Rocks of Island Falls" as mapped by Ekren and Frischknecht (1967) in the Island Falls quadrangle and is coeval with portions of the Allsbury Formation of Neuman (1967).

The exposure is located on the east side of Interstate 95 at a point .60 miles north of the township line between T 1 R 6 W and Herseytown (T 2 R 6 W). The graptolites were obtained from a sequence of thinly interbedded black, finely cleaved, pyritiferous slate and light gray calcareous and non-calcareous, fine grained sandstone and siltstone. The fine-grained sandstone and siltstone beds ($\frac{1}{8}$ " - 2" thick) are parallel and cross laminated; several show well developed ripples. The graptolites were obtained from the slate phase at the excavated edge of the outcrop.

The collection has been studied by Dr. William B. N. Berry who reports the following forms (written communication to Forbes dated October 14, 1969):

Monograptus of the *M. dubius* group (*M. praedubius* Boucek?)

Monograptus priodon (Bronn)?

Monograptus sp (*Monograptus* of the *M. nudus* group)

Dr. Berry points out that the specimens are badly deformed, being "... either pulled out or squashed ...". A Late Llandovery age (probably in the span of Ellis and Wood zones 24-25) is assigned to the above assemblage. Dr. Berry further states that he thinks "... the Late Llandovery age is fairly reliable even though I can't put any exact names on the forms present."

The terrain surrounding the fossil localities is unmapped so far as the authors know. R. B. Neuman (Neuman, 1967) has mapped exten-

sively in the Stacyville quadrangle to the west; E. B. Ekren and F. C. Frischknecht (Ekren, 1961, Ekren and Frischknecht, 1967) have studied the Island Falls quadrangle to the north of the Sherman quadrangle. The authors have made only a brief study of exposures along Interstate 95 north and south of the fossil locality in an attempt to determine the lithologic character of the immediate sedimentary section.

The sequence from which the graptolite collection comes is part of a discontinuously exposed slate and fine sandstone section. The facing direction at the fossil locality is clearly to the southeast as evidenced by cross-lamination and ripple structures in the thin fine sandstone layers. The structural trend of this exposure and those nearby to the south is N8 - 25°E with dips averaging 82°SE. Apparently stratigraphically below the beds with the graptolites and exposed in a large roadcut a short distance north on the interstate highway is a moderately folded thinly interbedded sequence displaying approximately equal proportions of dark slate and light-gray fine sandstone or siltstone. The fine-grained sandstones and siltstone show many of the turbidite features described by Bouma (1964) with the lower laminated (B) and cross-laminated (C) intervals dominating. The sandstones and siltstones appear to be largely quartz-carbonate mixtures with mica concentrations defining the lamination; an orange weathering color is characteristic.

Stratigraphically above the graptolitic horizon are thin to massively bedded, light gray, calcareous, micaceous, orange-weathering, fine-grained sandstones. Internal structuring in these beds is limited to local parallel lamination. The soles of the beds, however, display parallel and anastomosing ridges and grooves. Paleocurrent measurements on three separate beds and simple unfolding yields an average trend direction of S82°E (N82°W) for these linear sole features (range N89°E to S70°E).

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No directional sense can be assigned to these readings; no corrections for local or regional plunges can be made since these beds have not been placed in a structural framework.

Neuman (1967) assigns most of the southeastern corner of the Stacyville quadrangle to the Allsbury Formation of Ekren and Frischknecht (1967). In particular, he maps the terrain as underlain by the slate member which consists of "medium-to dark-gray, greenish-gray, and red slate and siltstone and a few beds of fine-to medium-grained sandstone." Locally within the slate, beds "...as much as 10 feet thick are formed of light-gray calcareous siltstone with closely spaced thin slaty partings." Fossil localities from the slate member in the southeastern corner of the Stacyville quadrangle (Neuman, 1967, p. 24) suggest an age range for that member of Late Llandovery to Early Ludlow. Thus, the graptolitic beds along Interstate 95 are coeval with the early phases of the slate member of the Allsbury Formation.

A very close lithologic correlation can be made between the exposures on Interstate 95 and the "Rocks of Island Falls" as mapped in the southeast corner of the Island Falls quadrangle by Ekren and Frischknecht (1967). Their description of the slate and sandstone phase of this unit is, in part, as follows:

"At Island Falls and May Mountain the rocks of Island Falls consist of medium-gray, green-gray and dark-gray, thin-bedded slate, siltstone and very fine to fine-grained sandstone. Thin beds of silty limestone averaging less than 6 inches occur throughout . . . The thickness of sandstone beds averages about 3 inches but is locally as much as 3 feet. The beds show fair grading and nearly everywhere are finely cross laminated.

The sandstone is quartzitic, and, in most exposures, calcareous . . . The abundance of quartz and calcite in the sandstone is a distinctive feature . . ."

No fossil localities in the "Rocks of Island Falls" unit have been found in the Island Falls quadrangle; however, Ekren and Frischknecht (1967) assign a Late Llandovery to Early Ludlow range to the sequence based upon localities in the Smyrna Mills quadrangle, on-strike

to the east, and upon stratigraphic relationships of the unit to the Mattawamkeag Formation. These workers consider the "Rocks of Island Falls" to occupy the same stratigraphic position as the Allsbury Formation.

The fossil locality on Interstate 95, therefore, helps to establish the age range of a basal Silurian slate and sandstone sequence extending southwestward from the vicinity of Smyrna Mills (Smyrna Mills Formation of Pavlides and Berry (1966) and Pavlides (1968)) through the southern half of the Island Falls quadrangle into the Shin Pond, Stacyville and Sherman quadrangles (Allsbury Formation and "Rocks of Island Falls"). The limited paleocurrent information available at these exposures is consistent with transport of the coarser material (fine sandstones and siltstones) from the west (or northwest) toward the east (normal to the trends of the Early Silurian facies) in general agreement with the models proposed by Neuman (1967, p. 35-36) and Ekren and Frischknecht (1967, p. 22).

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Geology of Limestone Hill, Stratton Quadrangle, Somerset County, Maine

Kost A. Pankiowsky¹

INTRODUCTION

Location and Topography

Limestone Hill is located north of Flagstaff Lake in the northeast corner of the Stratton Quadrangle, Maine (Fig. 1). Limestone Hill is a ridge, with highest elevation above sea level of 1990 feet, elongated in the north to south direction, and having gentle slopes to the north and west, steeper slopes to the south, and very steep slopes (in places with an inclination of 50°) to the east. Exposures of bedrock are numerous along the top of the ridge, at elevations above 1900 feet, and are poor lower on the slopes. Small outcrops are also found along the east side of the dirt road south and southwest of Limestone Hill.

Previous Work

Limestone Hill, has been known geologically since 1874, when C. H. Hitchcock and J. H. Huntington did their reconnaissance of north-

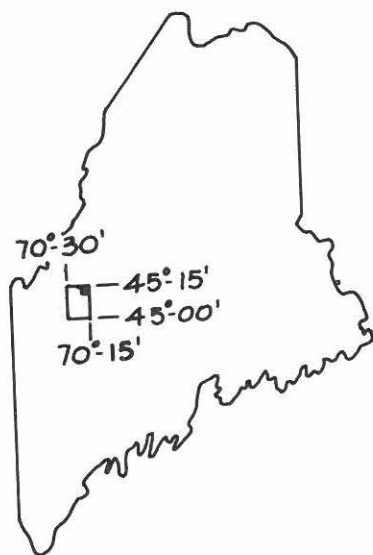


Fig. 1. Map of Maine showing location of the Stratton Quadrangle, and within it, the location of Limestone Hill.

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western Maine. In their report (Hitchcock and Huntington, 1874, p. 212) they report the existence of a band of Helderberg limestone outcropping on an island in Flagstaff Pond and another band on the west peak of Flagstaff Mountain (Limestone Hill), the latter containing fossils which "are very obscure". The locality was recently mentioned by Boucot and Heath (1969, p. 46) who list the fossils collected on Limestone Hill by the writer in 1958, and who give reasons for assigning to them a Late Llandovery age.

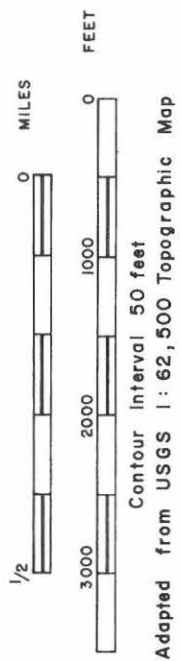
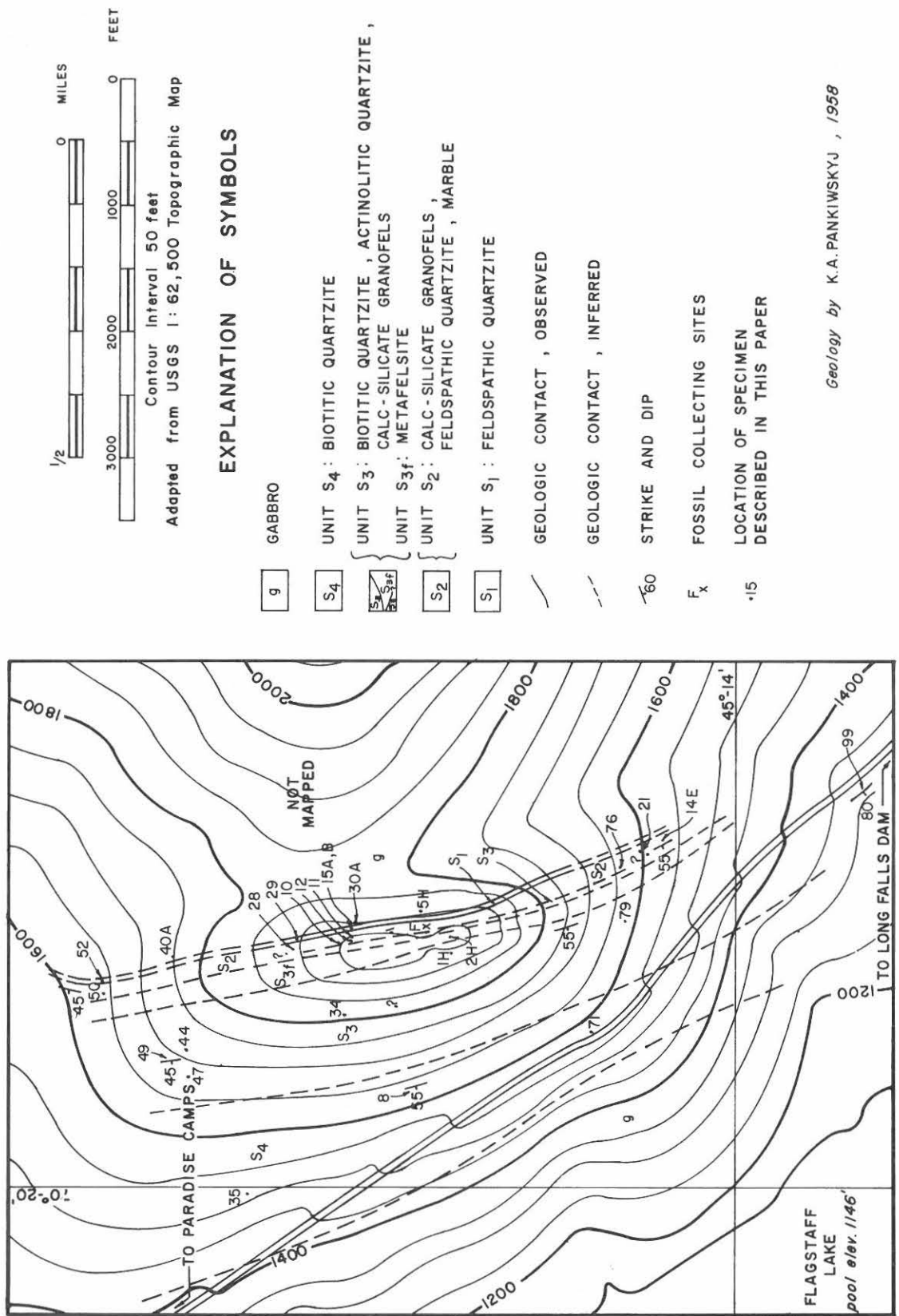
Acknowledgments

The field work and fossil collecting was carried out by the writer in the summer of 1958, when he was employed as field assistant to A. Griscom of the U. S. Geological Survey. Thanks are extended to Mr. Griscom for his advice during the mapping and for permission to publish this paper. The rock samples and thin sections examined by the writer are the property of the U.S. Geological Survey. The writer is grateful to Drs. W. F. Brace and A. J. Boucot, for visiting him in the field and offering useful suggestions. Additional thanks are extended to Boucot who served as the writer's advisor during the preparation of his bachelor's thesis at the Massachusetts Institute of Technology (Pankiowsky, 1959), and who identified and dated the fossil material.

STRATIGRAPHY AND LITHOLOGY OF THE METASEDIMENTARY ROCKS

General Statement

The stratigraphic section on Limestone Hill has been subdivided into four metasedimentary units of Silurian age, numbered from one to four in order of decreasing age, and one metamorphosed igneous unit, S_{3f}, mapped as part of metasedimentary Unit S₃. (Fig. 2). This section is as follows:



EXPLANATION OF SYMBOLS

- g GABBRO
- S4 UNIT S4 : BIOTITIC QUARTZITE
- S3f UNIT S3 : BIOTITIC QUARTZITE , ACTINOLITIC QUARTZITE ,
CALC - SILICATE GRANOFELS
- S3 UNIT S3f : METAFELSITE
- S2 UNIT S2 : CALC - SILICATE GRANOFELS ,
FELDSPATHIC QUARTZITE , MARBLE
- S1 UNIT S1 : FELDSPATHIC QUARTZITE
- GEOLOGIC CONTACT , OBSERVED
- - - GEOLOGIC CONTACT , INFERRED
- $\frac{60}{\text{---}}$ STRIKE AND DIP
- F_x FOSSIL COLLECTING SITES
- 15 LOCATION OF SPECIMEN DESCRIBED IN THIS PAPER

Geology by K.A. PANKIWSKYJ , 1958

Fig. 2. Geologic Map of Limestone Hill, Somerset County, Maine.

No. of Unit	Lithology	Thickness (in feet)
S ₄	Brownish-grey feldspar-rich biotitic quartzite	700-800
S ₃	Purplish biotitic quartzite and/or pale greenish-grey actinolitic quartzite with pods of light grey-green calc-silicate granofels; flaggy pale purplish-green quartzite;	350-450 25- 35
S _{3f}	grey metafelsite	60-130
S ₂	Bedded calc-silicate granofels and feldspathic quartzite, grey marble; greenish-grey fossiliferous calc-silicate granofels, grey marble	50- 75 50- 75
S ₁	Massive light greenish-grey diopside-and actinolite-bearing feldspathic quartzite	10- 20
TOTAL		1245-1585

On the very steep east slope of Limestone Hill, this section is cut off at its base by a concordant intrusive of hornblende gabbro. On the dirt road southeast of the hill, the metasedimentary rocks are folded and fractured and intruded by apophyses of gabbro, granite, and igneous rocks of intermediate composition. To the west and north, the slopes of Limestone Hill are devoid of outcrop and thus the relationship of the metasedimentary rocks to other rocks is not known.

Unit S₁

The oldest exposed stratigraphic unit, Unit S₁, is a light grey to slightly greenish-grey, massive, diopside- and actinolite-bearing feldspathic quartzite, weathering typically to a smooth greyish-tan color. Diffuse green bands rich in diopside occur locally in the feldspathic quartzite. These bands are about 5 mm thick and are spaced from 5 mm to 20 mm apart. Unit S₁ is cut off at the base by a sill-like body of hornblende gabbro. Toward the top of the unit the rock becomes richer in calcite, due to the appearance of fossil fragments, and grades over a stratigraphic distance of five to ten feet into Unit S₂. The stratigraphic thickness of Unit S₁ ranges from 10 feet to 20 feet.

Quartz, plagioclase, and orthoclase make up the bulk of the rock in a tight network of an-

hedral grains. Their sizes range from 0.05 mm to 0.12 mm and average 0.07 mm. The quartz and plagioclase (oligoclase on the basis of relief against balsam and poorly developed twinning after the albite law) are clear; the orthoclase (identified on the basis of moderate negative relief and $2V_x$ of about 60°), however, is clouded with minute inclusions. Actinolite typically forms scattered randomly oriented bladed grains up to 0.15 mm in length, though some is found in the form of radiating clusters. Most grains have ragged terminations, and some appear like bundles of optically continuous tightly packed needles. Diopside and clinzoisite show a blocky habit and are up to 0.12 mm in size. Sphene forms rounded grains 0.05 mm in diameter, commonly with an opaque inclusion in the center; and also as glomerocrysts up to 0.10 mm in size composed of many minute grains. Accessories are magnetite, zircon, rutile, apatite, and calcite.

Unit S₂

The second oldest stratigraphic Unit, S₂, is subdivided into two parts: the lower part, which contains fossils, and the upper mineralogically similar part, which is barren. The lower part consists of interbedded greenish-grey fossiliferous calc-silicate granofels and light grey sugary marble. The fossil-bearing granofels contains up to 20 percent of fossil fragments composed of polycrystalline masses of calcite grains up to 1 mm in size, showing extensive polysynthetic twinning. In addition, calcite is present as twinned single large grains, and as small untwinned grains in the matrix. This matrix is composed of a tightly packed network of minerals whose relative proportions to one another differ from section to section, and even in parts of one section. Diopside forms scattered grains as well as dense masses of blocky or round grains 0.02 mm to 0.04 mm in size. Polysynthetically twinned wollastonite is present in pure masses of fibrous crystals up to 0.3 mm long which are oriented parallel to the bedding in lenses up to 0.5 wide and 10 mm long. Wollastonite is also found as randomly oriented smaller grains associated with other minerals. A common association is as a jacket surrounding a fossil fragment. Grossularite is found in grains 0.02 mm in diameter, typically forming local glomerocrysts, some of which measure up to one centimeter in diameter. Idocrase forms large poikilitic grains up to 3 mm across.

Sphene is found both as opaque-centered round grains and as glomerocrysts. Clinozoisite, magnetite, and zircon are common accessories. Deeply clouded, low birefringent grains, studded with tiny inclusions with a high relief are thought to be orthoclase, due to their similarity to the orthoclase in Unit S₁. Quartz and oligoclase are found in variable amounts as clear anhedral grains ranging in size from 0.01 mm to 0.1 mm.

The above described calc-silicate granofels is associated with bands, lenses, and beds of sugary marble. In thin section this is seen to contain a few percent fossil fragments, but most of the calcite occurs as small untwinned grains ranging in size from 0.02 mm to 0.2 mm. Diopside, wollastonite, quartz, plagioclase, and traces of sphene and magnetite are found scattered between the calcite grains. Grossularite is also present as large poikilitic grains, up to 2 mm in diameter, enclosing grains of calcite, diopside, and wollastonite. Spongy idocrase forms grains up to 1 mm in diameter.

The upper part of Unit S₂ does not contain recognizable fossil fragments. It is composed of interbedded light grey marble and a rock made up of inter-layered bands of diopside- and actinolite-bearing feldspathic quartzite and of a calcareous calc-silicate granofels. All these rock types are mineralogically similar to types already described: the marble to the marble in the lower part of Unit S₂; the calc-silicate granofels to the fossil-bearing granofels, provided one takes out the fossil fragments; and the quartzite to the rock comprising Unit S₁. The upper part of Unit S₂ occupies a ridge at the top of the steep east-facing slope of Limestone Hill. Within this ridge are found several depressions containing angular blocks of quartzite, granofels, and marble. Blocks of the last have been found up to two feet in diameter. It is possible that these depressions represent localities from which marble was obtained to be burned for lime in the early days of colonial settlement of the area.

The lower, fossiliferous part of Unit S₂ ranges in thickness from 50 feet to 75 feet, and the upper barren part also from 50 feet to 75 feet. Near the top of Unit S₂, both marble and calc-silicate granofels become scarce, whereas the actinolite- and diopside-bearing feldspathic quartzite becomes slightly reddish in color due to the appearance of bands containing biotite.

The overlying Unit S₃ is mapped with the disappearance of layers of marble.

Unit S₃

Unit S₃, ranging in thickness from 350 feet to 450 feet, comprises the bulk of the well-exposed section on Limestone Hill. Within it are mapped: a 25-foot to 35-foot thick member characterized by a pronounced flaggy appearance, and a 60-foot to 130-foot thickness of metafelsite. Unit S₃ is characterized by the presence of light greenish-grey diopside-rich calc-silicate granofels in a matrix of darker greenish-grey actinolite-bearing quartzite or reddish biotite-bearing quartzite. The diopside-rich rock type ranges in geometric form from thin streaks (thin sheets in three dimensions) to pods up to 3 cm thick and 25 cm long, but is also found as poorly defined, irregular blob-like patches. No pattern was found to delineate the exposures of actinolite-bearing from the biotite-bearing quartzite.

The light greenish-grey diopside-rich rock contains from 10 percent to 40 percent diopside in blocky grains averaging 0.05 mm in size, but in places composed of glomerocrystic masses composed of grains up to 0.4 mm in size. Up to 4 percent total of actinolite, biotite, chlorite, and clinozoisite accompany the diopside in grains up to 0.07 mm in longest dimension. Pods and bands thicker than 5 mm commonly contain grains of grossularite 0.02 mm to 0.04 mm in diameter. Quartz and labradorite form grains 0.05 mm in size in a ratio of 2:3 to 3:2.

The dark greenish-grey actinolite-bearing matrix rock contains up to 15 percent actinolite in lath-like grains up to 0.07 mm in length. A few percent of diopside, clinozoisite, biotite, and chlorite are commonly encountered. Quartz and labradorite appear in roughly the same mutual ratio as in the diopside-rich bands. Sphene is a common accessory.

The reddish biotite-bearing matrix rock contains up to 20 percent biotite in flakes up to 0.08 mm in diameter and 0.04 mm thick. As much as 10 percent actinolite is present in some samples, as well as several percent of diopside, clinozoisite, and chlorite. Quartz and labradorite appear as in the other types in this unit.

The strikingly good looking color-banded pods, which crop out next to the dirt auto road south of Limestone Hill are part of Unit S₃ (Specimen 99 as shown on Fig. 2). In thin section, the color banding clearly reflects the

mineralogical zoning in the pods. The centers of the pods are dominated by grossularite, with smaller quantities of diopside, quartz, and plagioclase—in hand specimen this region is brown. Around this, diopside predominates, with smaller quantities of grossularite, clinzoisite, quartz, and plagioclase—in hand specimen this zone is light green. Around this, quartz and plagioclase comprise a total of 85 percent of the rock, with the remainder made up of scattered grains of chlorite, actinolite, diopside, and clinzoisite—this zone is pale grey-green. This zone is followed sharply by the purplish biotite-bearing quartzite.

Noted about half-way within Unit S₃ is a stratigraphic thickness of rock ranging from 25 to 35 feet, which, though mineralogically similar to the bulk of Unit S₃, is differentiable in the field on account of a well-developed flagginess. The rock breaks easily into flaggs, ranging in individual thickness from 5 mm to 2 cm, with rusty stained surfaces. In thin section (Specimen 49, Table 1) the development of the flaggs is seen to be controlled by thin zones of magnetite crystals, some of which have become oxidized to hematite. The rock is rust stained to a distance of 1 mm from these oxidized bands.

Unit S₄

Above the green pod-bearing quartzite comprising Unit S₃ is mapped a section, measuring upward from 750 feet in thickness, of feldspar-rich biotitic quartzite. The bulk of the section is poorly exposed on the gently sloping west side of Limestone Hill. Outcrops of this unit, are, however, also found along the east side of the dirt road skirting the west side of Limestone Hill. Here the metasediments are strongly deformed, fractured and folded, and intruded by apophyses of gabbro, granite, and rocks of intermediate composition, all presumably derived from the main bodies of intrusive rock which underly the topographic lows of Flagstaff Lake.

In hand specimen, rocks of Unit S₄ are distinguishable from those of the lower units because of their larger grain size. They are grey to slightly brownish-grey in color, and individual grains of quartz, plagioclase, and scattered flakes of biotite are readily discernible with the naked eye. Graded beds, about 1 cm thick, can be seen in some of the exposures along the road. In thin section, quartz and oligoclase form a

compact recrystallized network with the individual grains 1.0 mm to 1.5 mm in diameter. Scattered randomly between them are grains of biotite, a large number of which are about 0.5 mm long and only 0.02 mm in the *C* crystallographic direction. Thicker grains are also found, attaining widths in the *c* direction of up to 0.25 mm, and these commonly form composite grains intergrown with muscovite. Rare grains of epidote are found as weakly pleochroic crystals up to 1.5 mm long and 0.5 mm wide. Magnetite is present in scattered grains 0.07 mm in size. Accessories include zircon, apatite, and garnet.

IGNEOUS ROCKS

Gabbro

Unit S₁ is cut off on its lower side by a body of gabbro, which contains layering roughly parallel to the bedding in the metasedimentary section on Limestone Hill. This layering, which is well seen in outcrop, is due to the presence of coarser-grained bands in a rock which is typically fine-grained. These bands are up to 1 cm thick and are made up of scattered plagioclase laths, typically about 3 mm by 1 mm (but attaining sizes of up to 1 cm by 3 mm in some of the coarsest layers), set in a mass of pyrobole whose color varies from dark green to black. The finer grained type is diabasic in texture, with the individual feldspar laths and pyrobole masses about 1 mm in length and 0.3 mm in width. In thin section, the coarse and fine types are seen to be of the same bulk mineralogy, differing only in grain size. The plagioclase is in well-formed laths, practically all being twinned according to albite, Carlsbad, and pericline laws. The grains are strongly and continuously zoned from a core of An₆₉ to a rim of An₄₅. Some grains have sausseritized cores, and relatively fresh rims. Filling the intergranular spaces between the plagioclase grains are single grains and, much more commonly, aggregates of grains of a weakly pleochroic hornblende. Some of the larger grains contain remnants of a very pale green augite. It is presumed that most if not all of the hornblende is a replacement of augite. Some of the hornblende is seen altering to biotite which, in several samples, is itself being replaced by a pale chlorite. Irregular masses of pyrite are found with the hornblende. Accessories include zircon, sphene, apatite, and epidote.

**TABLE I MODES OF METASEDIMENTARY ROCK SPECIMENS
FROM LIMESTONE HILL
(based on visual estimates)**

	15A	15B	29	52	10	14E	1	11	12	21	28	50	76	1Ha	
Quartz	30	42	40	45	4	1		25	50	40	30	10	5	20	15
Plagioclase	10	10			—	—									
			54	48											
K feldspar	50	40			—	tr	?	?	2	1?	½?	?	3?	—	
Biotite	—	tr	—	—	—	—	—	—	—	—	—	—	—	—	
Chlorite	—	—	—	—	—	—	—	—	—	—	—	—	—	—	
Clinozoisite	—	tr	tr	tr	—	—	½	tr	—	tr	?	?	tr	tr	
Diopside	5	2	3	3	1	2	55	26	31	30	50	35	20	83	
Actinolite	2	5	2	3	—	1	tr	tr	—	—	—	—	—	1	
Wollastonite	—	—	—	—	1	—	½	12	12	15	tr	30	35	—	
Idocrase	—	—	—	—	—	1	—	—	1	—	—	—	1	—	
Grossularite	—	—	—	—	—	4	2	tr	—	½	12	20	tr	—	
Calcite	½	½	½	½	93	90	15	10	15	20	25	10	5	—	
Sphene	½	½	½	½	tr	tr	½	tr	tr	tr	tr	tr	tr	tr	
Opaque oxide	tr	tr	tr	tr	½	tr	½	½	tr	tr	tr	tr	tr	tr	
Zircon	tr	tr	tr	tr	—	—	—	—	—	—	—	—	—	—	
Apatite	—	—	—	—	—	—	—	—	—	—	—	—	—	—	

Explanation of Table I

Specimen No.	Unit	Rock Type
15A	S ₁	Actinolite and diopside-bearing feldspathic quartzite
15B	"	"
29	"	"
52	"	"
10	S ₂	Marble
14E	"	"
1	"	Calc-silicate granofels
11	"	"
12	"	"
21	"	"
28	"	"
50	"	"
76	"	"
1Ha	S ₃	Diopside-rich calc-silicate granofels - pod
1Hb	"	Biotite-rich quartzite - matrix
7a	"	Diopside-rich calc-silicate granofels - pod
7b	"	Biotite-rich quartzite - matrix
34a	"	Diopside-rich calc-silicate granofels - pod
34b	"	Actinolitic quartzite - matrix
44a	"	Diopside-rich calc-silicate granofels - pod
44b	"	Actinolitic quartzite - matrix
47a	"	Diopside-rich calc-silicate granofels - pod
47b	"	Actinolite-rich quartzite - matrix
79a	"	Diopside-rich calc-silicate granofels - pod
79b	"	Biotite-rich quartzite - matrix
49a	"	Flaggy actinolite-rich quartzite - composition between fracture surfaces
49b	"	" - composition at fracture surface
8	S ₄	Biotitic feldspar-rich quartzite
35	"	"
71	"	"

**TABLE I MODES OF METASEDIMENTARY ROCK SPECIMENS
FROM LIMESTONE HILL
(based on visual estimates)**

1Hb	7a	7b	34a	34b	44a	44b	47a	47b	79a	79b	49a	49b	8	35	71
	58	45		48	28	60			16	45	40	40	33	50	35
73			30				49	79							
	25	20		40	22	42			8	35	40	20	60	44	50
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
20	tr	20	—	½	tr	½	—	—	—	12	2	—	4	5	4
—	tr	½	—	3	tr	tr	—	—	—	—	1	—	tr	1	—
—	tr	—	½	—	tr	tr	tr	—	—	—	½	tr	—	—	—
2	15	—	40	—	40	—	40	25	76	4	—	—	—	—	—
4	—	15	1	5	—	5	1	20	—	1	12	30	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
—	—	—	25	2	—	—	—	—	—	—	½	3	—	tr	—
—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
tr	tr	tr	tr	tr	½	½	tr	½	tr	tr	—	—	—	—	—
—	tr	—	—	tr	—	—	—	—	—	—	1	8	½	tr	½
—	tr	—	—	tr	tr	tr	—	tr	—	—	tr	—	tr	tr	tr
—	—	—	—	—	—	—	—	—	—	—	—	—	½	—	—

**TABLE II MODES OF IGNEOUS ROCK SPECIMENS
FROM LIMESTONE HILL
(based on visual estimates)**

	5H	30a	40a	2H
Quartz	—	—	—	25
Plagioclase	44	41	45	50
"Myrmekite"	—	—	—	20
Augite	—	1	—	—
Hornblende	45	45	54	—
Biotite	—	2	tr	2
Chlorite	—	½	—	1
Clinzoisite	tr	—	—	—
Epidote	tr	tr	tr	—
Sphene	tr	—	—	—
Sulphide	½	½	½	½
Zircon	—	tr	—	—

Explanation of Table II

Specimen No.	Unit	Rock type
5H	g	Hornblende gabbro
30a	"	"
40a	"	"
2H	S _{3f}	Metamorphosed felsic volcanic

Unit S₃f

Starting at 30 to 40 feet above the base of Unit S₃ there is a dearth of outcrop for a distance perpendicular to strike of 60 to 130 feet. In this interval, which comprises a shallow groove-like valley along the entire ridge of Limestone Hill, is found heavy float of a metamorphosed igneous rock. Because of lack of exposure of contacts of this rock with Unit S₃, and because of recrystallization due to metamorphism, the writer was not able to establish whether this rock is a premetamorphic sill, flow, of pyroclastic deposit. In hand specimen can be seen squarish white megacrysts of feldspar and clear round megacrysts of quartz ranging in size up to 5 mm; shiny flakes of biotite up to 0.5 mm in size; and a few crystals of pyrite up to 1 mm in size, each surrounded by a rust colored halo. The matrix appears as a crystalline light grey mass resembling a mixture of finely powdered salt and pepper. In thin section this matrix is seen to be composed of quartz, oligoclase, and myrmekite, rounded, embayed, and intergrown in an amoebalike fashion. Individual grains range in size from 0.03 mm to about 0.5 mm, though a few grains of myrmekite attain diameters of 0.8 mm. Larger grains of oligoclase have better preserved crystal outlines, and are presumably phenocrysts. The oligoclase, whether in the phenocrysts or in the matrix, is crowded with fine sericite; the oligoclase in the myrmekite, however, is relatively clear. Strongly pleochroic biotite flakes are found in single grains up to 0.1 mm in length, and in glomerocrysts which reach 1.5 mm in diameter. Intergrown magnetite (?), muscovite, and chlorite showing vestiges of biotite appear as ragged glomerocrystic clumps up to 0.2 mm in size.

GEOLOGIC AGE

Fossils were collected by the writer from three localities in the lower part of Unit S₂. The thickness of the entire Unit S₂ in the region of the collections is about 100 feet, and the three localities are respectively 17 feet, 35 feet, and 50 feet from the bottom of the unit. Boucot (1969, p. 46) lists the most important forms in the three localities, and assigns to

them an age span in the range of C₃ - C₆ of the Upper Llandovery. However, he states that an Early Wenlock age cannot be completely ruled out.

STRUCTURE

The metamorphic section on Limestone Hill appears to be a part of a gently folded anticline plunging approximately due west at 50°. The attitude of the beds ranges from N20°E - 45°NW in the northern part of the hill, through about N15°W - 50°SW along the main ridge, to N40°W - 80°SW south of the hill, in the creek just off the dirt road (Specimen Location No. 99).

METAMORPHISM

The assemblage wollastonite - diopside - grossularite - idocrase - quartz is characteristic of the pyroxene hornfels facies of metamorphism (Turner, 1968). The presence of the three phases: wollastonite - quartz - calcite indicates either a lack of equilibrium in a system open to CO₂, because this would require a definite partial pressure of CO₂, or it indicates a system from which CO₂ is not free to escape. The first possibility is the more likely because a gabbroic intrusion is only twenty feet from the base of Unit S₂, which contains the assemblage. The gabbro is partially hydrated and, it appears to the writer, could absorb the CO₂ from the reaction calcite + quartz = wollastonite + CO₂, if there was sufficient time.

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The Fish River Lake Formation and its Environments of Deposition

Gary M. Boone¹

INTRODUCTION

As part of a former mapping program in northern Aroostook County one of the objectives was to clarify the stratigraphic relationship of rocks along the east margin of the Seboomook slate in the Fish River Lake quadrangle, since earlier studies (see below) produced conflicting opinions on structural relations and stratigraphic sequence. Locations of exposures along the contact led me to extend the mapping into the northwest part of the Winterville, and south-central part of the Eagle Lake quadrangles (Fig. 1). Furthermore, the interest generated by plant fossils associated in the same formational unit with marine faunas, and the varied depositional conditions recorded, led me to study, partly in detail, and partly in reconnaissance, (Boone, 1958) the rocks here designated as the Fish River Lake Formation.

In the intervening decade, others have studied parts of this sequence to the northeast of the map area in Fig. 2, and rocks in adjacent units to the southeast (e.g., Pavlides and others,

1964; Coombs, Horodyski, and Naylor, 1970). Extensive lumbering operations have opened new roads, creating new exposures, and more significantly, new fossil localities have been found (Pavlides and others, *ibid.*). The Fish River Lake sequence, however, has not been restudied as such, and therefore the main purpose of this report is to re-establish the rocks of this sequence as a formational unit, and secondly, to provide a summary of geological interpretations relating to the area up to the beginning of the 1960's. The observations, particularly as they were taken during a year of exceptionally low lake-levels, should prove useful to those in the future who may be lured to study these rocks in the much greater detail that they deserve.

I am indebted to John Rodgers for his enthusiastic interest during the period of study, and to Arthur Boucot for identifying most of the fossils. The kindness and interest on the part of the McNally's at Fish River Lake are warmly remembered.

Previous Work

The earliest known geological reconnaissance of northern Maine was made by James T. Hodge, who was assistant to C. T. Jackson, state geologist. At Jackson's request, Hodge travelled from north central Maine to the shore of the St. Lawrence River in 1837. He observed outcrops of intercalated slate, limestone, and graywacke from the headwaters of the Allagash River to its confluence with the St. John. He noted that slate is the most abundant (Jackson, 1838, p. 63-64). His observations were made in the western part of the Allagash synclinorium as here defined. In 1861, Hitchcock described in scarcely more detail the geographic distribution of rocks in northern Maine. Most of the facts are contained in A. S. Packard's report to Hitchcock, (Hitchcock, 1861, p. 420-425). Packard listed the fossils from the lower Helderberg limestone of Square Lake, and concluded from his reconnaissance that lower Helderberg rocks

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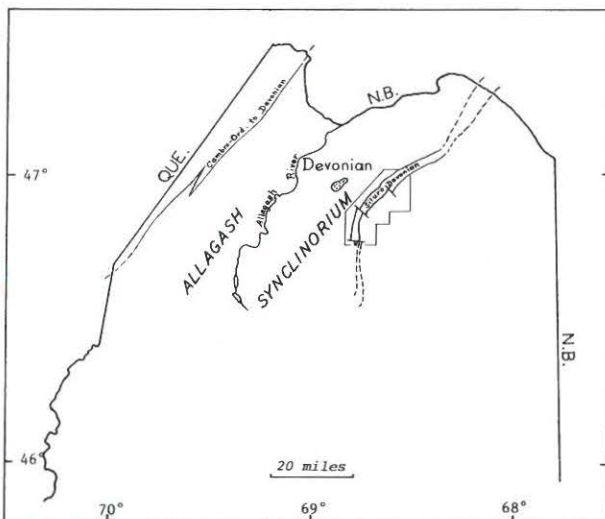
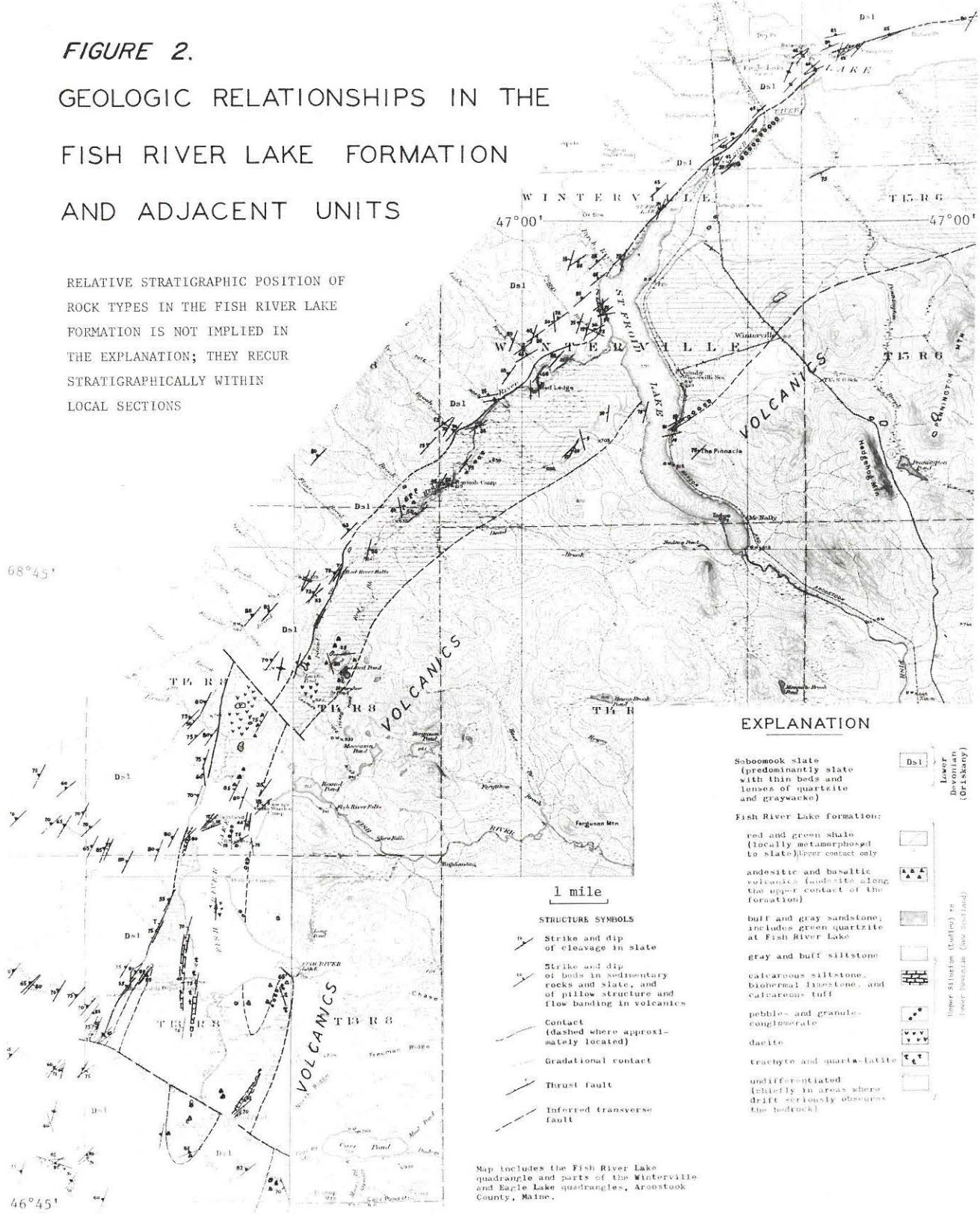


Fig. 1. The Allagash synclinorium, and ages of related rocks in northern Maine. The map area of this report, describing the Fish River Formation, is outlined. Irregular dashes: Deboullie igneous complex (Boone, 1962). Western limit of Devonian rocks in the synclinorium from Boudette and others (1967).

FIGURE 2.
GEOLOGIC RELATIONSHIPS IN THE
FISH RIVER LAKE FORMATION
AND ADJACENT UNITS

RELATIVE STRATIGRAPHIC POSITION OF
 ROCK TYPES IN THE FISH RIVER LAKE
 FORMATION IS NOT IMPLIED IN
 THE EXPLANATION; THEY RECUR
 STRATIGRAPHICALLY WITHIN
 LOCAL SECTIONS



EXPLANATION

- Schoonook slate**
 (predominantly slate with thin beds and lenses of quartzite and graywacke) Dsl
- Fish River Lake formation:**
 - red and green shale (locally metamorphosed to slate); lower contact only
 - andesitic and basaltic volcanics (andite along the upper contact of the formation)
 - buff and gray sandstone; includes green quartzite at Fish River Lake
 - gray and buff siltstone
 - calcareous siltstone, biohermal limestone, and calcareous tuff
 - pebbles- and granule-conglomerate
 - dacite
 - trachyte and quartz-tuffite
 - undifferentiated (chiefly in areas where drift seriously obscures the bedrock)

- 1 mile**
- STRUCTURE SYMBOLS**
- Strike and dip of cleavage in slate
 - Strike and dip of beds in sedimentary rocks and slate, and of pillow structure and flow banding in volcanics
 - Contact (dashed where approximately located)
 - Gradational contact
 - Thrust fault
 - Inferred transverse fault

Map includes the Fish River Lake quadrangle and parts of the Winterville and Eagle Lake quadrangles, Aronstook County, Maine.

extend southwest to Portage Lake, that these are overlain by clay slate to the northwest, and that beds generally dip moderately to steeply northwest. Hitchcock apparently did not visit this part of Maine, and the details of northern Maine geology on his map of 1862 were probably derived from Packard's description. The slate immediately to the west of the Helderberg rocks is designated as Devonian on Hitchcock's map, and the large area of "clay slate" west of the Devonian slate is not assigned a tentative age. Keith (1933) considered the slate Silurian in age, and the more arenaceous and volcanic rocks bounding the slate on the southeast, Devonian. It has generally been recognized since 1956 that Packard's early interpretation was essentially correct.

Regional Relations

The principal structural feature of regional importance established by previous work and by part of my own mapping is a broad synclinorium (the Allagash synclinorium, fig. 1) of Devonian slate that trends northeast across central and western Aroostook County.

Boucot (1961) established a lower Devonian, Oriskany age for the Seboomook Formation in the Moosehead Lake region, and termed the structural setting in which this and other chiefly Devonian units are preserved, the Moose River synclinorium. These rocks, dominantly expressed by the slate lithology of the Seboomook, were traced northward by Boucot into the Allagash and St. John River drainage basins where fossils that also indicated a lower Devonian age were found (Boucot, 1956, personal communication). These findings established a continuity of age, along strike to the northeast, with the rocks in the vicinity of Ft. Kent where Nylander (1940) reported the occurrence of brachiopods of Oriskany age. This slate terrain, including the slate highlands west of the Fish River chain of lakes, comprises the most extensive sedimentary sequence in northwestern Maine. It is flanked both to the northwest and to the southeast by well-delineated structural belts containing older rocks. Because the northeastward structural extension of the Moose River synclinorium bifurcates around a series of structural culminations exposing older rocks east of Churchill Lake (Doyle, 1967), the northwesterly extension of this structure into the Allagash region is here referred to as the Allagash synclinorium, named for the Allagash

River which follows its trend in the central part of the structure.

Stratigraphic Sequence

Three major rock-stratigraphic units occur within the map area shown in Figures 1 and 2.

Seboomook Slate

Pelitic to psammitic slate, including minor graywacke and quartzite; largely unfossiliferous.

Fish River Lake Formation

Largely unmetamorphosed sandstone and siltstone, in part calcareous. In part contains marine fauna; in part, flora consisting of plant stems. Presence of granule- and pebble-conglomerates. Interbedded flows of mafic and felsic volcanics.

Volcanics of the Winterville area

Predominantly mafic volcanics; minor felsic flows and fine-grained arenaceous, calcareous sedimentary rocks. Rare cherty layers, and medium-grained dioritic rocks.

The name *Seboomook Slate* is used for the slate terrain of the present map area on the basis of (1) physical continuity with the type Seboomook of the Moosehead Lake region (Boucot, 1961); (2) the occurrence of lower Devonian fossils in the slate along the Allagash River (Boucot, personal communication, 1956), (3) the occurrence of the Oriskany brachiopod *Leptocoelia flabellites* in slate near the village of Fort Kent on the St. John River (Nylander, 1940), and (4) the presence of Helderberg (New Scotland) fossils in siltstone below the base of the slate sequence at Fish River Lake.

I propose the name *Fish River Lake Formation* for the stratigraphic succession of rocks between the slate and the predominantly volcanic sequence. The association of fossiliferous, continental and marine arenites, calcarenites, and rudites with basaltic and felsic flows renders it a distinctive mapable unit whose age ranges from upper Silurian (Ludlow) to lower Devonian (Helderberg).

The volcanic sequence of the Winterville area was studied only in reconnaissance; hence no detailed description is given below. The relative abundance of mafic and felsic volcanics is imperfectly established, but basaltic or an-

desitic volcanics appear to prevail. Pervasive and closely spaced fractures, and the general absence of flow-banding and vesicular structure made it difficult to determine attitudes of flows. Coarse-grained dioritic rock occurs on Pennington Mountain at the east end of the map area, and is surrounded by outcrops of basaltic flows but the structural relationship between the two could not be determined. Both are less fractured and appear younger than the surrounding extensively deformed volcanics and interspersed strata. From its outcrop pattern, the dioritic rock seems likely to represent part of a small pipe-like injection or the coarse-grained facies of a volcanic conduit.

FISH RIVER LAKE FORMATION

General Characteristics

The relatively non-resistant rocks of this formation crop out modestly from beneath a widespread blanket of glacial deposits in a belt from Fish River Lake northeast to St. Froid and Eagle Lakes. The rocks are mainly clastic and volcanoclastic arenites with less common rudites and calcarenites, within which a variety of volcanic rocks are intercalated. The rock sequence forms a distinctive unit that separates the Seboomook Slate from the volcanic sequence of the Winterville area. The finer-grained sedimentary rocks in this unit are generally thin-to medium-bedded; the lava flows and conglomerates are locally very thick. The entire sequence is sparsely fossiliferous, containing marine faunas and a poorly preserved continental flora.

The formation is named for the outcrops of its typical rock types within and around the shores of Fish River Lake. The outcrops at Fish River Lake are not as continuous as others elsewhere in the unit but, whereas exposures elsewhere are restricted to one or two rock types, nearly all the lithologic variations of the formation are exposed at Fish River Lake. The name is considered appropriate for another reason: St. Froid and Eagle Lakes, along whose shores the formation is also exposed, belong to a chain of lakes locally known as the Fish River chain.

Composition

The most abundant rocks are thin-bedded arenites that range in texture from coarse siltstone to coarse sandstone. They are predominantly quartzose but arkosic and graywacke va-

riants occur locally. Granule- and pebble-conglomerates also contain detrital feldspar. Carbonate-bearing rocks, more restricted in distribution, consist of calcareous and dolomitic sandstone and siltstone, biohermal reef rock, and calcareous tuff.

Siltstone and sandstone: Both types range in color from buff to dark gray-green. Those of buff or gray color are commonly cemented by a mixture of quartz, calcite, and hematite within which black opaque material (not magnetic or of metallic luster, and therefore probably carbonaceous) is common. Dolomite is rare. The dark gray-green rocks are more thoroughly indurated, and are cemented chiefly by quartz. Although the source of the green color is probably ferrous iron (in sulfide?) it is not evident in thin section.

The most noteworthy characteristic of these rocks is the angularity of their constituent grains. The appearance of the quartz clasts in thin sections gives the impression that angularity is independent of grain size, and thus that the fine-grained, as well as the coarse-grained rocks, were rapidly transported and deposited. Most of the grains are quartz; plagioclase and garnet are present in small quantities. Staining to detect the presence of potassic feldspar revealed only a few grains per thin section except in considerably arkosic specimens.

Bedding is evident in the sandstones as color changes and widely spaced parting, and more commonly in the siltstones as closely spaced parting approaching a coarse fissility. Cross-beds are tabular in shape, and truncated sharply by superposed parallel beds.

Measurement of the attitude of cross-beds and their angular relation to parallel bedding at an outcrop in Red River, near the mouth of Labbe Brook, indicates a northeast up-current direction. Similar measurement of a crude lineation produced by alignment of plant stem fossils in coarse siltstone that crops out on the southeast bank of Fish River Thoroughfare, also indicates a northeast current source. The size of plant stems, the frequency of their occurrence, and the perfection of their preservation decrease from northeast to southwest along the strike of the formation, from the vicinity of the northwest shore of St. Froid Lake to Fish River Lake.

Siltstone and sandstone are only rarely fossiliferous. The flora of rather poorly preserved (commonly macerated) plant stems occurs in

buff or gray sandstone and siltstone. The stems are preserved in the form of carbonaceous molds and impressions. As may be expected, the stems are larger and better preserved in sandstone, implying that transportation was of shorter duration for the detrital material of the coarser-grained rocks. The classification and inferred age of the flora is considered following the section dealing with the invertebrate fauna of the formation, and its role in the interpretation of the environment of sedimentation is considered under a separate heading.

I did not observe marine fossils in the buff or gray siltstones and sandstones in which the plant stem flora abounds. Faunas consisting of brachiopods, gastropods, trilobites and corals occur, however, in dark calcareous siltstone, calcareous tuff, dark gray-green sandstone and granule-conglomerate. These rock types crop out chiefly southwest of Red River Falls (Winterville quadrangle), and are common at Fish River Lake (Fig. 3). At the northeast corner of the map area (Fig. 3), fragmented brachiopods and a tetracoral were found in pebble-conglomerate on the north shore of Eagle Lake, but they were too poorly preserved to permit even generic determination.

Conglomerate: The matrices of both pebble- and granule-conglomerates are quartz - rich. Commonly possible-conglomerate is cemented by calcite; granule-conglomerate, by quartz. The matrix of the granule-conglomerate is, in fact, orthoquartzite. The coarse fragments of both types consist of quartz (and jasper), basaltic and felsic volcanics, pink potassic feldspar, granite, and pegmatite. Fragments of the latter two types are rare.

Both pebble- and granule-conglomerate are massive or poorly bedded. Crude bedding is indicated in calcareous pebble-conglomerate by differential weathering. At Fish River Lake, the quartzitic granule- and pebble-conglomerates on Blueberry Island are well bedded; large blocks of similar conglomerate cropping out at water level 700 feet west of Blueberry Island contain faint traces of lenticular cross bedding.

Calcareous rocks: Rocks consisting chiefly of carbonate minerals crop out very rarely within the present map area. The well known reef limestone of Square Lake, from which Helderberg fossils have been described (Billings, 1862) crops out northeast along the strike of the formation, in the south central part of the Square Lake quadrangle. Within the present

map area, biohermal limestone crops out at water level 1500 feet north of Blueberry Island in Fish River Lake. With the exception of tuff containing abundant comminuted shell fragments that crops out near the northwest shore of Island Pond (Winterville quadrangle), no other predominantly calcareous rocks were discovered within the established boundaries of the formation. Quartzitic siltstone with calcite cement occurs locally at the south end of Fish River Lake, also in the vicinity of Sheldon Ridge, and along the northwest shore of St. Froid Lake. Gray dolomitic sandstone crops out in a railroad cut north of Fish River Thoroughfare, on the south bank of the thoroughfare between Route 11 and the railroad bridge, and along the north shore of St. Froid Lake. Dolomitic rocks were not observed elsewhere.

Volcanic rocks: Several types of volcanic rocks are interbedded with the sedimentary sequence. They constitute an orogenic association of olivine-poor basalt, hypersthene andesite, quartz latite, spherulitic dacite, and trachyte. Flows were not observed superposed one upon another, as they so commonly are in the volcanics of the Winterville area. Nevertheless, superposed flows may be present but not adequately exposed to be recognized. Volcanic agglomerate is rare.

Petrographic study of the basaltic and andesitic lavas revealed that many of the dark aphanites of basaltic appearance are, in fact, andesites. The rocks are considerably altered, and bent twin lamellae in plagioclase and broken or distorted crystals may indicate that they were deformed during the period of folding. Mafic minerals are extensively altered. Serpentine-magnetic clots in andesite near the northwest shore of Fish River Lake are pseudomorphous after equant crystals of pyroxene. Andesite containing diopsidic augite crops out in the bed of the North Branch of Fox Brook near its confluence with the South Branch. Hypersthene-biotite andesite is well exposed along the north shore of Eagle Lake, near Oak Point. Despite the variability of mafic mineral content, the plagioclase of the andesites is rather uniformly sodic, ranging from An₃₂ to An₄₂. Amygdaloidal structure is a prevailing characteristic of these and other volcanics in the formation. Chloritized and hematite-rich rock fragments occur abundantly as inclusions. Andesite and the felsitic volcanics are commonly spherulitic; the compositions of the spherulites were not determined.

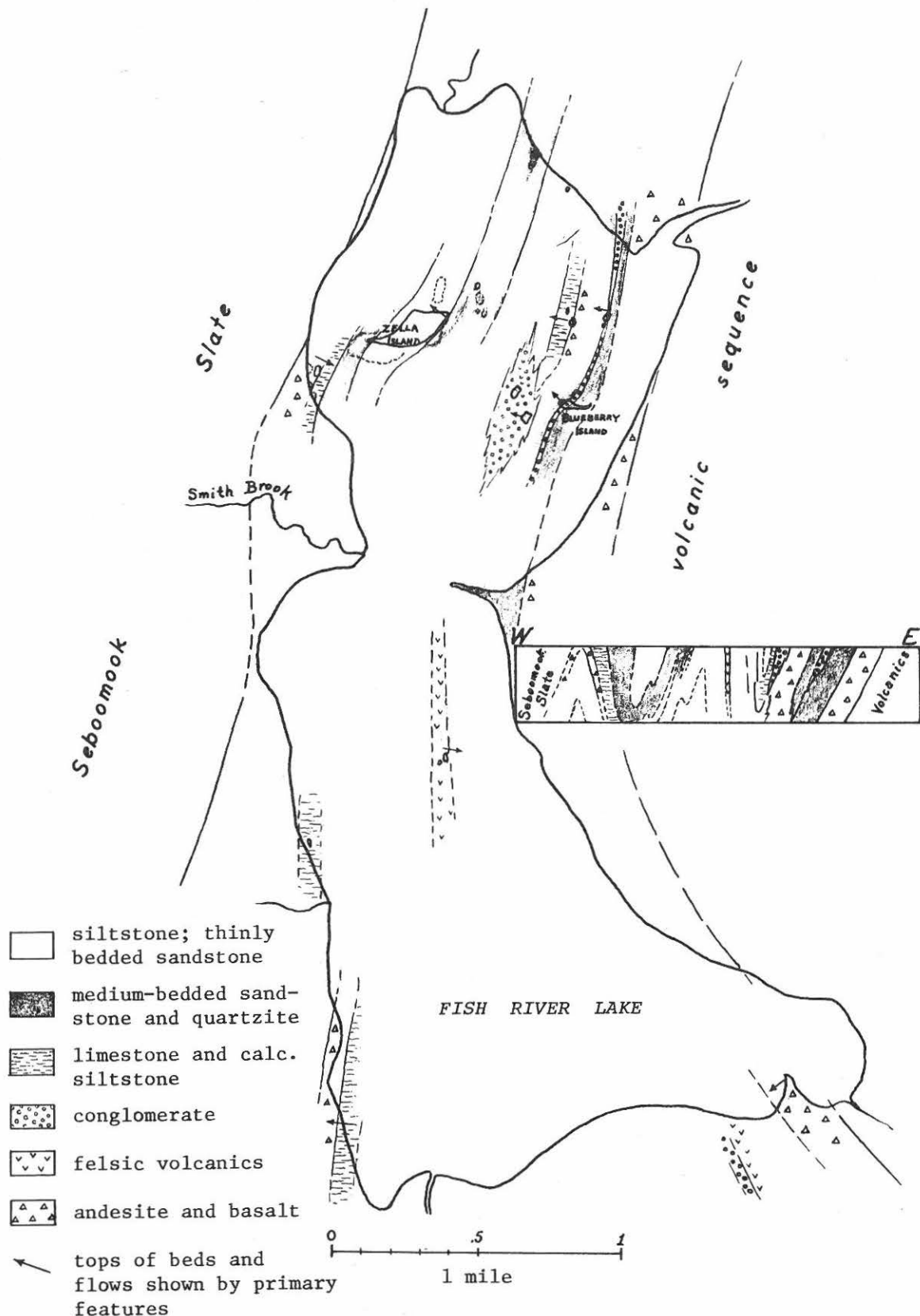


Fig. 3. Outcrop map of Fish River Lake composite structure section (inset at right) shows interpretation of major folds in the Fish River Lake Formation across the north central part of the lake.

Amygdaloidal, porphyritic quartz latite exposed on the hill-side immediately south of Fish River Lake contains andesine phenocrysts many of which are wholly or incompletely mantled by sanidine. Sparse phenocrysts of biotite are partly chloritized. Sanidine microlites are abundant in the groundmass.

Dense white or yellowish-gray rocks that are similar in appearance to massive chert are, in fact, aphanitic, felsic (dacitic?) volcanics, in part spherulitic. These rocks crop out 0.4 mile south of Luciffee Pond, 1.2 miles north of the outlet of Chase Brook at Fish River Lake, and at low water level in the central part of Fish River Lake where they display distinct pillow structure. Abundant laths of oligoclase-andesine impart to the rocks a trachytic texture. Very small spherulites of undetermined composition and calcite-filled amygdules are common. Dark, extremely fine-grained patches may represent recrystallized glass.

An isolated exposure of volcanic agglomerate one mile southwest of Mud Pond, south of Fish River Lake, consists predominantly of andesite fragments in a matrix of similar composition.

Fauna

Rocks containing marine fossils are rare. As large a collection as time and logistics permitted was taken at the few localities where marine faunas were discovered. Dr. A. J. Boucot, then at Massachusetts Institute of Technology, kindly studied five faunal collections, and the substance of his report, transmitted in 1957, is given here together with a brief description of each locality.

West end of Island Pond, Winterville quadrangle: This locality is roughly midway between the upper and lower contacts of the formation. Richly fossiliferous, fine-grained, light green to white calcareous tuff crops out in a small exposure near the shore. It is rather massive in general, but thin fossiliferous layers occur 6 to 8 inches apart along which the attitude of N.17°W., 30° E. was recorded. It is assumed that the beds also face eastward. Roughly five feet of section is exposed containing the following fauna:

Leptaena, sp.
Chonetes, sp.
Dalmanella (sensu strictu) sp.
Douvillina, sp.

The assemblage indicates a post-Ordovician, pre-Devonian age. Although the assemblage cannot be placed with accuracy, a late Silurian age is suspected. Amygdaloidal basalt, green quartzite, and pebble-conglomerate crop out nearby on the ridge to the north, and on the small island near the north end of the pond. The conglomerate is in all respects similar to that occurring on Blueberry Island in Fish River Lake.

Fish River Lake: The most westerly of two small islands north of Blueberry Island at Fish River Lake consists of volcanics and sediments dipping steeply to the northwest and facing in the same direction. Including the water-level outcrops 150 feet northwest, slightly over 100 feet of section is exposed; a ropy basaltic lava flow is exposed at the base, followed by sedimentary breccia that encloses weathered, oxidized fragments of the basalt, dark green calcarenite containing crinoid fragments, interbedded coarse limestone and dark green quartzite, and at the top, interbedded shaly limestone and nodular coralline limestone. The section contains the following fauna:

Atrypa reticularis
Proteus, sp.
Zaphrentis, sp.
Leptostrophia, sp.
Chonetes jerseyensis

Fossiliferous dark green quartzite occurs interbedded with conglomerate on Blueberry Island, Fish River Lake, and contains abundant shells of *Chonetes* sp. and also the following:

Nucleospira, sp.
a *Stropheodontid*
Crispella vanuxemi
Leptaena, sp.

Both of the assemblages listed above are indicative of a late Silurian (Cobleskill) age.

Dark calcareous cleaved siltstone crops out at low water level on the west shore of Fish River Lake, near Blanchard's Depot Camp. The siltstone here, although cleaved and deformed, contains an abundant fauna, from which were identified the following:

Leptaena "rhomboidalis"
Kozlowskiella perlamellosa
Meristella, sp.
Schuchertella, sp.
Atrypa reticularis
Coelospira concava

Eospirifer macropleura
Sphaerirhynchia vellicata
Orthostrophia, sp.
Strophonella, sp.
Levenea, sp.
Leptocoelia aff. *flabellites*

According to Boucot, the fauna is definitely of New Scotland, Helderberg age; the presence of *Leptocoelia* is novel inasmuch as it has hitherto not been reported in strata older than Oriskany age.

Bedding is closely parallel to cleavage at this locality, and both dip steeply westward. It is assumed that the beds face westward as well, but no conclusive evidence could be found.

North of Sheldon Ridge: Dark blue calcareous siltstone (similar to that outcropping at the west shore of Fish River Lake) crops out on the hillside half a mile north-northeast of Mud Pond, south of Fish River Lake. Fossils obtained from this locality were identified as follows:

Kozłowskiella perlamellosa
Leptaena "rhomboidalis"
Strophonella cf. *funiculata*
Leptostrophia sp.
"Fardenia" pecten
Atrypa reticularis
Resserella sp.

The fauna is probably of Silurian age, but cannot be accurately placed within the Clinton to Cobleskill interval. The presence of phacopid and dalmanitid trilobites suggests, however, a lower Devonian, Helderberg, age (H. B. Whittington, 1957, personal communication).

Flora

Interbedded with these rocks, particularly from Winterville township south to the north half of Fish River Lake, are sandstones and siltstones that carry carbonaceous impressions and molds of imperfectly preserved plant stems. These occur most abundantly in sandstone at Red Ledge on Red River, but the most completely preserved specimens occur in medium-bedded sandstone along the northwest shore of St. Froid Lake. Most of the stems are simple, elongate fragments generally less than a quarter of an inch in width, but a few have been found that are branched. The better preserved stems are striated parallel to their lengths. The stems are tentatively referred to the lower Pale-

ozoic genera *Psilophyton* or *Hostimella*, being similar to *Psilophytales* fossils in the upper sandstone member of the Perry formation of Finger Lakes age, southeastern Maine (Smith and White, 1905), and to a stem of *Hostimella* figured by Dorf from the lower Devonian of Wyoming (1934, pl. 43).

It is one of the most interesting characteristics of the Fish River Lake Formation that a flora of supposed "continental" type is associated with marine faunas of upper Silurian and lower Devonian time. The rarity of fossilized land-dwelling plants in the Silurian system is well known, and we may suppose either that they became common in this region before becoming widespread in the Devonian of eastern North America, or that marine faunas indicative of later Silurian time, with which the flora is interbedded, lingered on into the lower part of the Devonian in this area.

Numerous alternations between shallow shelf and estuarine environments characterized the deposition of the formation throughout its mapped extent (fig. 2). Throughout the entire sequence, deposition was likely interrupted in different localities during many short intervals, at different times. There is no lithologic evidence to support recognition of a widespread disconformity or unconformity within the sequence of rocks here described. I therefore regard the Fish River Lake Formation chiefly as late Silurian, but with the upper siltstones and volcanics extending across the systemic boundary to the New Scotland time-stratigraphic horizon.

Correlation

On the basis of the occurrence of *Leptocoelia flabellites* in the Seboomook Slate near Fort Kent, Maine (Nylander, 1940), the upper boundary of the Fish River Lake Formation lies within the Helderberg interval. The time-span of deposition of the Fish River Lake Formation allows correlation with several more precisely dated units.

The Square Lake Limestone

The few scattered outcrops of massive reef limestone around the west and north shores of Square Lake contain a brachiopod- and coral-rich marine fauna that places the age of the limestone definitely within the lower part of the Helderberg series (Williams and Gregory, 1900; Cooper, 1942). The Square Lake lime-

stone occurs well within the projected strikes of the lower and upper contracts of the Fish River Lake Formation. In view of the New Scotland age of the fauna from siltstone on the west shore of Fish River Lake, the limestone at Square Lake is equivalent to the upper part of this formation.

Chapman Sandstone

The Chapman Sandstone, cropping out 35 miles across strike to the southeast, is of New Scotland age (Williams and Breger, 1916; Boucot and others, 1964), and is therefore equivalent to the uppermost part of the Fish River Lake Formation.

Dalhousie Formation and subjacent units

Interbedded shale, limestone, tuff and lava flows of the "Dalhousie Beds" described by Alcock (1935; 1947) as of lower Devonian age, are lithologically similar to parts of the Fish River Lake sequence, but are late Helderberg or younger. The formations of Ludlow age, lying below the Dalhousie Formation (see Naylor and Boucot, 1965, for a summary of these units), more strongly resemble the lower part of the Fish River Lake Formation.

To the southwest of the Dalhousie area, and roughly on strike with the structural trend northeast of the mapped area (fig. 2), are gray and buff siltstone and sandstone that are exposed in road cuts on provincial route 17 in New Brunswick. The age of this sequence is not known, but the beds strongly resemble the sections of the Fish River Lake Formation exposed along the northwest shore of St. Froid Lake, and along Fish River Thoroughfare between St. Froid and Eagle Lakes.

The Gaspé Sequence

The Silurian and Devonian systems are well represented in the Gaspé peninsula, and the formations bracketing the Silurian - Devonian boundary have been studied in considerable detail (summarized by McGerrigle, 1950; Boucot, Cumming, and Jaeger, 1967). Although more than 200 miles separate the Siluro-Devonian of northern Maine from that of the Gaspé, it is worthwhile to suggest a tentative correlation, inasmuch as the Fish River Lake Formation and its enclosing rocks are on strike with the Gaspé sequence, which contains faunas of more European aspect.

The lower to middle Devonian is represented in eastern Gaspé by variously interbedded shale,

limestone, and shaly limestone. These have been divided into three formations which are, from youngest to oldest:

Grande Greve Limestone Cap Bon Ami Formation St. Alban Formation

The Fortin Formation, younger than the Grande Greve, contains conglomerate as well as slate. The St. Alban is the oldest of the Devonian formations. Lying beneath the St. Alban are Silurian units composed of shale, limestone, conglomerate, and volcanics. The downward sequence of sedimentary rocks including shale and conglomerate followed by volcanics bears lithologic similarity to the Fish River Lake sequence.

Both the St. Alban and the Grande Greve contain faunas of Helderbergian aspect in which are found genera and species similar to those in the upper part of the Fish River Lake Formation. The Grande Greve also bears some similarity to the New York Oriskany, by the presence of *Hipparionyx* and *Rensselaeria*.

Genera and species common to the Fish River Lake, the St. Alban, and the Grande Greve Formations are the following (generic and specific names as used by Boone, 1958):

<i>Fish River Lake</i>	<i>St. Alban</i>
<i>Atrypa reticularis</i>	<i>Atrypa reticularis</i>
<i>Dalmanella</i> , sp.	<i>Dalmanella perelegans</i>
<i>Leptaena "rhomboidalis"</i>	<i>Leptaena "rhomboidalis"</i>
<i>Leptostrophia</i> sp.	<i>Leptostrophia magnifica</i>
<i>Nucleospira</i> , sp.	<i>Nucleospira ventricosa</i>
<i>Strophonella</i> , sp.	<i>Strophonella punctulifera</i>
<i>Meristella</i> , sp.	<i>Meristella</i> , sp.
<i>Orthostrophia</i> , sp.	<i>Orthostrophia canadensis</i>
<i>Schuchertella</i> , sp.	<i>Schuchertella</i> , sp.
<i>Eospirifer macropleura</i>	<i>Spirifer plicata</i>
<i>Bronteus canadensis</i> (in Square Lake limestone)	<i>Bronteus canadensis</i>
dalmanitid and phacopid trilobites	<i>Dalmanites griffoni</i>
	<i>Phacops logani</i>
	(from McGerrigle, 1950, p. 56-59.)
<i>Fish River Lake</i>	<i>Grande Greve</i>
<i>Chonetes</i> , sp.	<i>Chonetes canadensis</i>
<i>Dalmanella</i> , sp.	<i>Dalmanella lucia</i>
<i>Leptostrophia</i> , sp.	<i>Leptostrophia magnifica</i>
<i>Leptaena "rhomboidalis"</i>	<i>Leptaena "rhomboidalis"</i>
<i>Leptocoelia</i> aff. <i>flabellites</i>	<i>Leptocoelia flabellites</i>
<i>Coelospira concava</i>	<i>Coelospira concava</i>
<i>Strophonella</i> , sp.	<i>Strophonella continens</i>
<i>Eospirifer macropleura</i>	<i>Eospirifer murchisoni</i>
<i>Schuchertella</i> , sp.	<i>Schuchertella becraftensis</i>
dalmanitid and phacopid trilobites	dalmanitid and phacopid trilobites
	(from McGerrigle, 1950, p. 69-72).

Because of environmental control, all three faunal assemblages compared above may be facies faunas. Other species might now distinguish more specifically the paleontologic age of these and the younger faunas of the Fish River Lake Formation.

STRUCTURE—Upper Contact

One of the principal objectives of the field mapping was to determine the stratigraphic relationship of the Fish River Lake Formation to the Seboomook Slate. Detailed study along the contact revealed that bedding in the slate and in the adjacent sedimentary rocks dips steeply northwest and, with rare exceptions, faces northwest. The direction in which beds face was determined on the basis of graded bedding and cross-bedding; intersection of bedding and cleavage in slate near the contact gives the same indication, but was not used as a criterion.

Bedrock exposures along the north shore of the east arm of Eagle Lake (hereafter referred to simply as the north shore of Eagle Lake) expose the contact clearly for approximately a mile along the strike. The exposures provide a "type section" of the contact with which less well exposed parts of the contact in the Winterville and Fish River Lake quadrangles can be compared.

The following rock types are found to lie directly beneath the Slate at different places along the north shore: red and green shale, hypersthenebiotite andesite, gray-green shale with epidote nodules, and gray shale. Where mottled red and green shale directly underlies slate, the rock types listed above occur in stratigraphic succession from the contact southeastward. There appears to be no structural discordance between bedding in the sedimentary rocks and in the slate. If an unconformable or disconformable relationship existed prior to folding, it is now obliterated. It appears likely, however, that at least some erosion took place before the deposition of the shale sequence that is now the Seboomook Slate, inasmuch as a given rock-type, among those listed above, is not found directly underlying the slate along the entire extent of the contact in the map area. Therefore I believe that disconformable relations existed, but only in part, between the Seboomook and the Fish Lake Formations. Along the north shore of Eagle Lake the two units are clearly in sedimentary contact. In the

southwest part of the Fish River Lake quadrangle, however, thrust faulting has locally modified the contact relationship.

Elsewhere along the contact, red slate (or equivalent red shale) occurs near the mouth of the unnamed brook south of Labbe Brook, along the two brooks between Fish River Lake and the North Branch of Fox Brook, and on the lower part of the North Branch of Fox Brook. Altered andesite flows occur at the latter three localities, and also at Lucifree Pond, along Island Pond stream, at Red River Falls, and on McNally Brook.

The section of rocks identifying the contact on the north shore of Eagle Lake averages 20 feet in thickness. If larger exposures were present elsewhere along the contact a larger part of the "Eagle Lake contact section" might be found at localities southeast of Eagle Lake.

Lower Contact

Reddish-brown conglomerate with a tuffaceous matrix and with a maximum exposed thickness of ten feet rests upon basaltic and andesitic volcanics on the east shore of St. Froid lake, a mile south of Winterville Station. This contact has been selected as the lower contact of the Fish River Lake formation in this area, as sedimentary rocks are not found interbedded with volcanics immediately to the south, nor are thick sequences of volcanics found interbedded with sedimentary rocks to the northwest. The structure of the volcanics exposed beneath the conglomerate is not clearly evident.

A basal conglomerate may not persist, however, along the strike of the lower contact. At Fish River Lake, volcanics ranging from basalt to dacite are interbedded with sedimentary rocks typical of the formation; however, sandstone, shale, and limestone also occur, but seemingly more sparsely, in the volcanic sequence of the Winterville area. Basal conglomerate similar to that at St. Froid Lake was not found elsewhere. Thus the relationship between the Fish River Lake Formation and the predominantly volcanic sequence to the east may be (a) unconformable at certain localities, and (b) one of gradation by interbedding at others. Interbedded basalt and green quartzite at the northeast end of Fish River Lake may be an expression, although on a small scale, of gradational relations in the contact zone.

A stratigraphic sequence typical of the Fish River Lake Formation is at present not known

to be infolded with the volcanic sequence of the Winterville area, nor with the Seboomook Slate. With a possible exception in the southeast corner of the Fish River Lake quadrangle, the upper and lower contacts have not been repeated by folding. On the basis of the contact relations reviewed above, and on the prevalence of northwest dips within the Fish River Lake Formation, there seems little doubt that the regional structure along the strike of the formation is essentially homoclinal.

Folds and Faults

Lack of outcrops and uncertainty as to correlation of small sections along strike prevent any detailed interpretation of the folds within the formation. Outcrops across the strike are more abundant in and around the north shore of Fish River Lake than elsewhere in the map area, and the least amount of projection along strike is required here to obtain a tentative interpretation of the structure and also thickness of the formation (Fig. 3).

All rocks outcropping from the outlet of the lake to Zella island, including Blueberry island and the smaller islands nearby, dip steeply westward and face in the same direction. The direction in which sandstone, shale, and conglomerate face on the north shore of the lake is indeterminate. Outcrops of andesite west of Zella Island, indicate by the increase in size of amygdules eastward, that the section there faces east. Similarly, spherulitic dacite with well-defined pillow structure at shoaling outcrops in the central part of the lake, faces eastward. Basalt, shale, and calcareous siltstone outcropping around the south shore of the lake, dip steeply westward and face in the same direction.

Therefore the rock sequence at Fish River Lake is repeated across two anticlines, the axial trace of one striking approximately north through the middle of the lake, and that of the other lying west of the amygdaloidal andesite west of Zella Island. The inferred relations are shown in the cross section of Figure 3. Taking into account the "major" folds and frequency of minor folds, the formation is estimated to be approximately 3000 feet in maximum thickness at the north end of Fish River Lake.

In view of the dominantly homoclinal structure of the formation in the present map area, a tentative correlation of local sections along the strike of the formation is shown in Figure 4.

A thrust fault of unknown displacement occurs near the east margin of the slate at Red River Falls, and at least one thrust fault is exposed at three localities southwest of Fish River Lake. Structural relations south of Fish River Lake are imperfectly established. The brecciation of red shale and altered andesitic volcanics in the vicinity of the faults suggests that the Seboomook has overridden the Fish River Lake Formation at the contact. However, typically undisturbed contact relations occur on the lower part of the North Branch of Fox Brook. The contact was not established farther south. East of the contact zone described above, red shale and dark volcanic rock have been found at three localities on strike with each other, again typical of contact relations noted elsewhere. It is possible that the folded contact surface may be repeated by thrust-faulting from the North Branch of Fox Brook northeast toward Fish River Lake, but lack of outcrops prevented tracing the contact in detail in the critical area immediately south of Smith Brook.

The close proximity of slate to basaltic volcanics on Sheldon Ridge indicates that the Seboomook - Fish River Lake contact has been offset by a transverse fault located north of Sheldon Ridge. The Fish River Lake Formation is also inferred to be offset by a transverse fault southwest of Lucifée Pond, in view of the offset relationship of slate and dacitic volcanics.

Environments of Sedimentation

Features considered in determining the environments of sedimentation are: the predominance of both coarse - and fine-grained arenites; the intercalation of beds carrying, on the one hand, marine faunas, and on the other, a continental flora; the angularity of the clastic grains; the closely spaced alternation of thinly laminated siltstone and sandstone with deltaic cross-bedding which is common throughout the area; and the presence of pillow structure in lava flows intercalated with the sedimentary sequence.

The distribution of lithologic assemblages, sedimentary and volcanic structures, and fossils lead to the conclusion that sedimentation took place under generally shoaling conditions northeastward, ranging from near-shore marine, to estuarine, and possibly fluvial. The formation is characterized by abrupt facies changes along and across strike, reflecting deposition under rapidly changing environmental conditions.

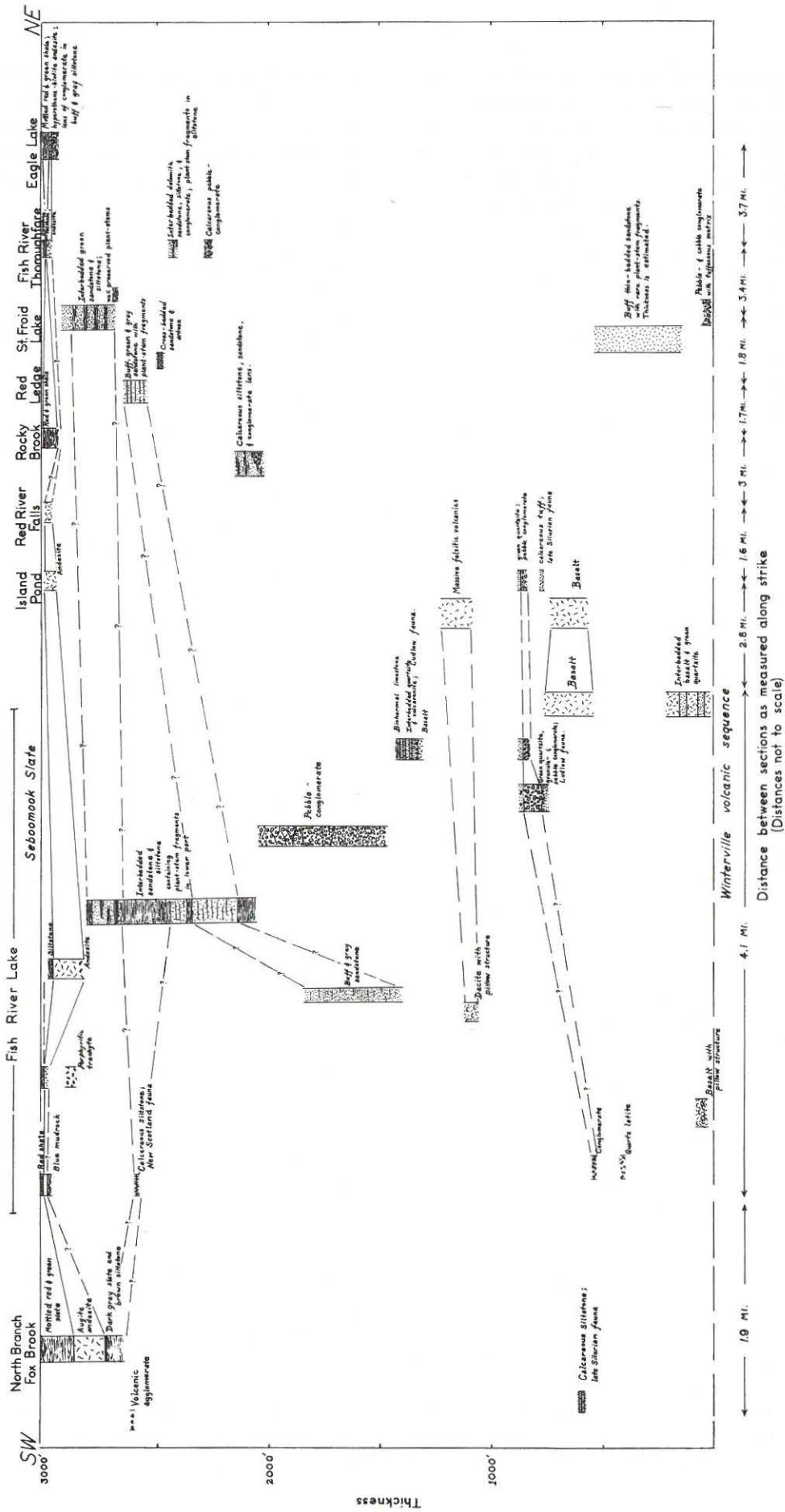


Fig. 4. Stratigraphic position and correlation of local sections in the Fish River Lake Formation.

Within the stratigraphic framework of reference, limited as it is by the lack of extensive sections across the formation (Fig. 4), a general contrast in environments appears to exist between the northeastern part of the formation and the southwestern part, as exposed along the strike for 24 miles within the map area. Most of the rocks in the northeastern part consist of coarse-grained arenites displaying deltaic cross-bedding with abundantly interbedded siltstone; volcanics are rare, and do not show pillow structure. Rocks containing a marine fauna were not observed, although sandstone beds containing plant stem fragments are common in nearly all outcrops. The only locality exposing calcareous rock is Limestone Point at Square Lake. In the southwestern part, particularly in the vicinity of Fish River Lake, quartzite, calcareous siltstone, and biohermal limestone containing marine faunas ranging in age from Late Silurian to Early Devonian are interbedded with conglomerate, sandstone containing plant stem fragments, and massive basaltic volcanics. Basalt and dacite with pillow structure also occur.

These relationships indicate that fluctuations of uplift and subsidence (and hence marked shifting of marine, near-shore, and perhaps subaerial environments) took place frequently during deposition. Nevertheless, throughout the period of alternating transgression and regression, it appears likely that marine conditions prevailed across that part of the formation that now crops out in the vicinity of Fish River Lake. The conditions cannot be specified in detail, but the sea floor may have occupied a near-shore part of a shallow shelf. The colonial corals of the biohermal limestone and the shells of brachiopods show virtually no abrasion, and hence were not transported far; these fauna are preserved in their natural environment. The littoral zone seems not to be represented in rocks containing a marine fauna. Intermittent volcanism erupted lavas onto the shallow marine shelf, with the formation of pillow structure in the vesicular flows.

The conglomerates in the vicinity of Blueberry Island occupy large channels in the underlying quartzite, and show festoon cross-bedding. The conglomerates thus record sequences beginning with uplift, accompanied by scouring of the underlying sands, followed by rapid subsidence and accumulation of new layers of finer-grained sediments that buried and

preserved the conglomeratic zones. The common alteration of plant-fragment sandstone with siltstone may indicate the same sequence that operated, in this case, in a much milder degree in an estuarine or deltaic environment.

The Bordering Land

From provenance information in the rocks of the Fish River Lake Formation, we can construct a paleogeographic framework in regard to the kind of rock material constituting the adjacent land during the period of deposition, and its position relative to the accumulated sediments. Most of the coarser arenites are feldspathic in varying degrees, and some are distinctly arkosic. The granules and pebbles in conglomerates are composed, in order of decreasing abundance, of basaltic and felsic volcanics, quartz, pink potassic feldspar, jasper, and granite. The volcanics of the Winterville area, lying to the east and stratigraphically beneath the Fish River Lake Formation, consist predominantly of mafic and felsic volcanic rocks. There seems little reason to doubt that they constitute, at least in part, a volcanic island or archipelago that experienced intermittent uplift during late Silurian and early Devonian time and furnished some if not all of the detrital sediments and possibly some of the lava flows now found in the Fish River Lake Formation. If this is so, then evidence should exist that material was brought into the sedimentary environment from the east.

Regional Slope

Three criteria were noted during the mapping of the Fish River Lake Formation that indicate current direction, and allow reconstruction of prevailing surface slope. Firstly, sandstone and arkose outcrops in the bed of Red River and in Fish River thoroughfare show deltaic cross-bedding, which, if the section were restored to a horizontal position, would indicate the flow of currents from northeast to southwest. Secondly, plant stem fragments in sandstone along Fish River Thoroughfare would be aligned northeast-southwest after restoration of the section to the horizontal. There may be some equivocation here as to whether currents flowed from the northeast or from the southwest. Thirdly, near the base of an amygdaloidal andesite flow that crops out west of Zella Island, Fish River Lake, some of the amygdules have an inverted tear-drop shape (facing toward the top of the

flow) and these "tear-drop" amygdules are now overturned toward the southwest.² This may indicate a local, if not regional, southwesterly slope at the time of extrusion of the lava. (From the preservation of volcanic textures, the shape of the amygdules appears not to be a result of deformation during folding of the rock sequence.)

It is then, likely that the volcanic sequence of the Winterville area as now exposed may be a southerly continuation of an island or chain of islands that bordered the late Silurian-early Devonian sea. The occurrence of marine strata in the southwestern part of the outcrop belt of the Fish River Lake Formation, and the deltaic aspects of the interbedded siltstones and sandstones to the northeast wherein the plant-stem fragments are larger and less macerated or decomposed, may be taken as corroborating evidence of a depositional surface that shoaled toward the northeast.

The Fish River Lake-Seboomook sequence records a net eastward or northeastward transgression of marine waters from upper Silurian to lower Devonian time. The widespread appearance of red, green, and mottled red and green shale at the upper contact of the Fish River Lake Formation may record the presence of extensive mud flats, frequently subaerially exposed, where red ferric oxide was stable, and possibly produced. The interval here recorded of widespread deposition of mud brought to a close the near-shore sedimentation characteristics of the Fish River Lake Formation, and heralded the onset of the thick accumulation of Seboomook argillaceous sediments.

²If the tear-drop shapes were derived by the upward flow of gas bubbles in viscous lava, then their long axes would be vertical if the bubbles were rising in motionless lava. It seems likely that the present shape of the "tear-drop" amygdules was produced by homogeneous shear in lava flowing toward the southwest.

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Devonian Slate Problems in the Northern Appalachians

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INTRODUCTION

The most widespread sedimentary unit in the northern Appalachians is a gray banded Devonian slate. The banding is formed by the alternation of dark-colored siltstone or mudstone and light-colored, fine-grained sandstone. This rock unit has been assigned a plethora of formational names over the great area of its distribution from central Gaspé, northern New Brunswick, eastern Quebec, northern Maine, and the Connecticut Valley region of western New England. Enough information is currently available about its physical characteristics, distribution relative to other Devonian rock units, and paleontologic contents to let us know that it provides a number of serious, currently unsoived stratigraphic questions. These problems are as follows:

1. In the Eastern Townships of Quebec, from Lake Memphremagog northeast to Lake Temiscouata, immediately below the Devonian slate are limestones which have yielded Onondaga, i.e., Eifel age Middle Devonian fossils. Elsewhere in the region from Gaspé southwest to northern Maine fossils found in the body of the unit are almost wholly of Becraft-Oriskany, i.e., Siegen age, with the exception of a few possible New Scotland, i.e., Late Gedinnian occurrences. The nature of the transition or contact, both stratigraphically and geographically, from the middle portion of the Early Devonian to the lower part of the Middle Devonian is unknown.
2. Recent studies in the Slate Belt of central Maine from Southern Aroostook and Penobscot Counties southwest almost to the New Hampshire line have shown that fossiliferous, fine-grained terrigenous rocks (not, however, to any significant degree of typical Seboomook aspect) span at least the Middle Ordovician through Early Devonian interval. Along strike in New Hampshire east of the Bronson Hill Anticline, similar rocks, uniformly metamorphosed and unfossiliferous to date, have been assigned to the Early Devonian.
3. The southeastern and eastern limits of known Devonian slate in central Maine and New Brunswick occur in synclinal structures far removed from the almost totally non-marine and volcanic Early Devonian (which contains

Early Gedinnian fossils in its basal part, the remainder not yet having yielded diagnostic fossils) of coastal New Brunswick, Maine, and Massachusetts. The Devonian slate is a flysch-like unit considered by most workers to indicate deposition under relatively deep water conditions. The geologically poorly known terrane between the Devonian slate and the coastal non-marine sequence may well include unmapped synclinal structures containing transitional facies. The gap between the two known Devonian belts ranges from about fifty to one hundred miles across strike with wide possibilities left open for investigation in both southern Maine and New Brunswick.

4. In the Lake Temiscouata region, Quebec (Lajoie, *et al* 1968, p. 635-638) Middle Devonian fossils occur in a thin limestone underlying the Devonian slate in several areas, as well as in limestone lentils within the slate; however, part of the slate may also be of Late Silurian age in this area if the structural relations have been correctly interpreted. Solution of this problem on either a structural or a stratigraphic basis is critical to furtherance of our understanding of the Devonian slate.
5. In the southeastern portion of the Eastern Townships fossiliferous calcareous rocks of Late Silurian age may grade laterally and vertically up into similar beds of possible Siluro-Devonian age, which in turn grade up into the Devonian slate, the slate in turn laterally to the Northeast in the Chaudiere River region being underlain in several places by Onondaga, i.e. Eifel age Middle Devonian limestone. It is important that the possible facies or unconformable relations be reinvestigated in the field in order to provide a more satisfactory solution.
6. Paleontologically the following opportunities exist for shedding further light upon the Devonian slate problem: examination for conodonts of the remaining unstudied carbonate lentils (particularly within the Eastern Townships) within the Devonian slate and conformably beneath it; an entire new program of studying hydrofluoric acid residues from the Devonian slate for chitinozoans and hystrichospheres.

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DISTRIBUTION

It is no accident that the Devonian slate of the northern Appalachians has been referred to colloquially as "The Great Gray Slate." Its occurrence in numerous synclinal structures extending from east-central Gaspé southwest and south virtually to Long Island Sound is linearly impressive as is its distribution across strike, reaching one-hundred miles in northern Maine. Geologists unfamiliar with northern Appalachian geology and terminology would have difficulty realizing the extent and areal importance of the unit because of the welter of formational names which have been applied to it in various areas over more than one-hundred years of study. It is appropriate, therefore, to review the distribution of the unit geographically, indicating for each area what formational terms have been employed in the past. This review will be done beginning in the northeast with Gaspé and ending up with the southern parts of the Connecticut River Valley region.

Quebec: Gaspé

McGerrigle (1946, p. 47-49) introduced the term Fortin Series for gray Devonian slates extending from the Matapedia Valley of western Gaspé in a wide swath eastward through the Peninsula to within ten miles of Percé. A subsequent map on a scale of 1:253,440 (McGerrigle, 1953) details the distribution of the Fortin as known at that time. Subsequent studies have not materially changed the picture.

The Fortin Series is somewhat unique in that its northwestern and eastern extremities become very calcareous, intertonguing ultimately with limestones of Early Devonian age (Devonian portions of the Gaspé Limestones). It is tempting to compare the calcareous Fortin with the Waits River Limestone, and the non-calcareous Fortin to the Gile Mountain Formation.

Quebec: Eastern Townships

The southeastern portion of the Eastern Townships adjacent to the International Boundary from northern Vermont and New Hampshire northeast to the Provincial Boundary with New Brunswick is underlain in large part by Devonian slate. To the northeast the term Temiscouata Group (McGerrigle, 1933, p. 116) was introduced for the Devonian slate revised recently by Lajoie, Lesperance, and Beland (1968, p. 635-638) after remapping and study of the region to the Temiscouata Formation. In general the term Temiscouata has

been employed by Quebec geologists working northeast from the Temiscouata region into the synclinal region to the south of Rimouski, changing to the equivalent term Fortin in the Matapedia Valley region.

To the southwest of the Temiscouata region the term St. Juste Group (Bland 1952, 1957) has been introduced for the Devonian slates exposed from the area of the Chaudiere River northeast to the International Boundary northeast of Lac Frontiere.

Southwest from the Chaudiere River to the Vermont-New Hampshire portion of the International Boundary the nomenclature applied during the last forty years to the Devonian slate is tangled. McGerrigle (1934) introduced the term Compton Formation in the Mount Megantic area, considering it to be of Ordovician age. Cooke (1950 and earlier) working southwest of the Mount Megantic area employed the term "quartzite-slate member" of the St. Francis Group, which he considered to be of Ordovician age. Finally Marleau (1958), mapping northeast of the Mount Megantic area, employed the terms Compton Formation and Seboomook Formation for two belts of Devonian slate, assigned both of them a Middle or Lower Devonian age, and included them both in the St. Juste Group. Clark (1934) employed the term upper Tomifobia Group for Devonian slate adjacent to the Vermont-New Hampshire line, assigning it to the Ordovician.

Northern New Brunswick:

Devonian slate is widespread in the western halves of Restigouche and Madawaska Counties, New Brunswick. A formal stratigraphic nomenclature has never been applied to this little-mapped region, but the distribution of the beds is shown by Potter, Jackson, and Davies (1968) as the D₁ unit in this area. The northern Maine term "Seboomook" or the Quebec term "Temiscouata" could be applied to the New Brunswick occurrences.

Northern Maine:

The widespread Devonian slate of northern Maine is assigned to the Seboomook Formation. The distribution of this unit on a 1:500,000 scale is well shown on the Preliminary Geologic Map of Maine (Doyle, 1967).

New Hampshire:

The Devonian slate of New Hampshire in the northern part of the state is assigned to

the Gile Mountain, Littleton, and Meeting House Slate. Billings (1955) considered both the Meeting House Slate and the Gile Mountain Formation to be of Ordovician age, but he assigned a Devonian age to the Littleton Formation. That portion of the Littleton Formation occurring around the domes and nappes of the Bronson Hill Anticline in the western portion of the State, and extending west to the Connecticut River, is uniformly underlain by strata of well demonstrated Silurian age. However, east of the Bronson Hill Anticline in south-central New Hampshire there is a more extensive region assigned by Billings (1955) to the Littleton Formation. Along strike to the northeast in Maine similar rocks have yielded Ordovician, Silurian, and Early Devonian fossils, which raises the possibility that the "Littleton" of south-central New Hampshire may also include pre-Devonian age beds.

Work along strike in northern Maine has shown a roughly similar Siluro-Devonian stratigraphy and history to that in New Hampshire. Within the Bronson Hill Anticline region the presence of a pronounced Taconic unconformity in both Maine and New Hampshire has helped sharpen stratigraphic contrasts within the Early Paleozoic, whereas to the southeast of the homologous region in Maine a relatively monotonous slate belt stratigraphy is present with no evidence for any Taconic unconformity (Pavlides, Boucot, and Skidmore, 1968). The possibility exists that a similar situation will be found in New Hampshire east of the Bronson Hill Anticline.

Vermont:

The distribution of the Devonian slate in Vermont is well shown by Doll, *et al* (1961), where it is assigned to the Gile Mountain and Littleton Formations.

Massachusetts:

The Devonian slate in Massachusetts west of the Connecticut River Valley Triassic is assigned to the Leyden Argillite, a strict senior synonym of the Gile Mountain Formation (Emerson, 1917, p. 47-48). East of the Connecticut River Valley Triassic the western New Hampshire term Littleton can be employed. To the east of the Bronson Hill Anticline's southern extension into Massachusetts there is a wide expanse of Paleozoic strata which may be in part Devonian.

Connecticut:

It is possible that the calcareous Wapawaug Schist, a Waits River lithology, west of New Haven is the southwesterly equivalent of the Devonian slate (Rodgers, *et al*, 1959). East of the Triassic in Connecticut Rosenfeld (in Rodgers, *et al*, 1959) concludes that the upper member of the Bolton Group is the southerly extension of the Littleton Formation, i.e. equivalent to the Devonian slate.

South of Long Island Sound:

Evidence is not available concerning the southerly limit of the Northern Appalachian Devonian slate, but there is no reason to suspect that the presently known limits approximate the real distribution. Knowledge of Piedmont geology from Virginia south is still so rudimentary as not to preclude extensions even into such a distant region.

The existence of a linear belt of Cambrian and, locally, Early Ordovician age banded argillites along and adjacent to the coast of greater Acadia in southern New Brunswick, adjacent Maine and the Boston area has been well known for over one-hundred years. St. Jean's (1964) recent discovery of paradoxidean trilobites, i.e., Middle or possibly Early Cambrian, in the North Carolina sector of the Carolina Slate Belt on the southeast side of the Piedmont in similar banded argillites associated with minor intercalations of tuffaceous and possibly extrusive volcanic rocks suggests the strong possibility that this Cambrian belt can be extended far to the southwest from the Boston area. It may be more than a coincidence that the distance across strike from the fossiliferous Cambro-Early Ordovician of the Valley and Ridge to the Cambrian of the Carolina, slate belt in North Carolina, about two hundred miles, is close to the distance across strike from the similar fossiliferous Cambro-Early Ordovician lithofacies on the west side of the Berkshires and Green Mountains to the Cambrian argillites of the Boston area (the measurement across strike from the St. Lawrence Valley region Logan's Line area to southeastern New Brunswick is closer to three hundred miles). This situation for the Cambrian holds out the hope that southwesterly extensions of the Devonian Slate will be found in the Piedmont from Virginia south (the gross similarity of the Piedmont Paleozoic rocks north of Virginia to those on the west side of the Green Mountains

makes it much less likely that Devonian slate will be found in the Piedmont north of Virginia).

AGE

Unfortunately one of the characteristics of the cyclicly banded, flysch-like northern Appalachian Devonian slate is the almost total absence of megafossils (micro-fossils other than conodonts extracted from the rare carbonate rocks associated with the Devonian slate have not hitherto been sought). The bulk of the known fossil localities with direct bearing upon the age of the Devonian slate are situated as follows: carbonate strata immediately beneath the Devonian slate, carbonate lentils within the Devonian slate, intertonguing sandstone and slate immediately adjacent (no more than one mile along strike) to major facies changes from the Devonian slate into shallow water, fossil-rich sandstone. Within the main body of the slate the typical, although extremely rare, fossil occurrences consists of solitary or very scattered shells.

The intimate involvement of the Devonian slate in Middle Devonian, Acadian age folding and metamorphism makes difficult the placement of a precise upper limit on its age for the following reasons: In most areas it constitutes the youngest pre-Acadian age stratigraphic unit. In those areas where post-Acadian age units are present they consist almost entirely of poorly fossiliferous, non-marine Devonian or Carboniferous units which themselves are difficult to date. Fossils useable for dating the Devonian slate have been found almost entirely in Gaspé, the Eastern Townships of Quebec, and northern Maine. Northern New Brunswick has provided a few occurrences, and northern New Hampshire a few as well. The bulk of the Littleton and Gile Mountain Formations together with their equivalents to the south in the Connecticut River Valley region have not yielded any fossil material useful for dating purposes. Almost none of the fossils from the Devonian slate or units helpful in dating it have been illustrated or described; reliance being placed on a few faunal lists and a mass of unpublished information accumulated by myself in connection with the routine examination of fossils collected by myself and others in the course of geologic mapping. Some of this data will be made available in a description of Maine brachiopods (Boucot, in preparation).

Quebec: Gaspé

In eastern Gaspé, adjacent to the facies change with the upper, Early Devonian portion of the Gaspé Limestones, fossils of probable Becraft-Oriskany, i.e. Siegen age have been obtained. In western Gaspé in the Matapedia Valley region a few faunules of Becraft-Oriskany age have been obtained. Structural and stratigraphic relationships of the Fortin Formation to both older and younger units in Gaspé are still poorly understood, despite the existence in the Peninsula of both older and younger Devonian rock units containing fossils, as well as abundant Silurian and older fossiliferous rock units.

Quebec: Eastern Townships

Brachiopods of Late Lower or Early Middle Devonian age have been identified from a limestone lentil in the Temiscouata region. In the same region, the Touladi Limestone, which underlies the Temiscouata Formation, contains conodonts of Eifel, i.e., Onondaga, Early Middle Devonian age, as well as corals and brachiopods consistent with this determination.

The Temiscouata Formation south of Rimouski, including that of northernmost New Brunswick, can be interpreted to be partially or wholly a facies of the known calcareous and terrigenous later Early Devonian (Grande Greve Limestone and overlying Gaspé Sandstone which together contain fossils and span the interval Becraft-Oriskany, Esopus, and Schoharie), but mapping in this region is too preliminary for us yet to decide whether the Devonian slate should be equated with this entire sequence or with only a portion of it. In view of the Middle Devonian fossils occurring in the Temiscouata regions nearby and their absence in the Matapedia Valley region it is clear that caution is appropriate.

In the Riviere Bleue area, southwest of the Temiscouata region, Lajoie, Lesperance, and Beland (1968, p. 635-638) mention the occurrence of Late Silurian corals (identified by Oliver) from limestones interpreted to lie below the Temiscouata slate with a transitional contact.

Further to the southwest in the Chaudiere River region the Devonian slate (St. Juste Group) has not yielded diagnostic fossils. Conformably below it, however, lies the Famine Limestone which has yielded a conodont fauna of Eifel age, and a coral fauna of Onondaga

age. Both faunas indicate an Early Middle Devonian date, and the evidence afforded by associated brachiopods is consistent with this view.

Southwest of the Chaudiere region the Devonian slate (Compton Formation, upper St. Francis Group) has not yielded diagnostic fossils. The thick calcareous unit below, i.e. the lower St. Francis Group, is in turn underlain by calcareous rocks in several localities which have yielded Late Silurian (both Ludlow and Pridoli) age brachiopods (see Boucot and Drapeau, 1969). The isolated Mountain House Wharf Limestone of the Lake Memphremagog area has yielded a coral fauna of Onondaga, i.e. Early Middle Devonian, age similar to that found in the Famine. This evidence suggests that an upper portion of the lower St. Francis Group may be of Early Middle Devonian age whereas the lower portion of the St. Francis Group is probably of Late Silurian age, grading up into beds of Early Devonian age. Boucot and Drapeau (1969) interpret the lower, calcareous member of the St. Francis Group as at least in part a southwestern facies of the Devonian slate occurring to the northeast of the lower member's northeastern termination.

Northern New Brunswick:

The Devonian slate of northern New Brunswick has yielded only a few scattered brachiopods of diagnostic value (*Leptocoelia flabellites*). Elsewhere within the body of the Devonian slate *L. flabellites* has only been found in well dated faunas of Becraft-Oriskany, i.e. Siegen age, although the absolute range of this taxon is New Scotland-Schoharie, i.e. Late Gedinian-Emsian. Isolated occurrences of *L. flabellites* must, therefore, be used with care and keeping in mind that in the northern Appalachians the taxon has been found in strata of Esopus, i.e. Late Siegen and/or Early Ems age, but never in beds of Schoharie age.

New Hampshire:

Fossils do indeed occur in the Littleton Formation, but it is critical to understand that diagnostic ones occur only in two, small, isolated fault blocks in northern New Hampshire (Boucot and Arndt, 1960) in lithologies somewhat different than the typically banded Devonian slate of the northern Appalachians. One occurrence is at the top of a sequence grading up from the Late Silurian in the Littleton Quadrangle, and the other at the base of the sequence

in the Whitefield Quadrangle, both yielding faunas of Schoharie age. Rigorously speaking neither of these New Hampshire occurrences can be traced into the main mass of the Littleton Formation. There is, however, no reason to conclude that these two fossil localities give us an upper and lower limit for the age of the entire Littleton Formation.

The Whitefield Quadrangle occurrence can be interpreted as a locality on the fringes of a Schoharie age island, and the Littleton Quadrangle occurrence, offshore from the same island. The presence of Silurian strata beneath the Littleton Quadrangle Devonian fossils and the absence of Silurian strata beneath the Whitefield Quadrangle locality supports this interpretation.

Northern Maine:

The majority of the fossil localities in the Devonian slate (Seboomook Formation) of northern Maine occur around and in the Moose River and Roach River synclinoria of Somerset, Franklin and Piscataquis Counties. Scattered occurrences are known from elsewhere in Piscataquis, Penobscot, and Aroostook Counties, but few have yielded much information of stratigraphic value.

Boucot and Heath (1969) have summarized the information about the Seboomook Formation and its fossil localities adjacent and within the Moose River and Roach River synclinoria. The bulk of the Seboomook Formation localities are within a mile or less of the facies change into the Becraft-Oriskany age Tarratine Formation, and contain fossils either consistent with a Becraft-Oriskany age or strongly supporting it. At the southwest end of the Moose River synclinorium, in the Beck Pond area, the basal Seboomook contains a few brachiopods of Late Helderberg, possibly New Scotland age. Slightly higher in the Seboomook, the Bear Pond Limestone Member has yielded a New Scotland type Helderberg fauna. Our problem with these Helderberg faunas is that not enough data is available about the taxa included within them from an evolutionary point of view to permit us to completely discount the possibility that while the fauna is certainly of Helderberg-New Scotland aspect it might not represent a distinct community of Becraft age. It is possible that additional data provided by conodonts might solve this dilemma, but for the moment the base of the Seboomook in this

area must be considered to have a possible lower limit of New Scotland, i.e. Late Gedinnian age.

Overlying the Tarrantine and Seboomook Formations of the Moose River and Roach River synclinoria is the Tomhegan Formation of Schoharie age. All the current structural data and their interpretation are consistent with the Seboomook in this region being of pre-Tomhegan age, but along the northernmost and northwesternmost boundaries of the Tomhegan Formation it is possible to infer that an upper part of the Seboomook could be of Schoharie age. It is important to note that at many localities in this region it can be well demonstrated that significant portions of the Seboomook Formation are of pre-Tomhegan age.

The Seboomook Formation of Piscataquis County, Maine has yielded only a few fossil localities away from the Moose River and Roach River synclinoria. All of these localities contain faunules supporting or consistent with a Becraft-Oriskany age.

The Seboomook Formation of northern Penobscot County, north and east of the Katahdin batholith, has yielded very few fossil localities; all faunules are consistent with a Becraft-Oriskany age. Adjacent to the East Branch of the Penobscot River the Seboomook has lateral facies relations in part with and in part underlies the Matagamon Sandstone of Becraft-Oriskany age.

In northern Aroostook County the main mass of the Seboomook Formation has yielded only a few fossil localities, all containing fossils consistent with a Becraft-Oriskany age. In several areas of the County the Seboomook is underlain by beds containing Helderberg type faunas suggestive of a New Scotland age, but presenting the same basic correlation problem as those associated with the Seboomook of the Moose River synclinorium in Somerset County. In southern Aroostook County a number of Seboomook Formation fossil localities have yielded faunules of New Scotland aspect.

Southeast of the Kathadin batholith and its extensions to the southwest in both Somerset and Penobscot Counties the Seboomook has not yielded diagnostic fossils. However, the unpublished studies of Griffin and Ludman indicate that the Seboomook in this region overlies Late Silurian, fossiliferous units of the Slate Belt. In west-central and southwestern Maine the units mapped as Littleton Formation have not yet yielded fossils.

Connecticut River Valley region with the omission of northwestern New Hampshire:

The Connecticut River Valley region, with the exception of the two limited areas of northern New Hampshire discussed previously, has not yet yielded any diagnostic fossils.

CONCLUSIONS

This review of northern Appalachian Devonian slate problems is intended to outline and encourage geologists and paleontologists interested in Devonian stratigraphy and paleogeography to center their attention in an area basic to our better understanding of the Appalachian story during the Middle Paleozoic. It is obvious that present data do not really permit unique solutions to any of the discussed problems, despite the necessity in individual papers of defining local ground rules regarding the Devonian slate. The overall rarity of megafossils combined with the internal structural complexity of the Devonian slate and its apparent stratigraphic homogeneity make the problems seemingly very difficult. Opportunity exists in every instance for attempting a solution of the problems by means of more detailed mapping, and in some instances by microfossil studies. With this information in view it is pertinent to go over the problems cited in the Introduction.

1 and 5. The problem posed by Middle Devonian fossils immediately below the Devonian slate in the Chaudiere-Temisouata region and of Early Devonian fossils well up in the slate to the southeast in northern Maine can be investigated in several ways (the possibility for regional onlap of a simple sort is discounted here). The first is to try to obtain conodonts from the upper part of the limestone in the lower member of the St. Francis Group to determine whether Middle or Early Devonian fossils or both are present in this carbonate facies underlying the southwestern part of the Devonian slate. Second is an attempt to obtain conodonts from known scattered limestone lentils within the Devonian slate itself in the Eastern Townships to give some idea whether or not all yield Middle Devonian faunas or whether more structural complexity is present than the mapping to date indicates. Third is to try to map key cross-sections extending from the known Middle Devonian portion of the slate southeast into the known Early Devonian to try to determine whether stratigraphic hiatus or northwest onlap can best explain the known facts.

2. The most effective way to deal with the problem of comparing the known Ordovician-Early Devonian sequence in the Slate Belt of central Maine with the "Littleton" east of the Bronson Hill Anticline is through the detailed mapping and stratigraphic work now in progress.

3. The attempt to discover synclines containing Early Devonian transitional facies between the southeasternmost known, flysch-like Devonian slate (Seboomook Formation) and the largely non-marine, volcanic Devonian of the coast to the southeast involves mapping, mapping, and more mapping supplemented by well designed traverses between the areas of known Devonian rocks.

4. The working out of relationships between the fossiliferous Silurian and the Devonian slate of the Lake Temiscouata region depends primarily upon more detailed mapping between Riviere Bleue and the Cabano region to the northeast, supplemented by study of the obtained fossils.

6. A thoroughgoing program of search and study of hydrofluoric acid residues for chitinozoans and hystrichospheres from the Devonian slate involves careful section measuring in the structurally simpler areas of Devonian slate combined with study of the fossils obtained from the sections, comparison of the faunas with those obtained from better known standard sequences elsewhere, followed ultimately by intensive sampling of the entire extent of the Devonian slate. It has never been determined whether or not hydrofluoric insoluble microfossils can withstand the consequences of regional metamorphism, and if so to what grade and degree of structural complexity.

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Rb-Sr Whole Rock Ages of Silurian-Devonian Volcanics from Eastern Maine

Paul D. Fullager¹ and Michael L. Bottino²

INTRODUCTION

Within the last few years we have determined the radiometric age of several volcanic units in northeastern North America (Bottino and Fullager, 1966; Fullager and Bottino, 1968a, b). These investigations were conducted primarily for the purpose of substantiating or, if necessary, revising the Silurian-Devonian portions of the radiometric time-scale.

We have recently completed Rb-Sr analysis on three additional Silurian-Devonian units—the Quoddy, Dennys and Pembroke formations. The analyzed samples were collected in the vicinity of Eastport, Maine.

ANALYTICAL PROCEDURES

Standard Rb-Sr laboratory procedures were followed in this study. Rb and Sr concentrations were determined by isotope dilution analyses. Unspiked solutions were analyzed to measure the Sr isotopic composition. The mass spectrom-

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etry for this study was done at the Planetology Branch, Goddard Space Flight Center, Greenbelt, Maryland.

The Eimer and Amend standard SrCO₃ was analyzed three times during this study; the average Sr⁸⁷/Sr⁸⁶ ratio is 0.7078 with the ratio normalized to Sr⁸⁶/Sr⁸⁸=0.1194. The standard deviation of a single analysis of the standard SrCO₃ is ± 0.0008. These results for the Eimer and Amend SrCO₃ are consistent with our past results (Bottino and Fullager, 1966).

For the age calculations, errors of ± 3% were used for the Rb⁸⁷/Sr⁸⁶ ratios, and ± 0.4% for the Sr⁸⁷/Sr⁸⁶ ratios. These error estimates are based on duplicate analyses done in several of our other Rb-Sr whole-rock studies. The ages and Sr⁸⁷/Sr⁸⁶ initial ratios were calculated by the least squares cubic method developed by York (1966). Errors are reported as one standard deviation. The Rb⁸⁷ decay constant used is 1.39 x 10⁻¹¹ year⁻¹.

ANALYTICAL RESULTS

The fossils of the Eastport area and surrounding regions are very similar to those of Great Britain. Therefore, it has been customary to use the terminology and systematic subdivisions of the British section. Fig. 1 modified from

		BRITISH SECTION		AMERICAN SECTION			
DEVONIAN	MIDDLE DEVONIAN	EIFELIAN		ONONDAGA	ERIAN	MIDDLE DEVONIAN	
	LOWER DEVONIAN	EMSIAN	UPPER	ZONE "B" OLIVER, 1960	ULSTERIAN	LOWER DEVONIAN	
			LOWER	ESOPUS			
		SIEGENIAN	ORISKANY				
GEDINNIAN	HELLERBERG	BE CRAFT	NEW SCOTLAND	CAY-UGAN			UPPER SILURIAN
		MANLIUS-COEYMANS					
LUDLOVIAN		COBLESKILL-RONDOUT	NIAGARAN	MIDDLE SILURIAN			
WENLOCKIAN		BERTIA			ALBION	LOWER SILURIAN	
	LOWER SILURIAN	LLANOV-ERIAN	C ₆ UPPER	MEDINA			
B ₃ MIDDLE							
B ₁							
A ₄ LOWER							
		A ₁	CLINTON				

Fig. 1. Correlation of American and British standard stratigraphic sections.

Table 1. Analytical Data

Sample	Rb ppm	Sr ppm	(Rb/Sr) _{wt}	Rb ⁸⁷ /Sr ⁸⁶	Sr ⁸⁶ /Sr ⁸⁸ (Sr ⁸⁷ /Sr ⁸⁶) _N	
<i>Quoddy Fm.</i>						
# 356	46.7	63.6	0.73	2.13	0.1204	0.7188
# 361	42.1	151.1	0.28	0.81	0.1202	0.7121
# 362	164.0	74.3	2.21	6.42	0.1205	0.7430
# 363	129.4	140.4	0.92	2.67	0.1200	0.7208
# 368	15.8	201.0	0.079	0.23	0.1202	0.7056
<i>Dennys Fm.</i>						
# 380	94.8	113.9	0.83	2.41	0.1201	0.7185
# 381	76.9	112.1	0.69	2.00	0.1201	0.7163
# 384	194.3	118.0	1.65	4.78	0.1204	0.7318
# 400	46.8	115.6	0.40	1.16	0.1200	0.7119
# 406	27.7	140.2	0.20	0.58	0.1195	0.7081
<i>Pembroke Fm.</i>						
# 387	134.0	75.6	1.77	5.14	0.1204	0.7333
# 388	73.7	101.7	0.72	2.09	0.1198	0.7145
# 389	141.6	97.2	1.46	4.24	0.1200	0.7276
# 390	103.0	81.9	1.26	3.65	0.1198	0.7253
# 391	36.4	99.2	0.37	1.07	0.1198	0.7096
# 392	3.32	91.0	0.036	0.10	0.1201	0.7047
# 416	71.0	36.5	1.95	5.66	0.1200	0.7367
# 418	184.1	39.7	4.64	13.53	0.1198	0.7789
<i>Dike Cutting Eastport Fm.</i>						
# 430	137.9	90.9	1.52	4.41	0.1197	0.7316
# 432	193.8	77.9	2.49	7.24	0.1199	0.7481

Boucot *et al.* (1964), summarizes the general relationships of the British and American sections. The new results are on volcanics which are between late Llandovery and earliest Gedinnian in age. The other pertinent units we have analyzed are the Arisaig volcanics which have a minimum geologic age of early Llandovery, the Eastport volcanics of Gedinnian age, and the Hedgehog volcanics of New Scotland age.

The Quoddy Formation is considered to be late Llandovery in age (Berry and Boucot, in press). The analyzed volcanics are from a location approximately 35 miles west of the type location for the Quoddy Formation. These samples were collected within a few tens of feet above shale containing graptolites tentatively identified by W. B. N. Berry as being late Llandovery to Middle Wenlock in age (O. Gates, personal communication); for this reason these volcanics are considered to be essentially the same age as the Quoddy Formation and perhaps part of the Quoddy Formation. Five samples were analyzed (Table 1). The calculated age is 418 ± 22 m.y. and the initial Sr⁸⁷/Sr⁸⁶ ratio is 0.7058 ± 0.0009 (Fig. 2).

The Dennys Formation is stratigraphically above the Quoddy Formation and is late Llandovery or perhaps Wenlockian in age (Berry and Boucot, in press). Five samples of volcanics were analyzed (Table 1). The least squares cubic calculation gives an age of 401 ± 5 m.y. and an initial Sr⁸⁷/Sr⁸⁶ ratio of 0.7051 ± 0.0002 (Fig. 3).

The Pembroke Formation is considered to be of Pridoli Age which is post-Ludlow and pre-

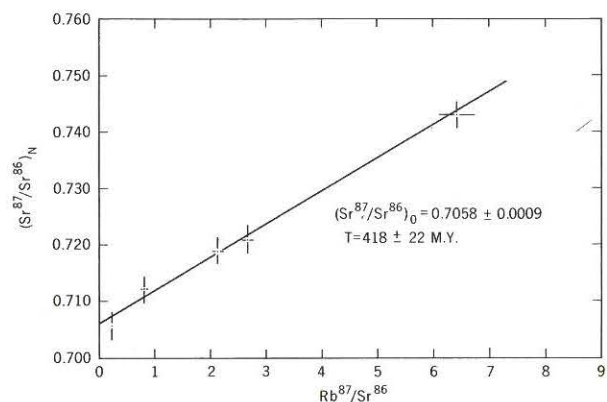


Fig. 2. Strontium-rubidium evolution diagram for whole-rock samples from the Quoddy Formation, southeastern Maine.

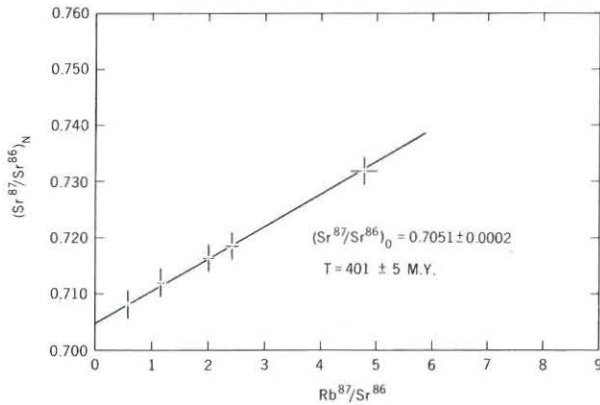


Fig. 3. Strontium-rubidium evolution diagram for whole-rock samples from the Dennys Formation, southeastern Maine.

Gedinnian (Berry and Boucot, in press). Eight samples of volcanics from this formation were analyzed (Table 1). The calculated age for the Pembroke volcanics is 402 ± 6 m.y., and the initial $\text{Sr}^{87}/\text{Sr}^{86}$ ratio is 0.7040 ± 0.0005 (Fig. 4).

The Eastport Formation is stratigraphically above the Pembroke Formation and the base of the Eastport formation is early Gedinnian in age (Berry and Boucot, in press). Two samples of a dike cutting Eastport Formation sediments were analyzed (Table 1) and give results suggesting an age of 400-420 m.y.

OTHER PERTINENT RESULTS

Eleven samples of Arisaig volcanics from Arisaig, Nova Scotia have a radiometric age of 428 ± 10 m.y. (Fullagar and Bottino, 1968a). The minimum geologic age for these volcanics is early Llandovery (Williams, 1914); the maximum geologic age is post-lower Ordovician. Consequently, this age of 428 m.y. is close

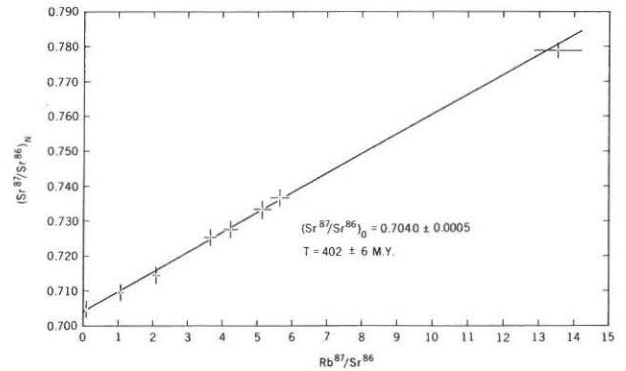


Fig. 4. Strontium-rubidium evolution diagram for whole-rock samples from the Pembroke Formation, southeastern Maine.

to a maximum age for the Ordovician-Silurian boundary.

Nine samples of volcanics from the Early Gedinnian or later Early Devonian age Eastport Formation were analyzed in a previous study (Bottino and Fullagar, 1966). The age and initial $\text{Sr}^{87}/\text{Sr}^{86}$ ratio of 412 ± 5 m.y. and 0.707 ± 0.001 given in that paper were not determined by York's squares cubic calculation (York, 1966). Re-evaluation of the data with York's calculation gives an age of 408 ± 3 m.y. and an initial $\text{Sr}^{87}/\text{Sr}^{86}$ ratio of 0.7076 ± 0.0009 .

The data for the Hedgehog formation volcanics also were reported previously (Bottino and Fullagar, 1966). These volcanics were collected about 150 miles north-northwest of Eastport, Maine, and are pertinent to this report because they have the same or nearly the same geologic age as the Eastport Formation. The Hedgehog Formation is New Scotland in age (Boucot *et al.*, 1964); this age perhaps is

Table 2. Summary of Rb-Sr Results

Volcanics	Radiometric Age*	$(\text{Sr}^{87}/\text{Sr}^{86})_0^*$	Geologic Age
Hedgehog	414 ± 5 m.y.	0.7054 ± 0.0017	New Scotland
Eastport	408 ± 3 m.y.	0.7076 ± 0.0009	Gedinnian
Pembroke	402 ± 6 m.y.	0.7040 ± 0.0005	Early Gedinnian
Dennys	401 ± 5 m.y.	0.7051 ± 0.0002	Late Llandoveryian or Wenlockian
Quoddy	418 ± 22 m.y.	0.7058 ± 0.0009	Late Llandoveryian
Arisaig	428 ± 10 m.y.	0.7060 ± 0.0009	Early Llandoveryian Minimum

*Least-squares cubic calculation (York, 1966).
Errors reported as one σ .
 $\lambda_{\text{Rb}^{87}} = 1.39 \times 10^{-11} \text{ yr}^{-1}$.

slightly younger than that of the Eastport Formation. The 413 ± 10 m.y. age and 0.706 ± 0.002 initial $\text{Sr}^{87}/\text{Sr}^{86}$ ratio previously reported were not determined by the least squares cubic calculation. Using the least squares cubic calculation to re-evaluate the data gives an age of 414 ± 5 m.y. and an initial $\text{Sr}^{87}/\text{Sr}^{86}$ ratio of 0.7054 ± 0.0017 .

DISCUSSION AND CONCLUSIONS

Table 2 summarizes the pertinent Rb-Sr data. Considering the analytical uncertainties (reported as one σ), the ages for the Quoddy, Dennys and Pembroke Formations are consistent with the ages for the Arisaig, Eastport and Hedgehog volcanics. More samples, and samples with greater ranges of Rb/Sr ratios were analyzed in the studies of the Arisaig, Eastport and Hedgehog volcanics; thus, the results for these volcanics probably are more reliable than the new results reported in this study.

The Hedgehog and Eastport volcanics are above but close to the Silurian-Devonian boundary. Even with the recalculation of the ages, our previous estimate of 413 ± 5 m.y. (Bottino and Fullagar, 1966) as a minimum age for the Silurian-Devonian boundary seems reasonable. Comparing the results in Table 2, but especially those for the Hedgehog, Eastport and Arisaig volcanics, suggests that the Silurian period in northeastern North America had a length of only 10-20 m.y.

ACKNOWLEDGEMENTS

Professor Olcott Gates of Fredonia State College, New York, is mapping the Eastport, Maine area. We are very grateful for Professor Gates' assistance; he provided suggestions for collecting sites and accompanied us in the field when we collected these samples. This study was supported by NSF grants GP-3605 and GA-1338.

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Structural and Stratigraphic Studies in the Sawyer Mountain Area, York County, Maine

Richard A. Gilman¹

INTRODUCTION

The Sawyer Mountain map area (Fig. 1) is located in Southwestern Maine at the mutual corners of the Kezar Falls, Newfield, Sebago Lake, and Buxton quadrangles (15 min. series). The metasediments are mostly pelites within the sillimanite metamorphic zone and belong to the Berwick and Rindgemere formations of Siluro-Devonian (?) and Devonian (?) ages respectively. Igneous rocks are represented by a small body of foliated biotite granodiorite and a larger stock of binary granite. These are believed to be correlative with the New Hampshire Plutonic Series of Devonian age and will not be considered further.

The purpose of the paper is to describe the structural styles and stratigraphy of the area and to offer a tentative structural-stratigraphic interpretation. The Sawyer Mountain area appears to be one of the best areas in the four quadrangles for detailed study, and it will undoubtedly prove to be a critical region for the understanding of broader scale structural and stratigraphic relationships.

The work supported by the Maine Geological Survey, represents the first detailed geologic mapping of the metasediments in the region. The writer was assisted in the mapping of the Sawyer Mtn. region by Michael Lis and has benefited by many discussions with Dr. Arthur M. Hussey II.

STRATIGRAPHY

The metasedimentary rocks of the area have been assigned to two formations: the Berwick formation of Siluro-Devonian(?) age and the Rindgemere formation of Early Devonian(?) age.

The Berwick formation consists of light to dark blue-grey, fine grained, granular biotite schists with uniform texture and pronounced bedding on the scale of 0.5 to 2.0 inches. Individual beds frequently show laminations due to concentrations of biotite. The rock consists

primarily of quartz and plagioclase (An₃₀₋₄₀) with about twenty per cent biotite and minor garnet. The color, bedding and mineralogy of this rock is sufficient to distinguish it from other metasediments in the area. Lithologically identical rocks have been mapped as Berwick formation in the Buxton quadrangle to the east (Hussey, personal communication). It appears to be possible to connect the rocks mapped as Berwick in the Sawyer Mountain area with those farther east in the Buxton quadrangle but intervening outcrops are scarce due to heavy glacial cover and low relief. The lack of structural and stratigraphic control prohibits thickness estimates.

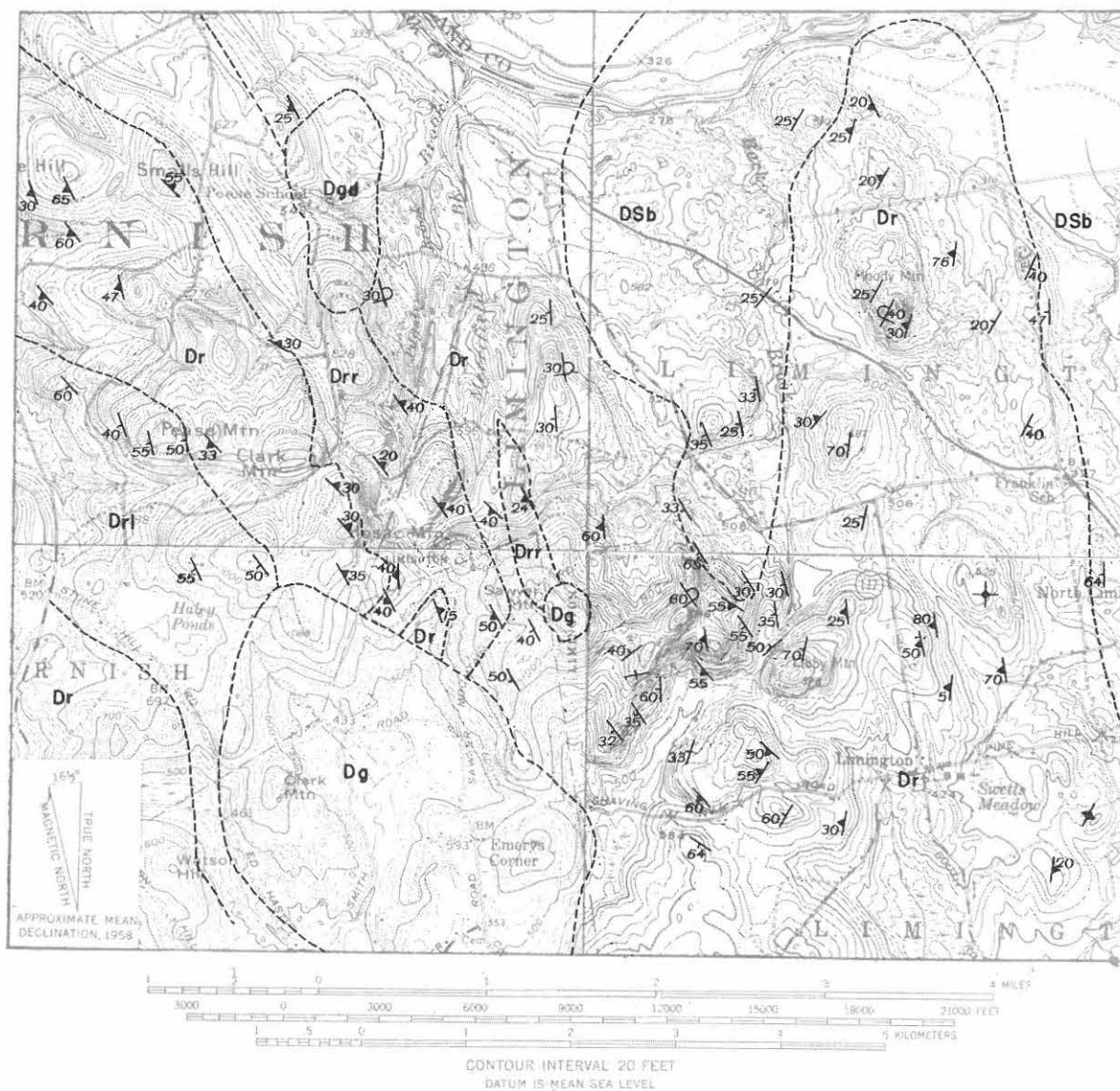
The most accessible outcrops of this rock type are found just east of the map area in the Saco River at Steep Falls in the Buxton quadrangle (Gilman, 1965).

The remaining metasedimentary rocks are collectively assigned to the Rindgemere formation in accordance with Hussey's work in Southern York County (Hussey 1962). This formation consists of three distinct lithologic types that appear to be mapable units but as yet insufficient regional data are available to warrant their classification as separate formations. The three lithologic types are (1) grey-brown, medium grained mica schist and migmatite, (2) rusty weathering, sulfide-bearing quartz-mica schist, and (3) green-grey, bedded, lime silicate granofels. Each of these types is briefly described below.

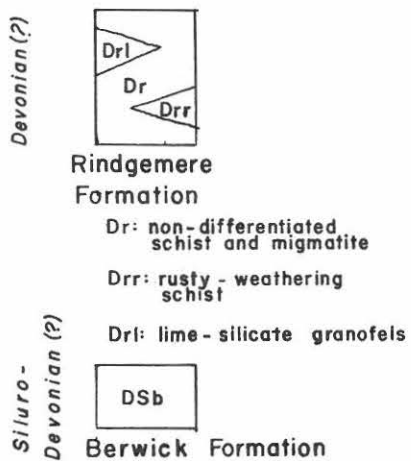
Grey-brown mica schists and migmatites are the most abundant rock type in the area. There is considerable variation between migmatitic and non-migmatitic types, but as yet these varieties cannot be mapped separately, nor can one determine the stratigraphic positions of a given type within this group. The two major lithologies of this group are (1) well-bedded, brown-grey, medium grained mica schist, and (2) medium-to-coarse grained, brown or grey migmatites.

The well-bedded mica schist is distinguished primarily by its 1/2 to 3 inch scale cyclic bedding which is frequently graded (Figs. 2 and

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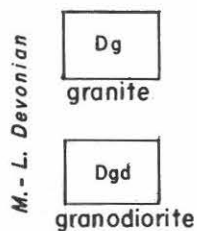


STRATIFIED ROCKS



PLUTONIC ROCKS

New Hampshire Plutonic Series



Approximate geologic contact

Bedding

Schistosity

Fig. 1. Geologic map of the Sawyer Mtn. area, Maine.

3). Graded beds consist of tan-brown granular quartzose schist grading upward into grey micaceous schist. This unit is also unique in that it is the only rock type in which staurolite and occasionally andalusite have been found in addition to the sillimanite generally present in the area. The best exposures of this well-bedded schist are to be found on the west side of Libby Mountain which is easily reached by a discontinued road starting from route 117 north of Limington. Moody Mountain also provides excellent exposures of this rock type.

Rocks of a general well-bedded character have been found to the north in the Saddleback Hills (Kezar Falls-Sebago Lake quadrangles) and to the south in the vicinity of North Alfred (Gilman, R. A., unpublished reports, Maine Geological Survey). In both cases the exposures are within two or three miles of the contact with rocks of the Berwick formation. Although exposures do not allow detailed correlations, I feel that the well-bedded schists are generally found only near the contact with the underlying Berwick formation.



Fig. 2. Well bedded mica schist with minor quartzfeldspar pods: Moody Mountain.

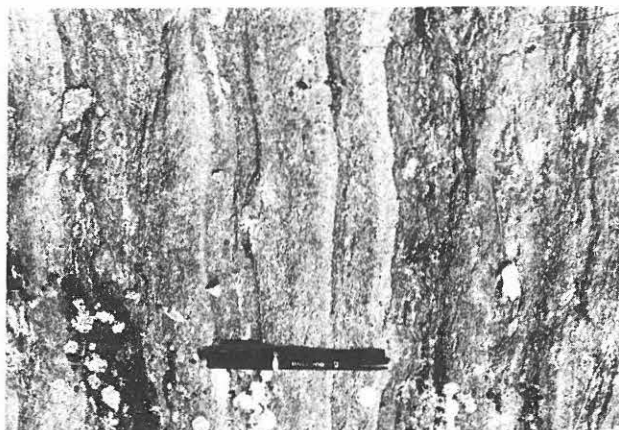


Fig. 3. Well bedded schist showing graded bedding: tops toward pen cap: Moody Mtn.

The migmatitic rocks are generally various shades of grey to brown, are medium to coarse grained, and contain abundant quartzofeldspathic pods and stringers. In general they do not show bedding, although in some cases compositional layering is present and may represent original bedding. This rock type is interbedded with all the other lithologies within the Rindgemere formation and distinguishing between the several layers on the basis of lithology is presently impossible. Even an approximate original thickness estimate for the schists and migmatite is meaningless due to lack of detailed structural control and the demonstrable isoclinal folding (such as found on Libby Mountain).

Sulfide-bearing quartz-mica schist is easily recognized in outcrop by its characteristic rusty or sometimes yellow and black weathering surface. The rock is composed largely of quartz-plagioclase-muscovite-biotite and sufficient sulfide (two to five percent) to impart the rusty weathering surface. Rusty units are also occasionally found within the schists and migmatites described earlier, but have been mapped separately only where they are the predominant rock type. The rusty unit thus defined is contained in a northwest trending belt passing between Sawyer Mountain and Hosac Mountain. This belt is known to extend as far north as Mt. Cutler (Kezar Falls quadrangle) but it has not been found south of the Clark Mountain granite body. Outcrop patterns would suggest this to be a rather thin unit (few hundred feet?) within the Rindgemere formation.

Lime-silicate granofels is characteristically a light grey-green color, is generally fine-grained and usually shows well preserved thin bed-

ding but may also be massive. One-to-three inch interbeds of fine grained granular biotite schists are frequently present. Major minerals are quartz, plagioclase, and diopside. Minor minerals consist of garnet, sphene, calcite, and occasionally large clots of vesuvianite. Bedding is frequently severely contorted into disharmonic folds. The field occurrence of this rock suggests that it is a thin unit within the Rindgemere, perhaps on the order of one or two hundred feet thick. The unit is well exposed on the west side of Pease Mountain and in roadcut ledges along Maine Highway 5 southwest of Pease Mountain.

STRUCTURE OF THE METASEDIMENTS

A detailed structural evaluation of the Sawyer Mountain area is not yet possible. However, the results of field work in the summer of 1968 clearly indicate a complex structural history. The following discussion will outline the observed structural styles thus far recognized and then formulate a tentative structural-stratigraphic framework for the Sawyer Mountain area.

Regional folds of bedding:

The Berwick-Rindgemere contact is the best in the area for purposes of determining the major fold patterns. Although never clearly exposed, this contact can be traced fairly continuously in much of the map area. The structural symbols on the geologic map show the locations where control is best documented. That the contact loops around between Sawyer and Libby Mountains is well substantiated. Although the contact is poorly exposed to the north, scattered outcrops suggest that it again loops around north of Moody Mountain. It is then fairly reliably traced down the east side of the map.

This looping pattern suggests a pair of plunging folds. While there remain some unresolved problems with this interpretation, it seems better substantiated than alternatives such as facies intertonguing or faulting. There is no direct field evidence of faulting and the geometry of minor folds in the valley west of Libby Mountain (Fig. 4) agrees well with these inferred major folds. This regional pattern is therefore interpreted as one of overturned, isoclinal(?) folds with axial surfaces dipping to the west. This is suggested by the structural symbols on the map where it can be seen that

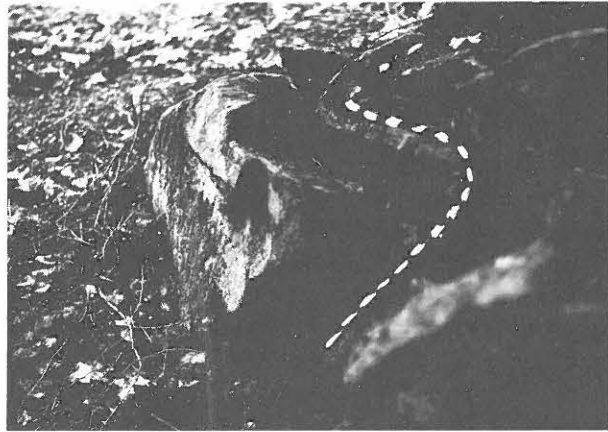


Fig. 4. Overturned fold in valley west of Libby Mtn. Fold axis trends S85°W; plunges at 11°. White line traces bedding.

strikes are generally either northwest or northeast but dips are nearly always to the west.

Considering the relative ages of the Berwick and younger Rindgemere rocks, the exposures of the Berwick formation northwest of Libby Mountain occupy a plunging anticline. Similarly the Rindgemere to the east occupies a syncline whose axis passes essentially through Moody Mountain. The attitude of axial plunge is west to southwest at about 30° as determined by plotting data from outcrops close to the contact on a stereographic net. Smaller scale folds such as that shown in Figure 4, found in the valley west of Libby Mountain also show a west-southwest plunge of overturned folds.

The interpretation of the Moody Mountain syncline is open to question because, while many exposures on Moody Mountain show graded bedding with tops west (Fig. 3), none as yet have been found with tops east corresponding to the other limb of the syncline. The syncline cannot be traced south of Limington, where exposures are scarce due to thick glacial cover and low relief.

West of Sawyer Mountain the units of the Rindgemere formation show a consistent westerly dip of both foliation and bedding. The few exceptions to this where easterly dips are recorded are perhaps related to a separate folding episode along northwest trends suggested by field work in the western parts of the Kezar Falls and Newfield quadrangles.

Small scale folding:

In addition to the small folds mentioned above, several styles of small scale folds have been found.



Fig. 5. Chevron folds with nearly horizontal axial plane. Fold axis trends S25°W, plunges at 14°, northeast side of Libby Mtn.

On the west side of Libby Mountain excellent exposures of well bedded schist show repeated reversals of tops directions as determined from graded bedding. One section of about two hundred feet across strike contained eight reasonably certain reversals in tops directions. It is therefore interpreted that this section is much thickened by isoclinal folding, however only one recognizable isoclinal fold could be found. How this is related to the larger scale fold pattern discussed above is not clear.

On the east side of Libby Mountain, on Moody Mountain, and in several other locations a chevron style of folding with a nearly horizontal axial surface has been observed (Fig. 5). This clearly deforms the schistosity as well as bedding and is tentatively considered as a separate folding episode, perhaps late in the metamorphic-tectonic history.

In addition, small asymmetric folds are common throughout the metasediments. As yet no systematic pattern of these folds has been found.

Finally, the schistosity commonly shows minute crenulations frequently in more than one direction.

Tentative Structural-Stratigraphic Framework:

From the data presently available it is my opinion that the major outcrop patterns in the Sawyer Mountain map area are due to tight (perhaps isoclinal) overturned folds plunging

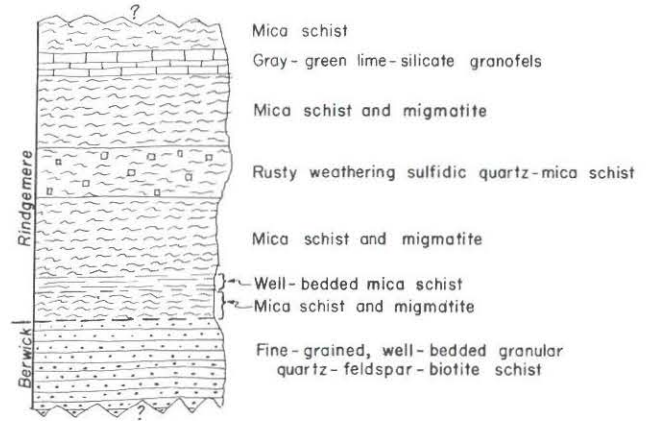


Fig. 6. Generalized stratigraphic section for the Sawyer Mtn. area. Relative thicknesses are based on map patterns and field exposures.

in a general southwesterly direction. It is also suggested that the general westerly dips west of Sawyer Mountain provide a simplified stratigraphic section from the Berwick formation at the base to progressively younger units within the Rindgemere formation to the west.

A generalized stratigraphic section is shown in Figure 6. Of special importance is the occurrence of well-bedded schists near the base of the Rindgemere, and the general interbedding of mica schist and migmatite with the more distinctive rusty schist and lime silicate granulate.

While beyond the scope of this paper, regional mapping in the western parts of the Kezar Falls and Newfield quadrangles suggests the presence of a large synclinal structure with a generalized north-south trend. The westward younging of the rocks in the Sawyer Mountain area is therefore consistent with the regional patterns as they are presently interpreted.

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Observations on the Origin and Development of the Wells Beach Area, Maine

Arthur M. Hussey, II¹

INTRODUCTION

Wells Beach is located along the southwestern coast of Maine, approximately 20 miles north of Portsmouth, New Hampshire, and 30 miles south of Portland, Maine, (Fig. 1). It is a segment of a barrier beach chain extending a distance of 9.7 miles from Ogunquit on the south to Kennebunk Beach on the north. The continuity of this fine series of beaches is interrupted at several places by outcrops of bedrock and glacial and postglacial surficial deposits. For purposes of discussion in this article, Wells Beach is taken as that part of the barrier chain extending from Moody Point north to the mouth of the Webhannet River (Fig. 2). Wells Beach is separated from the mainland by a tidal marsh approximately one mile wide through which the tidal channel of the Webhannet River and tributaries meander.

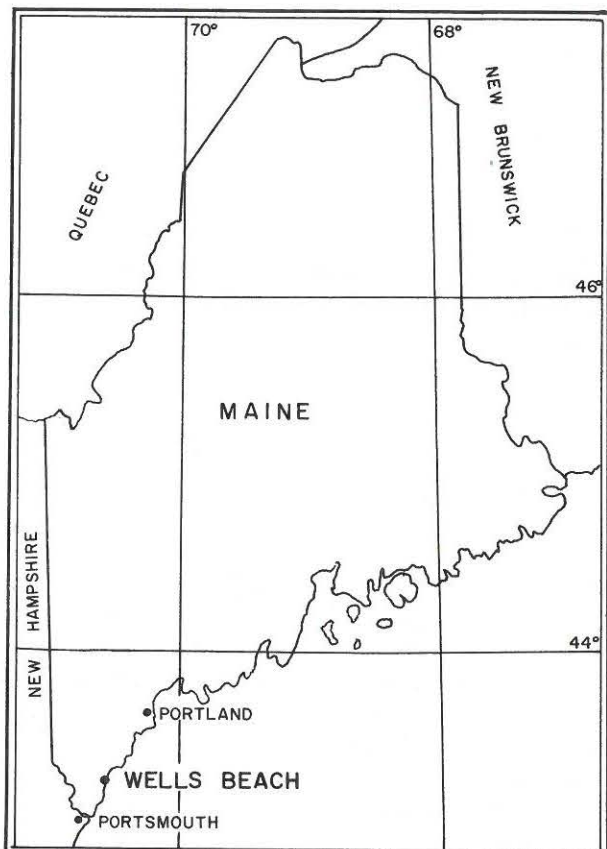


Fig. 1. Location of Wells Beach, Maine.

The writer has made intermittent observations on this segment of the chain for the past 20 years, particularly during the Springs of 1953, 1958, 1962, and 1969 when northeast storms severely eroded portions of the middle segment (Fig. 2) of the beach, exposing limonite-manganese oxide cemented gravel, marine clay, and peat with tree stumps rooted in place. These observations, along with the excellent regional study of surficial glacial and post-glacial stratigraphy by Bloom (1960), provide the background from which to frame some conclusions as to the origin and development of Wells Beach.

COASTAL MORPHOLOGY AND DEPOSITS

Beach Segments. Wells Beach consists of three segments (Fig. 2) partially separated by exposures of bedrock, glacial drift, marine clay, and cobbleboulder lag deposits. The southern segment extends in an arc from Moody Point to the next point north which for purposes of discussion is referred to as Point "A". This is the narrowest segment of the Beach, averaging slightly less than 200 feet in width. The original character of the Beach prior to residential development is obscured by a breakwater, houses, and a road which occupy the zone that would correspond to the upper part of the foreshore and all of the backshore. Relationships at the southern end of the breakwater near Moody Point suggest that the Beach in this part of the segment consisted of a narrow foreshore, which is a bedrock wave-cut bench with thin cobble and boulder veneer. The backshore was a narrow storm beach presumably with no coastal dune development behind it. Toward the northern end, well sorted beach sand is found on the lower part of the foreshore, and the upper part of the foreshore to the breakwater consists of well-rounded cobbles apparently not resting on a bedrock bench. Northeast storms periodically uncover compact peat with abund-

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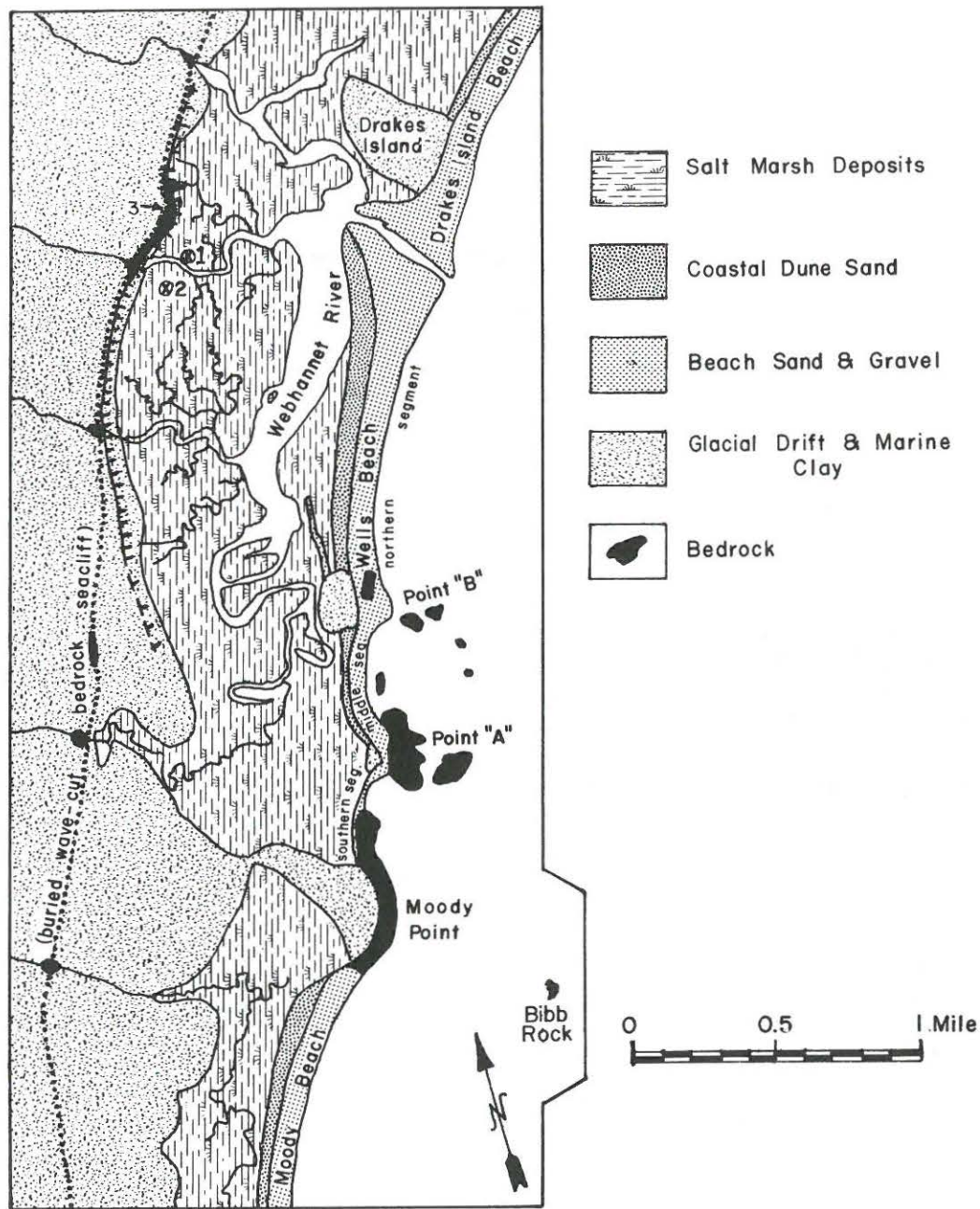


Fig. 2. Coastal morphological features and deposits in the area of Wells Beach.

ant wood fragments. The peat is exposed in a wave-smoothed pavement at the base of the breakwater about 6 feet below high tide level.

The middle segment extends northward in an arc from Point "A" to Point "B" (Fig. 2). At present, this segment is characterized by a foreshore of fine beach sand; a backshore of cobbles of diverse rock type, forming a relatively thin veneer over coarse feldspathic beach sand; and a zone of coastal dunes behind the backshore,

upon which the houses and road have been built. A lag deposit of cobbles and small boulders of diverse size and lithic composition is present along the lowest part of the foreshore and the adjacent shallow off-shore waters. This deposit is situated in the lee of an offshore ledge referred to locally as "Whalesback" (Fig. 3). The cobble veneer on the backshore has been gradually delivered from the south by predominantly northward longshore drift. The writer

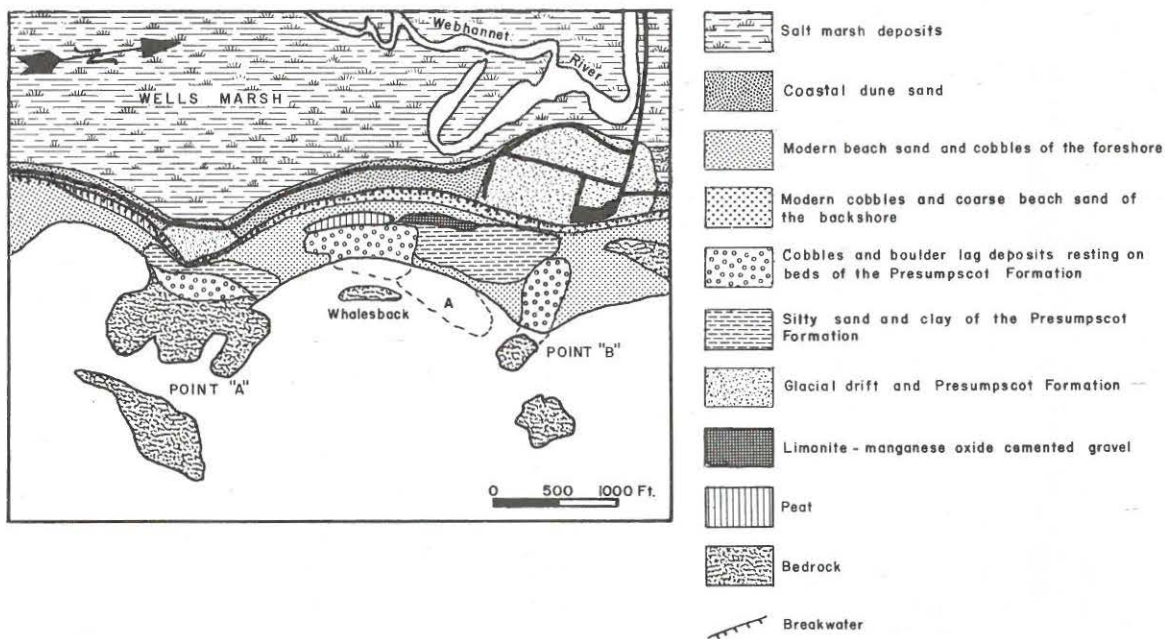


Fig. 3. Sketch map showing the distribution of surficial deposits exposed after the northeast storms of February and March, 1969. Area A marks the site of a bar formed by the sand eroded from the beach.

vividly recalls from summer visits that until about 30 years ago, the backshore consisted of the same quality of fine beach sand as the present foreshore, anchored by beach grass, and that the level of the backshore beach at the breakwaters in front of the houses was approximately four feet lower than at present. Cobbles were sparingly present at the lower edge of the backshore. During the last 30 years longshore movement of cobbles on the backshore has become very active. The cobbles, presumably derived from the southern segment of the beach and from the lag concentration in front of Point "A", have migrated gradually northward and are now found about halfway along the length of the northern segment of the beach. The reason for this acceleration of long-shore cobble transport is clearly related to the removal of an intertidal rip-rap groin that formerly existed at Point "A", extending from the breakwater there to ledges in front of the Point. The groin served to check northward longshore drift when the beach was submerged at high tide. According to accounts of older inhabitants of the beach, the groin was gradually broken up by northeast storms 35-40 years ago, and was not repaired. About 15 years ago most of the blocks of the groin, then lying helter-skelter on the beach

were bulldozed up against the breakwater at the Point in an attempt to protect the deteriorating breakwater against storm-wave damage, thus removing all obstructions to long shore current movement. An additional cause for accelerated longshore cobble transport is the position at the edge of the foreshore of the breakwater at Point "A" and along the southern segment of the beach. There, the breakwater interrupts the normal wave swash at high tide. Rather than expending energy in streaming up and back upon an equilibrium slope, the swash is thrown into turbulence upon hitting the foot of the breakwater resulting in more pronounced sediment movement by longshore drift.

Erosion during severe northeast storms, particularly those occurring during the early Spring of the year, has repeatedly shown that the modern beach sediments of the middle segment of Wells Beach are very thin, generally less than four feet over most of the foreshore. During the closely spaced sequence of violent northeast storms of February and early March 1969, the foreshore of this segment was eroded by surf action to a greater degree than heretofore noted. The result was extensive exposures of the following types of deposits that comprise the foundation of the Beach (Fig. 3):

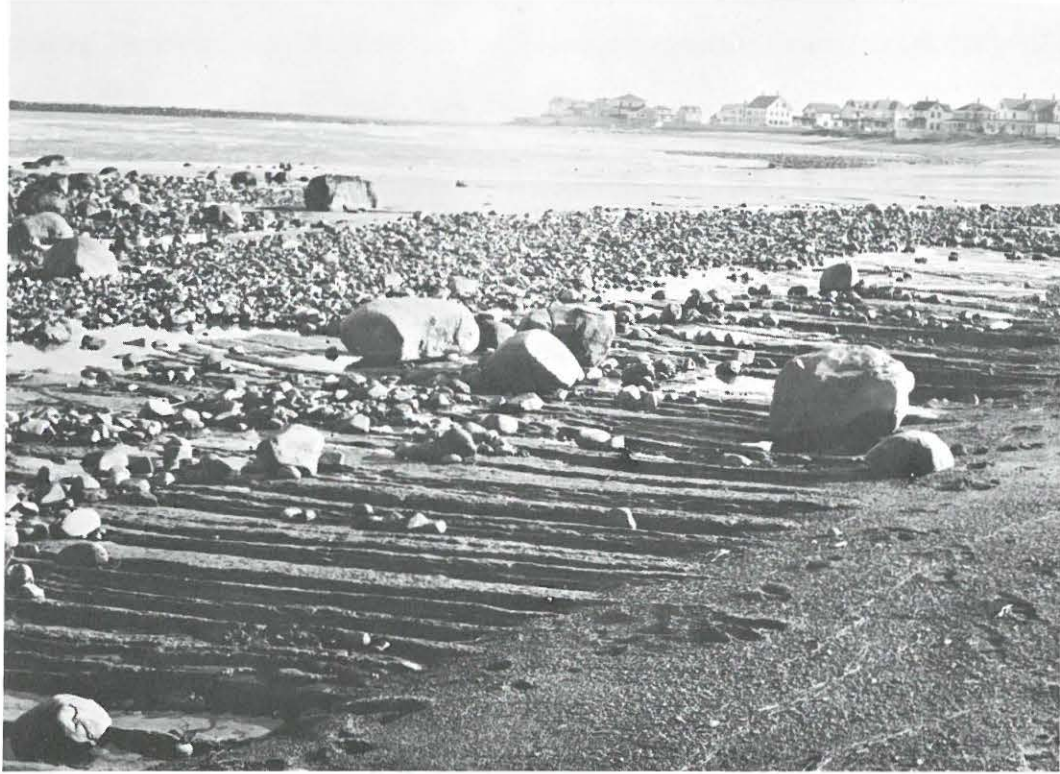


Fig. 4. Exposure of gently dipping alternating silty sand and silty clay of the Presumpscot Formation (commonly referred to as post-glacial marine clay). Smaller boulders and cobbles form a local veneer on top of marine clay, and are usually covered by modern beach sand in the summer. The larger boulders are exposed in the summer, and the greater exposure of the boulder in the right foreground indicates that approximately three feet of modern beach sand were removed during a single northeast storm of two days duration. View looking south along the middle segment of the beach from Point "B", 1 April, 1958.

1. Thinly interbedded buff silty sand and silty clay, inclined gently in varying directions (Figs. 4 and 5), and correlated with the Presumpscot Formation (generally known as the post-glacial marine clay) of Bloom (1960). These beds were exposed over approximately 40% of the foreshore of the middle segment of the Beach and were found to be present beneath peat, cobble veneers, and cemented gravel over another 30% of the foreshore.

2. A layer ranging from six inches to four feet thick, of compact, polymictic gravel cemented by limonite and manganese oxide, resting on Presumpscot beds. The layer is thickest at the landward margin of the foreshore where it has been least eroded by wave action (Fig. 6). Scattered patches (Fig. 7) were observed over most of the foreshore where Presumpscot beds were exposed, but the layer was not encountered beneath the peat in a 6 foot hand auger hole drilled through this peat and into the underlying silty sand and clay. This gravel is probably a very localized deposit, origin uncertain.

3. Dark brown compact salt marsh peat with abundant remains of *Spartina* (Fig. 8), beneath which is greenish gray sand becoming silty and clay-rich at a depth of about 4 feet. Rooted in the peat and sand are numerous stumps, mostly of white pine (*Pinus Strobus*), varying from an inch to a foot in diameter (Fig. 9). A radiocarbon age of 2810 ± 200 has been

obtained on a stump from this locality (Hussey, 1959). These stumps are found at a level approximately 6 feet below mean high water. The elevation of exposure of the peat suggests that it should have originally extended over the cemented gravel noted above. If it did, this might suggest that the limonite-manganese oxide cement of the gravel resulted from a bog-iron-type precipitation.

4. Concentrations of loose cobbles and boulders of diverse lithologies which are most heavily localized in the lee of offshore ledges (Fig. 3). The cobble concentration on the foreshore in the lee of Whalesback Ledge (Fig. 3) is approximately 1 foot thick beneath which is one foot of coarse sand becoming silty at a greater depth (probably the same as the Presumpscot beds exposed nearby). These cobble and boulder concentrations may represent lag concentrations of the coarser material from glacial drift deposits that formerly extended over the nearby ledges. The finest materials were removed by wave action and longshore drift.

The northern segment of Wells Beach extends one and one-half miles from Point "B" to the mouth of the Webhannet River. Drakes Island Beach (Fig. 2) north of the Webhannet River mouth is a continuation of this segment.

The northern segment is essentially a large spit built out from Point "B" by northward long shore currents. A 500- to 800- foot wide foreshore is composed of fine beach sand. The backshore of the northern half of this segment is of the same material as the foreshore, but the southern half is covered by the same thin cobble veneer that is present in the middle segment. A fine dune sand zone is located the whole length of the segment behind the backshore. Thickness of the modern beach deposits is not known.

Moody Point. Moody Point is a linear height of land approximately $\frac{1}{4}$ mile long trending northwest and rising 27 feet above sea level and above the surrounding tidal marsh to the north and south. Moody Point connects with the mainland over a low narrow saddle not over five

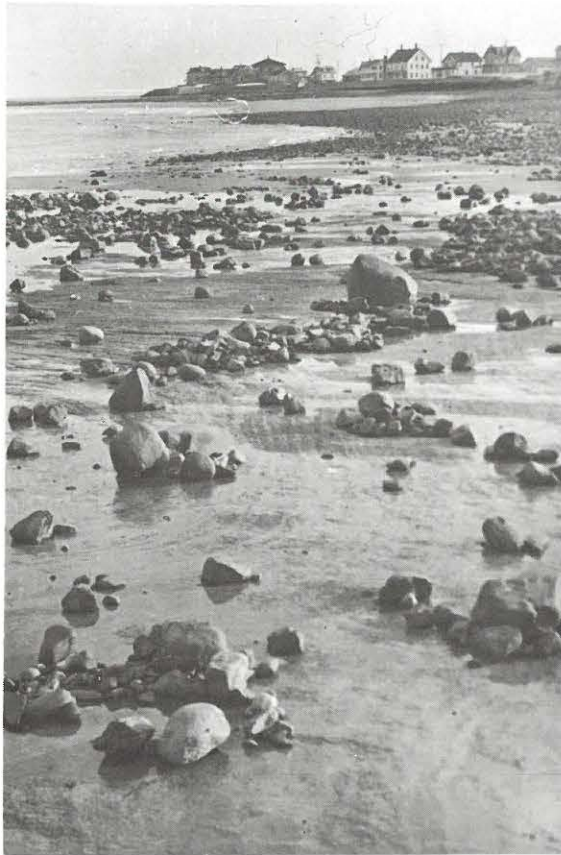


Fig. 5. Extensive exposure of the Presumpscot over a large portion of the foreshore near the northern end of the middle segment of the Beach after the storms of February and March, 1969. Scattered small boulders and cobbles are in part transported from the lag concentration in the background by longshore currents, and in part derived *in situ* from limonite-manganese-oxide cemented gravel. View looking south along the middle segment, 20 March, 1969.



Fig. 6. Very well indurated gravel cemented by limonite and manganese oxide exposed near the upper edge of the foreshore of the middle segment. The lighter colored cobbles are derived not from the gravel but from the cobble deposit localized in front of the breakwater 100 feet in the direction of the lower left corner of the photograph. April, 1955.

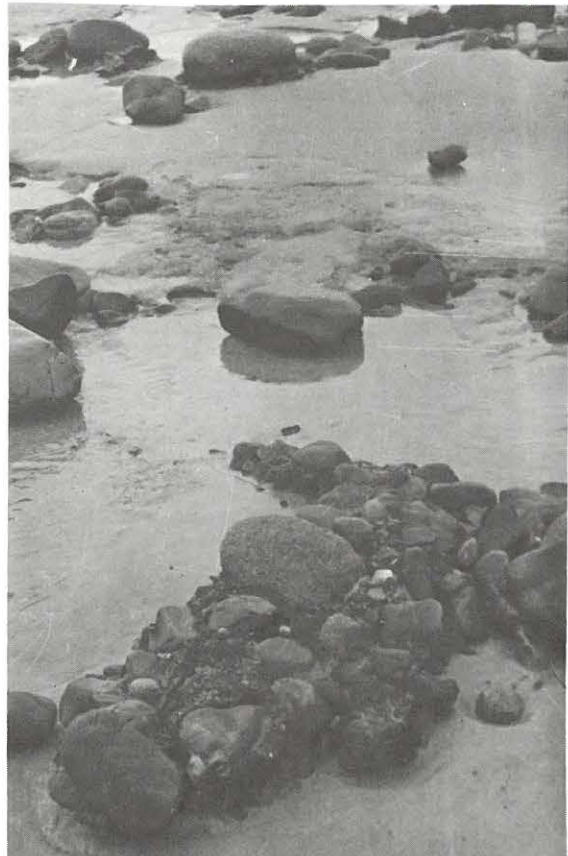


Fig. 7. Scattered remnant of the cemented gravel resting on Presumpscot pebbly silt and silty clay, exposed on the lower part of the foreshore of the middle segment of Wells Beach, 20 March, 1969.



Fig. 8. Exposure on the middle part of the foreshore of the middle segment of Wells Beach, of compact salt marsh peat. To the right is the lag veneer of cobbles in lee of Whalesback ledge. These cobbles are not cemented and rest upon sand which grades downward into silty clay of the Presumpscot Formation. The peat does not rest on the cobble veneer. Two layers of peat shown in the photograph are separated by 2 inches of pebbly sand. View looking north toward Point "B", 20 March, 1969.

feet above the surrounding marshes. The seaward edge of the point is a wave-cut bench formed on outcrops of the Kittery Formation and basalt dikes that intrude the formation. On the south side of the point, longshore drift is predominantly to the south while on the north side it is to the north. This is probably brought about by a line of shoals extending to Bibb Rock (Fig. 2). A thin veneer of cobbles and boulders of diverse lithologies rests on the wave-cut bench in front of the Point. The heterogeneity of rock types present suggests that the cobbles and boulders represent a lag deposit resulting from erosion of glacial drift resting on bedrock of the area of Moody Point, as well as contributions from wave plantation to form the bedrock bench.

Points "A" and "B". Points "A" and "B" owe their position and prominence to the extensive ledges of Kittery Formation both adjacent to the shore and in the shallow offshore zone. At Point "A", after the storms of 1958 and 1969, clay similar to that mapped by Bloom (1960) as Presumpscot Formation was extensively exposed. The lowest part of this clay exposed is very well bedded, and grades up into material containing numerous oxidized and well-weathered pebbles and cobbles that easily disintegrate under a hammer blow. Upon both the cobbly and the well-bedded clay is a discontinuous veneer of cobbles and small boulders that may be essentially a lag deposit. The landward part of



Fig. 9. Small intertidal tree stump rooted in coarse sand that lies beneath the peat, middle segment of Wells Beach. Thin layer of modern beach sand and cobbles covers the sand in which the stumps are rooted. April, 1955.

the point is a narrow prominence rising 15 feet above the beach, and is now protected by a breakwater. Reports of local residents indicate that prior to the construction of the breakwater, this point was subjected to relatively rapid erosion. Evidence for recent rapid erosion can be seen on ledges accessible at low tide in front of the Point. There, very fresh glacial striations and crescentic fracture marks are present on the lower three feet of a prominent 15 foot wide diabase dike that protrudes about 10 feet above a cobble and boulder lag deposit (Fig. 10). The higher parts of the glacially smoothed surface of the dike show marked pitting which has obliterated the individual striae. Such pitting is probably due to a combination of weathering and the surf action that constantly affects the ledge at high tide, and should likely be produced quite rapidly after exposure. For this reason it is felt that the lower three feet of the dike have been exposed very recently by wave erosion of overlying deposits.

The material underlying Point "A" as exposed briefly in an excavation near the crest of the hill, is apparently crudely stratified glacial drift. This drift overlies clay (which is exposed on the foreshore in front of the prominence, and may then correlate with drift of the Kennebunk ice advance of Bloom (1960). The point may represent a small remnant of a former hill much like that of Great Hill in Kennebunk Beach described by Bloom (1960, p. 71).

Point "B" is a more extensive prominence than Point "A". As at "A" cobbles and small boulders of diverse lithologies, rest on gently inclined



Fig. 10. Fresh glacial striations with crescentic fracture marks on lower 3 feet of diabase dike on ledges in front of Point "A". An abrupt line separates the portion with the fresh striae from the higher parts of the glacially smoothed dike where the surface has been etched by weathering and corrosion by surf action (upper left). Fountain pen in middle for scale. June, 1952.

beds of alternating silty sand and silty clay of the Presumpscot Formation (Fig. 4), forming a lag deposit on the beach and nearshore waters in lee of ledges which extend offshore approximately 1/2 mile (Fig. 2). Bedrock crops out through the sand of the foreshore at the northern edge of the Point. Just south of the exposures of bedrock are numerous large boulders of diverse lithologies scattered over the foreshore (Fig. 4). The nature of the material comprising the landward part of the Point is not known; however, from the presence of the cobble-small boulder veneer on the foreshore in front of the point, the writer feels that the higher parts of the Point may be composed of glacial gravel overlying marine beds of the Presumpscot Formation. It would thus be similar to Point "A". The occurrence of the large boulders on the beach in front of the Point is probably the result of wave erosion of boulder-bearing gravels, and distribution of the finer fractions elsewhere over the beach, thus lowering the larger boulders onto the beach without appreciable dispersal. The absence of large boulders over the higher parts of the prominence at Point "B" may be due to burial of the boulder-bearing gravels by glacial outwash deposits.

At the northeast edge of the Point, a narrow ridge of sand, rising about three feet above the level of the marsh, extends northward approximately 800 feet. Marsh peat and silt thins against the sand on both sides, indicating that the ridge was formed before the marsh had been built up to the present level. This ridge clearly represents a sand spit built out from Point "B" prior to formation of the salt marsh, and at times when sea level was somewhat higher.

Wells Marsh. Wells Marsh extends northward from Moody Point separating Wells Beach from the mainland. It extends around Drakes Island (Fig. 2) and terminates essentially in the area shown at the edge of Figure 2. Behind the northern segment of the Beach the meandering tidal channel of Webhannet River is very wide, and at low tide extensive mud flats are exposed.

Salt marsh peat and organic-rich silt are exposed at the surface over the extent of the marsh with the exception of the area of the tidal channels. These deposits are, however, very thin (see Fig. 11) and the salt marsh peat is not found lower than the present level of the mud flats in the tidal channels.

At several localities in the marsh, intertidal tree stumps rooted in peat are present, all at a level of approximately 2.5 feet below the level of the marsh which corresponds to high tide level. One stump from locality 1, Figure 2, collected by Dr. Robert Dow, Maine Department of Sea and Shore Fisheries, gave a radiocarbon age of 2980 + 180 years (Hussey, 1959), which is very close to the age determined for the stumps intermittently exposed on the beach.

The Mainland. The mainland behind the marsh is a gently undulating surface underlain by a thin blanket of the Presumpscot Formation over crudely stratified glacial drift. Outcrops of the Kittery Formation are locally common. Most streams entering from the mainland onto the level of the marsh descend abruptly along a line of bedrock cascades or falls which are aligned with a 40 feet high cusped bedrock-defended scarp fronting directly on the marsh (Locality 3, Fig. 2). The writer suggests that this alignment of cascades and the scarp represents a former wave-cut cliff now partially buried by glacial and post-glacial sediments. The level of the bedrock bench at the foot of the cliff is not known.

South of the bedrock scarp, for a distance of 1.3 miles, a conspicuous bench generally less than 300 feet wide and sloping from the marsh upward to the west to an elevation of 16-20 feet (Bloom, 1960, p. 109), marks the edge of the mainland. The bench gives way to a scarp 20-40 feet high developed in clay of the Presumpscot Formation and crudely stratified drift, and for a distance of about 1.5 miles closely corresponds to the inferred bedrock seacliff noted above. South of this, the scarp on the surficial material diverges eastward from the bedrock scarp. A similar bench and scarp developed on unconsolidated sediments extends approximately 0.5 mile north of the bedrock-defended scarp at Locality 3 (Fig. 2). Significantly, the surficial scarp is most strongly developed along the edge of the marsh just opposite the northern segment of Wells Beach. It loses definition, and the mainland surface descends evenly to the level of the marsh in the area opposite Points "A" and "B" and Drakes Island.

Bloom (1960, p. 108) suggests that this scarp of surficial material represents a wave-cut cliff, and the bench below, a wave-cut bench produced "at times when sea level was higher than present relative to land, perhaps during post-glacial progressive emergence." The coin-

cidental fading of the scarp in areas just landward of Points "A" and "B" and Drakes Island, and its well-developed character landward of the northern segment of Wells Beach supports a wave-erosional origin at a time prior to the development of the beach, at which time wave energy could reach the mainland unimpeded.

STAGES IN THE DEVELOPMENT OF THE WELLS BEACH COASTAL AREA

The stages in the development of the coastal area in the vicinity of Wells Beach are illustrated in Figures 12-14 and Figure 2. Coastal features described above which are pertinent to the reconstruction of these stages are:

(1) a wave-cut bedrock sea cliff now buried by glacial and post-glacial sediments;

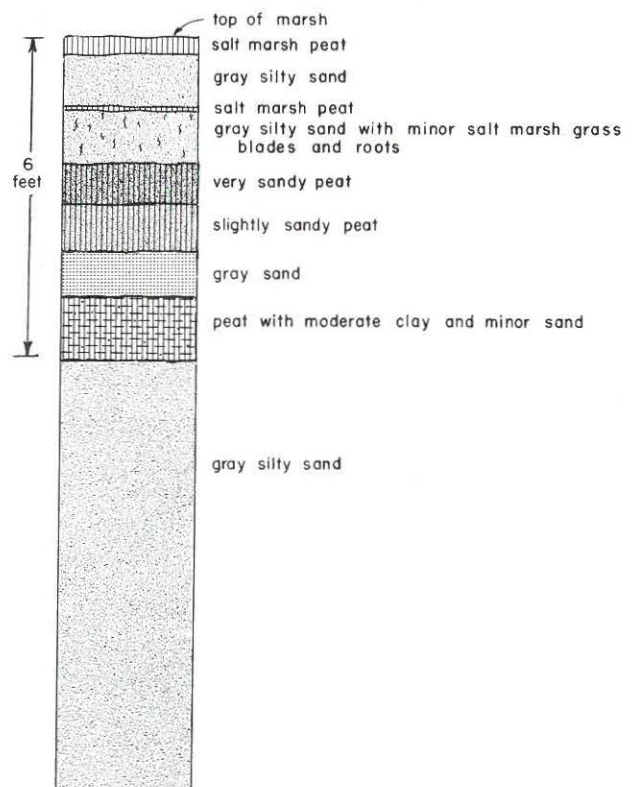


Fig. 11. Stratigraphic section of marsh deposits exposed in a 7 foot deep trench at locality 2, Figure 2. Trench was dug during dredging operations near the mouth of the Webhannet River. Discussion with the operator of the drag-bucket that was used to dig the trench indicated that the gray silty sand at the bottom of the 7 foot trench continued to a depth of at least 15 feet in a hole dug nearby but covered at the writer's visit.

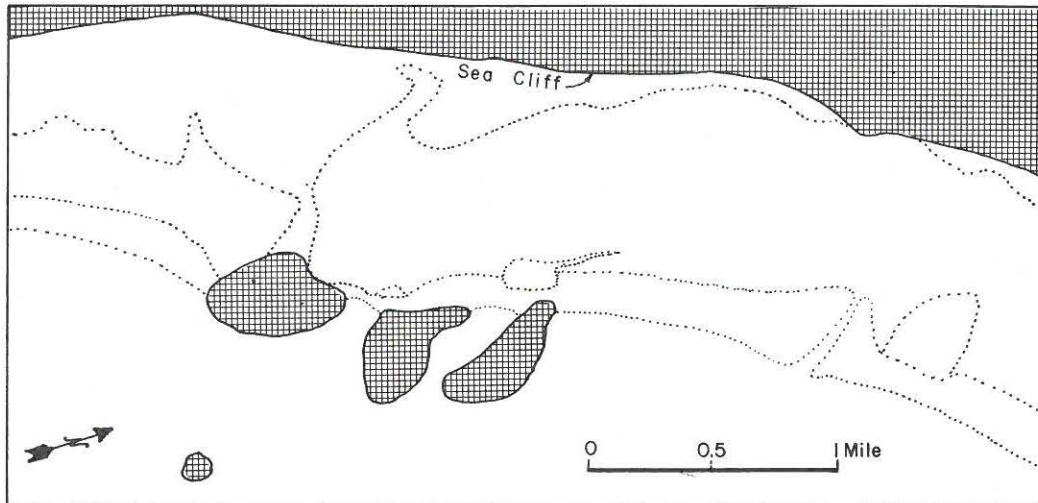


Fig. 12. Stage 1 in the development of Wells Beach: cutting of sea cliff now buried by glacial and post glacial deposits at a time prior to the last stage of glaciation. Cross-hatched pattern indicates terrain of unknown character, but with bedrock at or near the surface, on which the sea cliff was cut.

(2) a wave-cut scarp developed on glacial and post-glacial sediments on the mainland opposite the northern segment of the Beach;

(3) intertidal tree stumps dated as approximately 3000 years old, in the marsh behind the Beach, rooted in salt marsh peat and buried by approximately 2.5 feet of marsh peat;

(4) similar stumps exposed intermittently on the middle segment of Wells Beach, of essentially the same age, but four to five feet lower than those in the marsh;

(5) offshore islands, now represented by Moody Point, Points "A" and "B", and Drakes Island, which served to shield the mainland from strong effective wave attack and erosion and which served as anchors to which the beaches were attached;

(6) thin salt marsh deposits in the lagoon behind Wells Beach.

Stage 1 (Fig. 12). A bedrock scarp believed to be a sea cliff, inferred from the alignment of cascades over which streams from the mainland abruptly reach the level of the marsh, was formed prior to the last episode of glaciation. This scarp is buried by crudely stratified glacial gravel and post-glacial marine beds of the Presumpscot Formation. The sea level at the time of formation of this sea cliff may have been somewhat lower than at present because of lack of bedrock exposures of a complementary wave-cut bench in any of the tidal channels of the marsh. The ledges at Moody Point and Points

"A" and "B" probably were islands about a mile offshore.

Stage 2 (Fig. 13). After glaciation and deposition of the Presumpscot Formation, and during a temporary halt in the gradual reemergence of the land following the maximum marine invasion of the coastal area, a sea cliff was formed on the beds of the Presumpscot Formation and glacial drift. Bloom (1960, pp. 109 and 111) suggests that this strandline feature was formed between 7000-8000 and 3000 years ago when sea level was approximately 16-20 feet higher (the approximate elevation of the base of the scarp). This sea cliff was cut only on that part of the mainland opposite the northern segment of Wells Beach, clearly indicating that that segment had not yet been formed. The absence of a definitive scarp elsewhere along the strand line at this time is probably due to the protective influence of offshore islands now represented by Moody Point, Points "A" and "B", and Drakes Island, and the ledges offshore from these points. These islands would have acted as barriers to energetic open-ocean waves and thus would have shielded the mainland from effective wave attack. The reason for the absence of a sea cliff along the mainland opposite the Moody Beach segment of the barrier chain may be taken as evidence that this segment of the barrier chain had formed by this time, somewhat earlier than Wells Beach. During this stage much shoaling of the waters in the area of the present marsh must have taken place by depos-

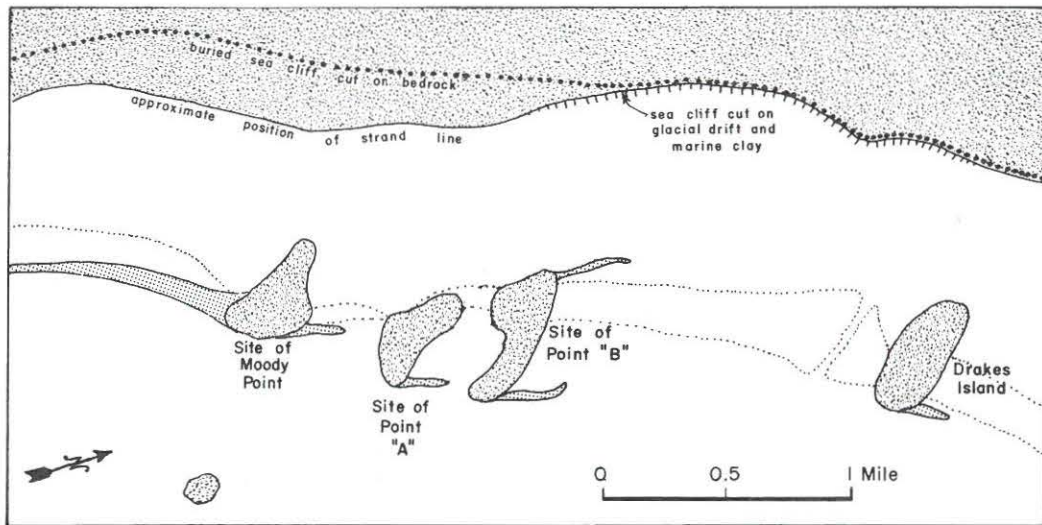


Fig. 13. Stage 2 in the development of Wells Beach: cutting of a sea cliff on glacial drift and Pre-sumpscot Formation deposits during a temporary halt in the fall of sea level following deglaciation. Sand spits began to form onto downdrift edges of offshore islands, later to become the various segments of Wells Beach and adjacent beaches. Patterns the same as in Figure 2.

ition of material eroded from the sea cliff and supplied by streams entering the ocean. The sand spit attached to the landward side of Point "B" was deposited at this time. Similar small spits, the forerunners of the different segments of Wells Beach, were probably formed northward from the seaward ends of the offshore islands.

Stage 3 (Fig. 14). At a time shortly before 3000 years ago, sea level relative to the land had fallen to its lowest level following glaciation,

probably as the result of completion of isotatic crustal rebound. Prior to this event, and probably shortly after the temporary halt in the emergence of the land that led to the formation of the sea cliff of Stage 2, the different segments of Wells Beach and adjacent beaches had been extended to form an essentially complete barrier chain isolating lagoons behind. This inference is based on the fact that the intertidal tree stumps found in the marsh are rooted in peat rather than in the sand and silt beneath the peat. Thus peat had already begun to form in a lagoon

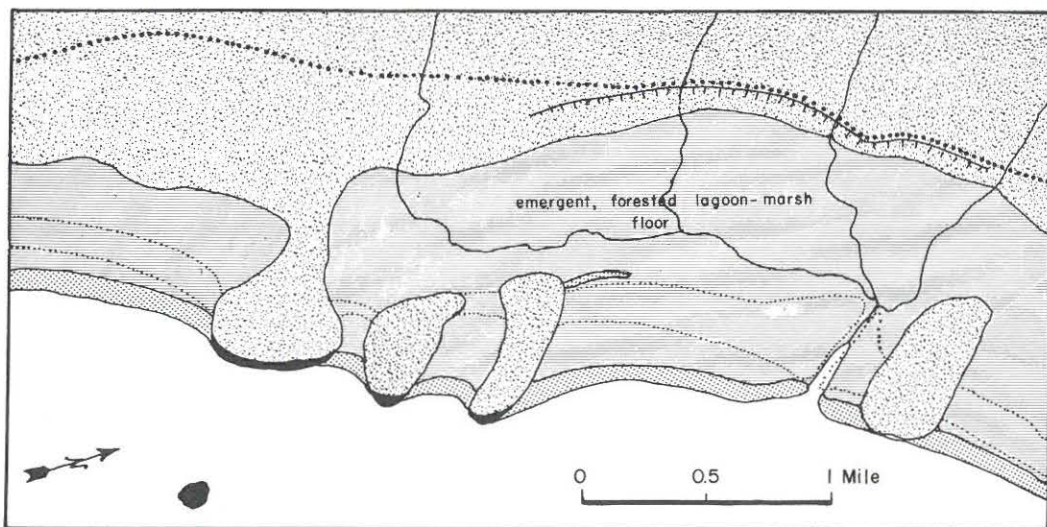


Fig. 14. Stage 3 in the development of Wells Beach: Development of spits into a barrier chain attached to offshore islands, followed by development of a pine forest on the emergent floor of the lagoon when the sea had fallen to its lowest level shortly before 3000 years ago.

behind a protective barrier beach prior to the low stand of sea level 3000 years ago.

Just prior to 3000 years ago the shallow lagoon behind the barrier beach chain was drained of salt water and upon this surface a pine forest became established. Sea level during this stage was at least three feet lower than at present because the stumps that are remnants of this pine forest are approximately 2.5 to 3 feet lower than the present level of the marsh which coincides with the present mean high tide level. If the marsh, formed between Stages 2 and 3, had been thoroughly laced by tidal channels as the present marsh is, sea level would have had to be perhaps an additional eight to ten feet (tidal range for this latitude) lower to prevent salt water incursion into the marsh sediments at high tide and during exceptional storms. Thus, prior to 3000 years ago, sea level probably stood 11 to 13 feet lower than at present.

Stage 4 (Present Configuration - Figure 2). At about 3000 years ago the pine trees growing on the salt marsh were killed, very likely as the result of a gradually rising sea level. The barrier beaches, attached to the ends of the islands represented by the ledges in front of Moody Point and Points "A" and "B" were approximately one-fourth to one-half mile further offshore than at present. The beaches were forced to migrate shoreward over former salt marsh deposits and the stumps of the drowned forest. The weight of the retreating beach ridge compressed the marsh deposits with the result that the stumps and peat exposed intermittently on the present foreshore of the middle segment of Wells Beach are approximately 3.5 feet lower than the same deposits in the marsh behind the beach today. The former offshore islands were stripped to a large extent of their cover of unconsolidated sediments by wave erosion, leaving ledges exposed. The ledges in front of Moody

Point have been extensively planed by wave erosion forming a wave-cut bench.

Stage 4 represents only a temporary configuration produced by a sea level that continues to rise, probably as the result of addition of glacial melt water to the ocean basins. If the process were to proceed without the human intervention that has already occurred, Wells Beach would continue to recede shoreward at the expense of the marsh behind, ultimately forming a fringing beach at the base of a wave-cut cliff on the mainland. However, the breakwater, extending the entire length of the beach, and around both Points "A" and "B", prevents an equilibrium balance from being maintained, by fixing the position of the backshore. At one place in the southern third of the middle segment, the breakwater has been extended to the upper edge of the foreshore and interferes with the equilibrium dynamics of the uprushing wave swash. The result has been to initiate erosion of sand from the beach. As eustatic rise of sea level continues, the presence of the breakwater will only serve to accelerate such removal of sand. It is thus unfortunate that the breakwater erected to protect the shore property and dwellings from the destructive action of storm waves is now causing the degradation of, and may ultimately destroy, the very feature for the enjoyment of which the dwellings were erected. Further elaboration on this very timely topic, though desirable, is beyond the scope of this article.

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