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MAINE GEOLOGICAL SURVEY

Robert G. Doyle, State Geologist

THE GEOLOGY OF THE CUTLER AND MOOSE RIVER QUADRANGLES WASHINGTON COUNTY, MAINE

By

OLCOTT GATES

Quadrangle Mapping Series No. 1

Department of Economic Development

Augusta, Maine

September, 1961

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INTRODUCTION

Acknowledgments

This report describes the geology of the Cutler and Moose River 7½ minute topographic quadrangles in Washington County, Maine. It is part of a program sponsored by the Geological Survey of Maine, Department of Economic Development, to study systematically the rocks of the state and to make this information readily available to those interested in the geology and mineral resources of Maine.

I am indebted to John R. Rand, former State Geologist, and Robert G. Doyle, present State Geologist, for their cooperation and assistance. Professor Carleton A. Chapman, University of Illinois, and Professor Philip H. Osberg, University of Maine, carefully read the manuscript and made many valuable suggestions. I have profited from discussions with my colleagues at The Johns Hopkins University; and Robert Morrison assisted me ably in the field during the summer of 1958.

I am very grateful to the fine people of the lovely town of Cutler for the unfailing kindness, generosity, and courtesy they extended to me and my family.

Previous Study

The rocks in the Cutler and Moose River quadrangles have been known in a general way for more than a century. In 1836 Dr. C. T. Jackson, a Boston physician and Maine's first state geologist, sailed in the Revenue Cutter *Crawford* from West Quoddy Head to Little River and Little Machias Bay. In his report (1837, p. 35) he described "enormous cliffs of greenstone trap, intersecting and overlying argillaceous and calciferous slate . . . " Hitchcock (1861, p. 185-189) noted that trap intruded slates in the Machiasport area and that this belt of trap extended through the Cutler area eastward into New Brunswick. Hitchcock's (1861, 1862) descriptions of the geologic complexity of the Eastport area, the finding of possible Devonian plants in the Perry formation by Rogers (1860), and the subsequent study of a much more diagnostic fauna by Dawson (1861) focussed geologic attention on the Passamaquoddy and Cobscook Bay areas for the next fifty years while the Cutler area was ignored. Canadian geologists Bailey and Matthew published a series of Canadian Geological Survey Reports of Progress (1872, 1876) in which rocks on the New Brunswick side of Passamaquoddy Bay were identified as Silurian volcanics, post-Silurian granites, and Devonian sedimentary rocks. Shaler (1886) made a two months reconnaissance of the Cobscook Bay area and found abundant Silurian fossils and volcanic rocks, intrusive gabbro and diabase, and emphasized the marked unconformity beneath the Perry formation of Devonian age. Smith and White (1905) described the geology of this formation in a monograph and showed that the rumors of coal in it were unfounded.

In 1914, Bastin and Williams published the Eastport Folio of the U. S. Geological Survey's Geologic Atlas of the United States. The south boundary of their map is the north boundary of the map of this report. They mapped a large open east-plunging anticline of Silurian shales and volcanic rocks cut off on the southeast by a shear zone, the Lubec shear zone, southeast of which are large masses of diabase intrusive into the Quoddy shale, the oldest Silurian rocks in the Eastport quadrangle. The diabase that makes up so much of the Cutler and Moose River quadrangles is an extension of this diabase. Since 1914, there has been no detailed study published of any bedrock geology from West Quoddy Head to Machias Bay.

SURFICIAL GEOLOGY

Topography

The Cutler and Moose River quadrangles, a land area of about 32 square miles, lie along the easternmost coastal section of Maine. The southwest trending coastline parallels a steep submarine scarp which extends to a depth of about 200 feet where it merges with the floor of the Bay of Fundy. Johnson (1925, p. 286-294) and Koons (1941) interpreted this submarine slope as the fault line scarp of a large fault (the Fundian fault) marking the western boundary of the Triassic fault basin they believed underlies the Bay of Fundy. Shepard (1930, 1942), on the other hand, considered the straight steep slope to be primarily the result of erosion by a great glacier that flowed down Grand Manan Channel and once occupied a large part of the Bay of Fundy.

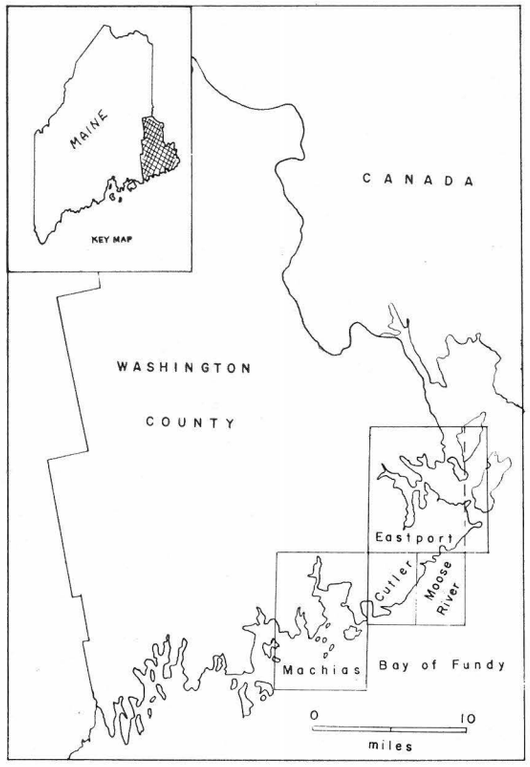

Fig. 1. Location of the Cutler, Moose River, and adjacent quadrangles in Washington County and of Washington County in Maine.

The coastline east of Little Machias Bay is quite regular, the only indentations being Moose River and Little River. It is generally rocky and bold with few offshore reefs or shoals. Some of the cliffs are as much as 160 feet high; gravel beaches are restricted to a few shallow coves; and the only tidal flats are at the heads of Moose River, Little River, and Little Machias Bay.

The present configuration of the coastline results primarily from pre-glacial and glacial erosion. Post-glacial marine erosion has been confined to local plucking and quarrying of jointed rocks, and to the cutting of cliffs in Pleistocene sediments in Little Machias Bay and in House Cove, Long Point Cove, and Holmes Cove. The steepness of some of the bedrock cliffs suggests that they were cut by wave erosion, but well preserved glacial striations on many of them indicate that marine abrasion and quarrying have been slight during post-glacial time. The absence of submarine or tidal zone wave-cut bedrock platforms also points to very little marine erosion since the departure of the glaciers.

The Cutler and Moose River quadrangles are in the Coastal Lowlands physiographic province according to Toppan's (1935) classification. The topography is generally subdued and has a maximum elevation of about 250 feet. Hills and knobs of bedrock are separated by broad shallow valleys, numerous swampy flats, and scattered heaths. Most hills are of resistant diabase and gabbro, although hills of sedimentary and volcanic rocks are also common. Drainage is poorly integrated and streams are meandering. The topographic map, however, gives a false impression of the small-scale relief. Vertical cliffs, narrow small steep-sided gullies and valleys, and steep bedrock knobs which rise 50 to 100 feet above the surrounding land surface are either completely omitted or shown greatly subdued on the map.

Prior to advance of the Wisconsin glacier, the area had already been reduced to low relief by a long period of erosion probably extending back through the pre-Wisconsin glaciations and much of the Tertiary. The area was presumably a rolling upland standing several hundred feet above the present floors of Passamaquoddy Bay and the Bay of Fundy. Sonic sounding to bedrock beneath unconsolidated sediments in Passamaquoddy Bay has revealed the course of a river which once flowed through Western Passage between Deer Island and Eastport and thence northeast between Deer and Campobello Islands to the present floor of the Bay of Fundy (Upson, 1954, p. 291). The main

drainage in the Cutler area may have followed East Stream valley northeast to join this now submerged river course.

Glacial erosion has rounded and striated some of the bedrock knobs, locally plucked their southeast sides to form steep cliffs, carved small roches moutoneés, and scoured some of the shear zones to give a distinct grain to the topography trending northeast parallel to the regional shearing. Moraine and gravels dumped by the glacier have filled valleys, formed ridges, and modified the drainage.

The topography can conveniently be divided into three main areas parallel to the southwest-trending coastline. The coastal area, extending about 3 miles inland, contains the most numerous hills of bedrock and the best bedrock exposures. The tops of many hills and knobs are free of unconsolidated surficial deposits; and scattered outcrops of bedrock occur in some of the lowland areas. The central valley area is bounded on the southeast approximately by a line running from Ackley Pond along French Ridge to Stone Mountain and thence through Warren Meadows. Most of this central area lies in the valley formed by East Stream and Eastern Marsh Brook. This broad valley contains only scattered knobs of bedrock along its eastern margin. It is filled by glacial and post-glacial rocks so completely that mapping of the bedrock is prohibited. The third or inland area of topography is in the northwest corner of the Cutler quadrangle where low ridges of bedrock trend northeast and are overlapped by the glacial gravels and clays through which Lively Brook meanders.

Glacial and Post-glacial Deposits

Pleistocene glaciers, of which only the Wisconsin has left a record, covered Maine and may have extended as far southeast as Georges Shoals (Shepard and others, 1934). As the ice melted back or shortly afterwards, southeastern Maine was flooded by the sea in which fossiliferous marine clay was deposited up the Kennebec River valley as far as Bingham and up the Penobscot as far as Lincoln (Goldthwait, 1949, p. 65); and deltas of glacial outwash formed where glacial streams entered the sea almost 100 miles inland from the present coast (Leavitt and Perkins, II, 1934, p. 199). Since withdrawal of the sea, the marine clay and deltas inland have been elevated by glacial rebound about 450 feet and along the coast from 150 to 250 according to various estimates (Antevs, 1928, p. 322; Leavitt and Perkins, II, 1934, p. 199; Goldthwait, 1951, p. 27; Flint, 1957, p. 251).

Leavitt and Perkins (II, 1934, p. 191-193) and Trefethen and coworkers (1947, p. 15) present evidence that the main withdrawal of the ice and advance of the sea were followed by a weak readvance of ice and second marine invasion.

Retreat of the glaciers and inundation of the sea left glacial till, moraine, outwash sand and gravel, and clay strewn over more than 90 per cent of the Cutler and Moose River quadrangles. In the coastal topographic area, the surficial sediments above the 100 to 120 foot contours consist largely of a mixture of boulders, cobbles, sand, and clay, varying considerably in the ratio of one to another and in the degree of stratification from place to place. Locally there are patches of distinctly stratified and sorted sand and gravels, as well as flats, now generally swampy, underlain by clay and strewn with boulders. Outcrops of bedrock are common on the hills and knobs and on steep slopes. Below the 100 foot contour outcrops of bedrock, except along the shoreline, are more scattered and discontinuous, boulders are less common, and bedded clays form flat valley floors into which meandering streams have been incised to a minor degree.

One of the most conspicuous topographic features of glacial origin in the coastal area is Pond Ridge, a glacial moraine, which extends from Sprague Neck in the Machias quadrangle continuously northeast several miles into the Cutler quadrangle and discontinuously as far northeast as Bog Brook.

Leavitt and Perkins (II, 1934, p. 44-46, 160) compared Pond Ridge and its continuation southwest to Sprague Neck with the Newington moraine bank of southwestern Maine described by Katz and Keith (1917). The latter used the term moraine bank for ridges of sand and gravel formed along an ice front standing in the sea. The steep straight northwest slope of Pond Ridge according to this interpretation is an ice-contact slope.

In the central valley, brown and black bedded clays, probably marine, are exposed in the banks of the streams; and bouldery gravels and moraine form the east-trending low ridges that traverse the valley. Coarse to medium-grained outwash gravels of well-rounded pebbles and cobbles flank the southeast side of Great Ridge.

Bastin and Williams (1914, p. 15) and Upson (1954, p. 290) conclude that the evidence is substantial in the Eastport area for a stand of the sea at about the 90 to 100 foot contour and that there is less

convincing evidence of a submergence to at least 200 feet. In the Cutler and Moose River quadrangles, erosion, redistribution, and modification by waves and currents of a mantle of glacial drift as sea level gradually lowered to about the present 100 foot contour is the simplest general explanation for the widespread distribution of bedrock outcrops and the mixture of various types of unconsolidated deposits above the 100 foot contour, especially in the coastal area.

Since withdrawal of the sea, changes in the topography and the surficial sediments in the Cutler and Moose River quadrangles have been slight. Surf has cut back unconsolidated sediments to form steep sea cliffs; an extensive boulder tidal flat has formed at the head of Little Machias Bay; some streams have incised the flat valley floors 5 or 6 feet and locally have cut through hummocks and ridges of glacial material; and some abandoned tidal flats and shallow lakes in kettle holes have been filled by vegetation to form swamps and heaths.

BEDROCK GEOLOGY

Introduction

The Pleistocene and Recent surficial unconsolidated sediments rest with profound unconformity on deformed and somewhat metamorphose sedimentary and volcanic rocks of Silurian age, keratophyre porphyry, and immense quantities of intrusive diabase. The distribution of these bedrocks was mapped by pace and compass traversing through as many topographic check points as could be found. Reliability of the mapping is excellent along the coastline and at altitudes above 200 feet where exposures are practically continuous. Inland, at lower altitudes, and in heavily wooded areas contacts are located within 200 feet at most places. West of a line from Ackley Pond along French Ridge and through Warren Meadows, the distribution of the map units is made according to the most likely of several alternative ways to connect up widely scattered and relatively few outcrops; and in the central valley no mapping of bedrock is possible because of the complete lack of outcrops.

VOLCANIC AND SEDIMENTARY ROCKS

The volcanic and sedimentary bedrocks of the Cutler and Moose River quadrangles have been divided into three formations, all of Silurian age. The Edmunds and Pembroke formations northwest of the central valley are extensions of formations defined by Bastin and Williams in the adjacent Eastport quadrangle. Southeast of the central valley, blocks of volcanic and sedimentary rocks separated by a multitude of diabase intrusions are defined in this report as the Little River formation. Glacial sediments in the central valley completely obscure the contact between the Little River and Pembroke formations. Presumably the Lubec shear zone mapped by Bastin and Williams continues from the Eastport quadrangle into the Cutler along the central valley and separates the two formations.

The legend on the geologic map implies that the Little River formation is older than the Pembroke and Edmunds. The only evidence for this is that the Little River formation is on the same side of the Lubec shear zone as the Quoddy Shale, which was considered by Bastin and Williams (1914, p. 3) to be the oldest formation in the Eastport quadrangle.

Silurian System

Little River Formation

The diverse group of volcanic and sedimentary rocks named here the Little River formation does not constitute a formation according to the established practices of stratigraphy. No top or bottom can be defined; the thickness is unknown; no stratigraphic succession applicable to the formation as a whole can be recognized; and no particular exposure or groups of exposures is sufficiently representative of all the rocks included in the formation to permit designation of a type section. The formation essentially consists of blocks of dominantly volcanic rocks, many of which are marine, surrounded and isolated from each other by interlacing dikes, sills, and plugs of diabase. Extensive cataclastic shearing, which obliterates bedding, and a dearth of outcrops inland further obscure the stratigraphy of the formation.

Age and correlation

The Little River formation is on the extension to the southwest of the rocks named the Quoddy Shale by Bastin and Williams (1914) in the Eastport quadrangle. The map of the Eastport quadrangle shows Quoddy Shale exposed at West Quoddy Head and along the west shore of Quoddy Roads. Further southwest the Quoddy Shale forms isolated blocks in a sea of diabase up to the south border of the Eastport quadrangle adjoining the Moose River and Cutler quadrangles. However, except for a thin sequence of argillite along the north shore of Moose River, most of the Little River formation bears little lithologic resemblance to the rocks in the type section and other good exposures of Quoddy Shale.

A few fossils found in a coarse limestone and tuff breccia near Hues and near Great Head (marked by F on the geologic map) were identified by A. J. Boucot as *Stricklandia lens ultima, Eospirifer, Merista, Isorthis, Plectodonta, Resserella, Plectatrypa?, Euomphalus,* and *Calymene.* Dr. Boucot writes of these collections, "Collections 1 and 2 are similar to what Williams reported from the Quoddy formation although they could be slightly older or slightly younger." "The *Stricklandia* is of stratigraphic significance as this particular variety is restricted to zone C-5 of the Upper Llandovery."

The Little River formation therefore appears to be the same age or nearly so as the Quoddy Shale and may represent a rather abrupt

facies change from shales and mudstones to volcanics. However, it differs so much in lithology from the Quoddy Shale that until we know more about the stratigraphy and structure of southeastern Maine, I prefer to assign a new formation name rather than to extend the Quoddy Shale into the Cutler quadrangle.

The stratigraphic relations of the Little River formation to the Edmunds and Pembroke formations are unknown because the geologic evidence is buried beneath the glacial deposits of the central valley. Some of the coarse tuffaceous breccias in the Little River formation exposed along the south shore of Little Machias peninsula and along the north shore of Almore Cove resemble rather closely coarse breccias of the Edmunds and Dennys formations in the Cobscook Bay area of the Eastport quadrangle. However, correlation of the Little River formation with either of these formations on this lithologic basis raises difficult structural and stratigraphic problems in tying the geology of the Cutler quadrangle to that of the Eastport quadrangle.

Lithology

The Little River formation for purposes of description can be divided into three principal rock types: (1) basaltic lavas and pillow lavas, agglomerate, and tuff; (2) coarse tuffaceous breccias made up of many different lithologies—mainly basaltic; and (3) bedded tuff and argillite with minor amounts of conglomerate, sandstone, shale, and limestone. On the geologic map, the basaltic volcanics and the generally basaltic tuffaceous breccia have been differentiated by overprint pattern without contacts. As a broad generality, the basaltic volcanics are most plentiful to the north and northwest; the breccias dominate the south and southeast; and the bedded sediments form lenses in both the other rock types, but more commonly in the breccia. However, attempts to map any of these lithologies as stratigraphic members of the Little River formation from block to block across the intervening intrusions of diabase failed to reveal any continuity or stratigraphic order.

Basaltic rocks. Best exposures of basaltic lava flows, pillow lavas, and agglomerate are at Eastern Head. Other good exposures, are on French Ridge and northeast of Cocoa Mountain. At Eastern Head, vesicular basalt flows rest on siliceous black shales and have pipe amygdules along their bases and spherical amygdules along their tops (Fig. 2). In other areas the basaltic flows display zones of calcite or chlorite

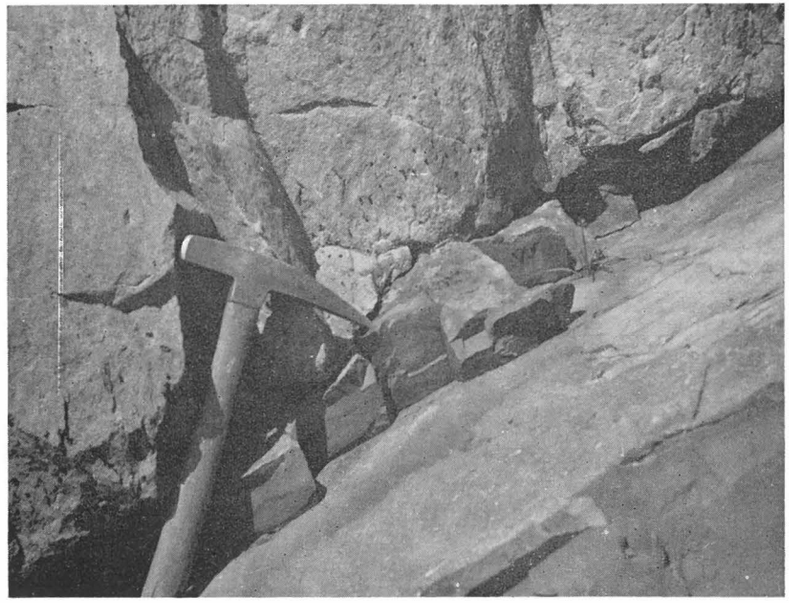

Fig. 2. Base of lava flow resting on silty shale. Note the pipe amygdules. Eastern Head.

amygdules which commonly are flattened parallel to the region-wide shear, lenses of basaltic rubble which are probably flow tops, polygonal jointing, and numerous irregular clots and veins of calcite. The flows at Eastern Head are not more than 20 feet thick. Away from the shoreline exposures are generally not sufficient to permit tracing of individual flows and hence to estimate typical thicknesses.

Locally, the amygdular basalt flows show excellent pillows, particularly at Eastern Head. There the pillows are irregularly elliptical with a long diameter of about 3 to 4 feet and a short diameter of 2 to 3 feet. A thin glassy selvage rims individual pillows, and amygdules tend to form concentric bands within the pillow. Fine black siliceous material — perhaps once fine ash — and coarse basaltic rubble occupy the triangular spaces between pillows or make up irregular lenses that meander among the pillows.

At some places, the lavas appear to have been injected into unconsolidated muds and fine tuffs. Rounded, bulbous fingers of pillow lavas or diabase intrude argillite or tuff; and laminations in the fine-grained sediments are distorted and tend to follow the outlines of the

intrusions (Figs 3 and 4). On Eastern Head finely banded siliceous argillite drapes around some of the pillows in underlying lavas and projects downward into spaces between pillows.

The lavas and pillow lavas in outcrop weather to a dark brown and are dark gray to greenish black on a freshly broken surface. Chlorite and calcite amygdules lie in a matrix which under the hand-lens appears to be largely fine-grained chlorite. During mapping, where pillows, rubble, or contacts were absent, if the outcrop showed abundant amygdules, the rock was classified as a basaltic lava. However, some of the intrusive diabase sills have amygdular marginal zones; and undoubtedly some of these have been mistakenly mapped as lava.

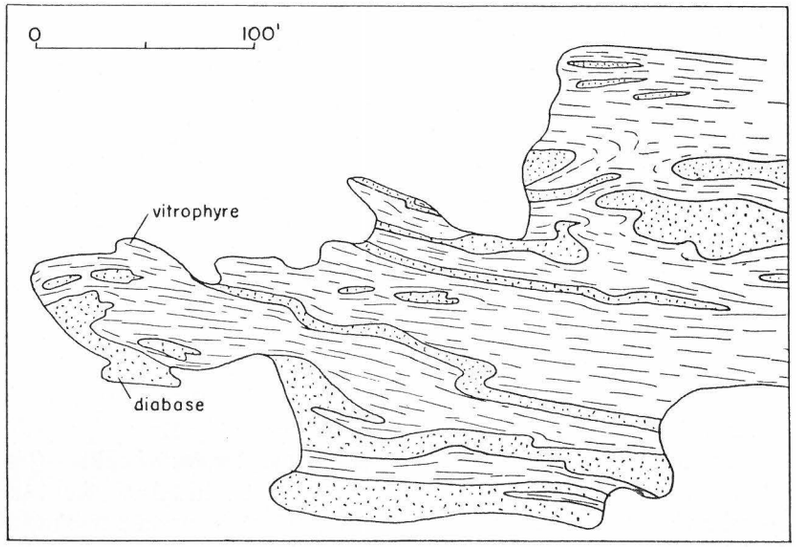

Fig. 3. Thin diabase dikes which appear to have intruded unconsolidated tuffaceous and siliceous muds, now tuffaceous vitrophyre. Map view along shoreline south of Moose River.

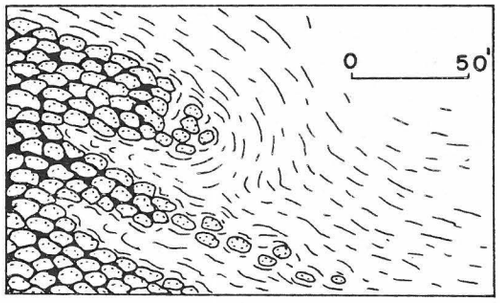

Fig. 4. Pillow lavas which intruded unconsolidated siliceous mud, now argillite. Note the detached pillows. Map view at Eastern Head.

The microscope shows that the lavas now consist of scattered small laths of dusty unzoned albite, small epidote grains, and magnetite or leucoxene granules in a matrix largely of chlorite. Some lavas, particularly the pillow lavas, contain scattered patches of carbonate replacing the groundmass or plagioclase. Although these rocks have been altered to low grade greenstones, remnant intersertal, ophitic, and intergranular textures clearly show that these rocks were once basalt or diabase.

Lenses and thick piles of basaltic agglomerate occur in many places associated with the lavas and pillow lavas. The largest mass, at least 300 feet thick, at Eastern Head abruptly interfingers with pillow lavas and bedded submarine tuffs and is magnificently exposed along steep sea cliffs. Eastern Head may be near the location of one of the original volcanic vents.

The agglomerate consists of spindle or football-shaped bombs of amygdular basalt in a fine-grained basaltic tuff matrix (Fig. 5).

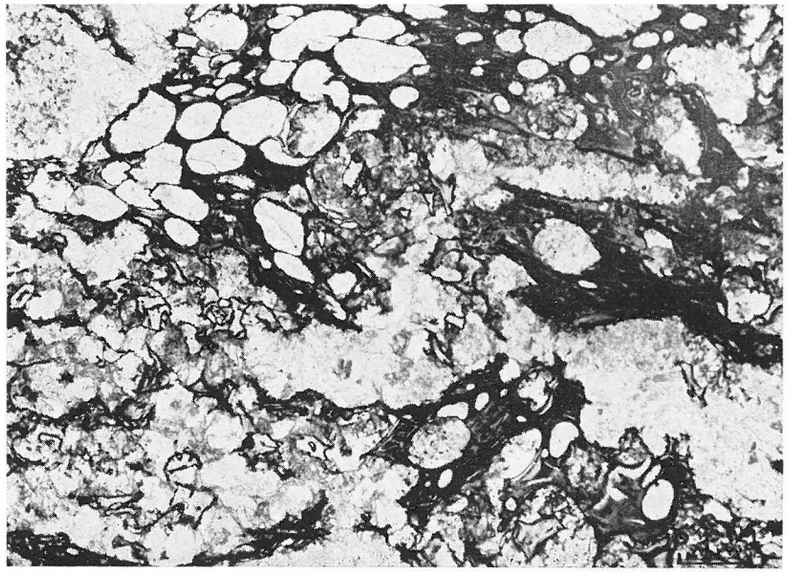

Fig. 5. Photomicrograph of basaltic agglomerate matrix. Vesicles are filled with calcite. The light irregular areas are calcite replacing lapilli. Plain light. North shore of Almore Cove.

Bombs range from a few inches to 3 feet in longest dimension; many have a ropey fine-grained surface; and all carry abundant amygdules. The microscope reveals that the bombs are made up of saussuritized plagioclase laths in a very amygdular groundmass largely of chlorite and carbonate. The bombs are in a matrix of basaltic shards and pumice lapilli whose vesicles are filled by chlorite, carbonate, and quartz, or any combination of these.

Commonly associated with the basaltic lavas and agglomerates are outcrops of fine-grained chloritic tuff, generally unbedded, but in places containing thin streaks or stout pods of bombs or rock fragments. Otter Point and Dennison Point contain excellent exposures. When sheared, these rocks are indistinguishable under a hand-lens from sheared basalt flows. In thin-section, it can be seen that the fine-grained varieties consist of shards, fragmentary saussuritized plagioclase crystals, fragments of basaltic chloritized glass, and pumice lapilli full of chlorite, carbonate, or albite amygdules. The coarse varieties in addition have fragments of amygdular basalt and diabase, tuff, and quartz porphyry. These rocks are clearly basaltic tuffs some of which incorporated other rocks, perhaps torn from the walls of vents.

Tuffaceous Breccia. The term tuffaceous breccia is here applied to rocks in the Little River formation that are unbedded and consist of rock fragments in a matrix of smaller rock fragments and tuffaceous material. This breccia occurs, as do the other rocks of the Little River formation, in blocks separated from neighboring ones by diabase intrusions; and different blocks of breccia may differ considerably in lithology.

On the geologic map tuffaceous breccia is differentiated from the basaltic volcanics by use of a different overprint pattern but contacts between them are deliberately not drawn. This usage reflects the relations between the basaltic tuffs and lavas and the breccia as seen in the field. The breccia and other volcanics are intimately mixed, and the overprint pattern designating breccia merely indicates areas where the breccia is the predominant rock. In many of these areas, the breccia is intruded by narrow irregular dikes of diabase or keratophyre; many small diabase intrusions pinch out in a mass of breccia or carry numerous inclusions of breccia and tuff; diabase, keratophyre, tuff, or bedded rocks may be cut or intruded by tongues and dikes of breccia; and masses of coarse breccia may cut across masses of fine breccia or vice-versa. Figure 6 illustrates many of these relations. In addition, in the areas designated as basaltic volcanics undif-

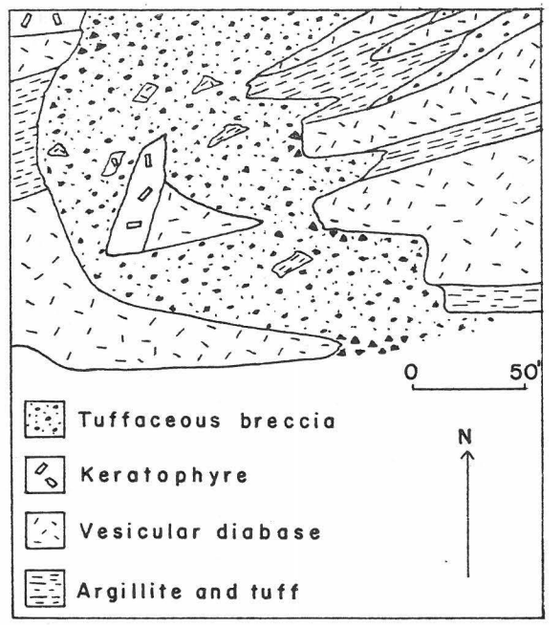

Fig. 6. Tuffaceous breccia intrusive into vesicular diabase, keratophyre, and argillite and tuff. Map view. West of Black Point Cove near Schooner Brook.

ferentiated, there are many lenses, pods, and irregular masses of breccia.

The bare hills in the vicinity of Norse Pond expose craggy outcrops of sheared tuffaceous breccia so intimately mixed with keratophyre porphyry that separation of the two on the map scale is impracticable. In some places breccia apparently intrudes keratophyre porphyry and in other places porphyry intrudes breccia (Fig. 7). Flattened and elongated blebs of prophyry, averaging about 3 to 4 inches long, are scattered through a finer-grained, sheared, chloritic matrix. The main variation in the breccia is in the size of its fragments. Coarse breccia carrying fragments up to a foot long forms lenses and pods commonly 10 to 20 feet long and 5 to 10 feet thick surrounded by fine-grained breccia. Lenses of fine breccia in coarse also occur.

In thin section, it can be seen that the matrix of the breccia consists of fragments of diabase, basalt porphyry, and keratophyre porphyry which are much sheared and flattened parallel to the shear planes. Irregular vugs and vesicles in the basalt fragments are filled

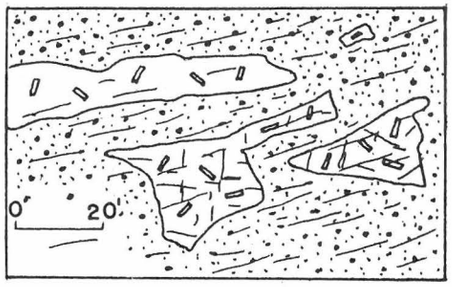

Fig. 7. Angular blocks of keratophyre porphyry in tuffaceous breccia which in turn has been intruded by a small dike of keratophyre porphyry. Map view. Norse Pond.

by quartz, epidote, albite, pale green chlorite, or calcite, or by any combination of these.

Sheared tuffaceous breccia of slightly different composition and lithology from that at Norse Pond is very well exposed on the narrow tongue-like ridge about 1½ miles southwest of Stone Hill. The fragments are larger than at Norse Pond, as much as 3 feet long, and consist of amygdular basalt and diabase, spindle-shaped basaltic bombs, and bedded fine-grained silty tuff. The matrix is much like that at Norse Pond but also contains some basaltic pumice and devitrified shards.

Sheared tuffaceous breccia exposed in a quarry on route 191 at the head of Little River consists of flattened fragments of basalt and diabase up to 4 inches long in a matrix of basaltic pumice lapilli, fragmentary altered plagioclase crystals, and chlorite. This breccia, in some places with fragments a foot long, is typical in the area extending from this quarry to Schooner Brook.

Excellent exposures of tuffaceous breccia occur in Little Machias peninsula, particularly along the shore west of Deer Island. Here blocks of every lithology recognized among the rocks of the Little River formation are enclosed in a matrix of basaltic pumice lapilli, tuff, and a wide variety of small rock fragments.

Blocks in the breccia are of the following rock types; bedded tuff; well-bedded and laminated argillite and fine tuff; lithographic limestone; bedded pebble conglomerate; amygdular basalt and diabase; and basaltic bomb agglomerate. Many of these blocks are as much as 20 feet long and 10 feet wide, although most range in size from 2 to 10 feet. Well-bedded blocks tend to be tabular, the long dimensions

parallel to bedding; and the unbedded or coarsely bedded blocks such as those of conglomerate or agglomerate generally form irregular lenses and pods (Fig. 8). Some of the blocks of laminated argillite or

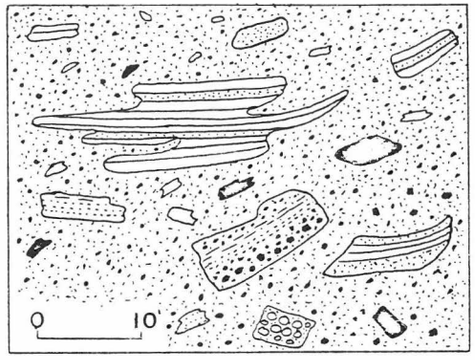

Fig. 8. Typical tuffaceous breccia carrying blocks of bedded argillite and siltstone, bedded tuff, limestone pebble conglomerate and basalt. View of a small cliff. Little Machias peninsula.

fine-grained tuff appear to have been somewhat plastic muds when incorporated in the breccia matrix. They are intruded by thin dikelets of matrix or they extend long fingers, whispy filaments, or ameboid-like projections into the surrounding matrix. Many of the tuffaceous blocks differ from the breccia matrix only in being bedded. The range of grain size and composition of fragments are the same in both block and matrix.

The matrix itself is almost as heterogeneous as the wide variety of rock fragments it encloses. In some places for a thickness of 10 or 15 feet and a length of perhaps 40 or 50 feet it faintly shows crude bedding which fades away into more typical unbedded breccia. In the unbedded breccia there are wide variations in grain size. In some exposures it is so fine-grained that a hand-lens is needed to see its fragmentary nature. At other places the included rock fragments and bombs are as much as 6 inches long. These variations in grain-size are distributed in irregular blocks and masses 20 to 50 feet long commonly approximately equidimensional, but with highly irregular, scalloped borders, some of which are sharp and some gradational. Coarse-grained masses are surrounded by fine-grained breccia and fine-grained masses by coarse-grained breccia (Fig. 9).

A thin-section of the matrix of tuffaceous breccia from the south shore of Little Machias peninsula shows irregularly shaped fragments

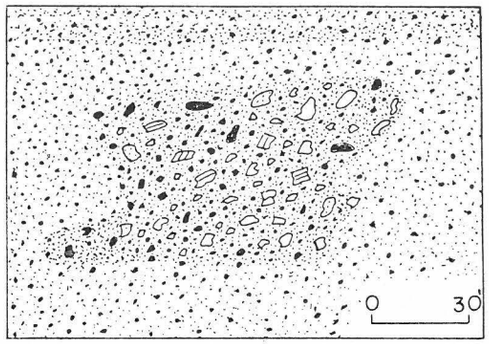

Fig. 9. An irregular mass of coarse tuffaceous breccia in fine tuffaceous breccia. View of a small cliff. South shore of Little Machias peninsula.

of basalt, basalt porphyry, amygdular basaltic pumice, keratophyre porphyry, saussuritized plagioclase crystals, argillite, silty limestone, and fossils cemented together by calcite, chlorite, clay minerals, and phlogopite. Calcite and phlogopite also fill the amygdules in the basaltic fragments.

In the interior of Little Machias peninsula, notably near triangulation stations Dennison and Davis, blocks of well-bedded tuff and limestone conglomerate several acres in area appear to be completely engulfed in tuffaceous breccia and tilted to various dips.

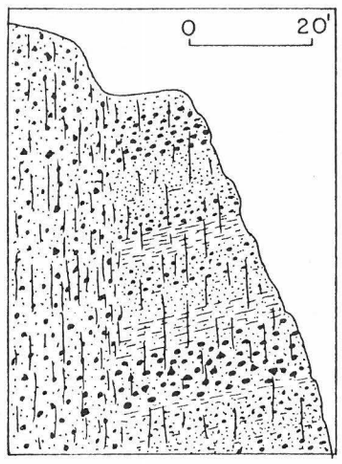

Fig. 10. Block of sheared bedded tuff cut off by unbedded tuffaceous breccia. Note the absence of a sharp contact. View of a small cliff. South shore of Little Machias peninsula.

Fig. 11. Typical tuffaceous breccia, considerably sheared. Fragments are siltstone, tuff, and basaltic bombs. Almore Cove.

Bedded rocks. Bedded sedimentary rocks are exposed along the north shore of Moose River, north of Holmes Cove, along the north shore of Almore Cove, and on either side of Great Head. Shearing, weathering, and absence of extensive outcrops have limited their identification inland, but scattered outcrops can be found in many places, particularly on French Ridge, the ridge west of Ackley Pond, and the interior of Little Machias peninsula. The bedded rocks occur as lenses in lavas, basaltic tuffs, and agglomerate; as large blocks surrounded by unbedded tuffaceous breccia; and as irregular shaped bodies isolated by intrusive diabase. They display a wide range of lithology from place to place.

The north shore of Moose River is parallel to the strike of a section of black argillite, siliceous sandstone and siltstone, and well-bedded tuff, all with a south dip. These rocks are the only ones of the Little River formation that resemble those at the type section of the Quoddy Shale in the Eastport quadrangle.

The dominant rock is a black to dark gray, banded argillite cut by slaty cleavage at an angle to bedding. There are also siliceous siltstones and sandstones made up of angular to subrounded grains of quartz and feldspar in a matrix of chlorite and calcite. Individual beds

range from an inch to a foot thick, and some have very small-scale cross-bedding. Well-bedded basaltic tuffs consisting of shards and lapilli make up beds ranging from thin laminations to 10 feet in thickness. Some of these beds are graded from coarse lapilli tuff at the bottom to a fine siliceous argillite at the top. Others have cross-bedding or scour between successive beds.

Fig. 12. Photomicrograph of the matrix of coarse tuffaceous breccia. Almore Cove.

Coarse massive tuffaceous conglomerate on the north shore of Almore Cove has yielded one of the two collections of Silurian fossils made in the Cutler and Moose River quadrangles. The conglomerate is made up of subangular to round cobbles and a few boulders scattered through a matrix of small pebbles, chlorite, and calcite. The boulders, cobbles, and pebbles are of tuff, aphanitic feldspar porphyry, sandy, bedded limestone, and laminated argillite. Beaks of brachiopods were collected both from the matrix and the limestone

cobbles. A penetrative shear through the conglomerate has flattened and stretched the pebbles and undoubtedly helped round their corners. This coarse conglomerate forms a bed about 20 feet thick in a sequence of laminated argillite, well-bedded basaltic and lapilli tuff, basaltic tuff breccia, and basaltic lava.

Coarse tuffaceous conglomerate, also fossiliferous, is spectacularly exposed on the south shore of Great Head in a continuous section about 80 feet thick. The conglomerate strikes east and dips south. The base of the section is a fine-grained basaltic tuff. Above this is about 10 feet of laminated dense limestone followed by 30 feet of well-bedded basaltic lapilli tuff containing pebbles up to 1 inch long of limestone, argillite, and chert. The tuff coarsens upwards, gradually loses its crude bedding, and grades into a coarse tuffaceous breccia full of angular to round fragments of blue dense limestone, bedded pebbly limestone, siliceous banded argillite, feldspar porphyry, porphyritic diabase, and bedded basaltic tuff. This fragmental material is firmly cemented by a bluish gray limey and chloritic matrix largely of tuff, lapilli, and small rock fragments. Many of the argillite and limestone blocks are tabular with sharp corners, being 3 or 4 inches thick and 2 or 3 feet long. This non-bedded coarse breccia includes several lenses up to 5 feet thick of banded argillite and limestone showing some graded bedding and many penecontemporaneous, irregular, disharmonic folds. Towards the top of the section the largest fragments gradually decrease to pebble size and a faint crude bedding becomes apparent. Brachiopods were found in both the limey cobbles and boulders and in the shardy and pebbly matrix.

This coarse tuffaceous conglomerate and breccia is probably genetically associated with the tuffaceous breccia of Little Machias peninsula. Discrete blocks of tuffaceous conglomerate occur in the tuffaceous breccia, and the matrix of the coarse conglomerate is indistinguishable from that of the tuffaceous breccia.

On the north shore of Great Head a sequence about 50 feet thick of tuff, argillite, and silty limestone is well exposed. Individual beds range from 3 inches to 10 feet thick. Many of the tuff beds are very well graded from a coarse lapilli tuff at the base to a fine-grained shardy siltstone at the top. Large discrete blocks, some several acres in area, of similar tuffaceous sediments were found surrounded by tuffaceous breccia in the interior of Little Machias peninsula.

Origin

The rocks of the Little River formation clearly record a period of volcanism, much of it explosive, in which both lavas and pyroclastic debris accumulated on the sea floor. Between eruptions, fine ash and fine limey muds formed fine-grained laminated and well-bedded sediments which were soon buried under lava flows or a rain of pumice and bombs.

The significant evidence of origin provided by the regional distribution and stratigraphy of the tuffaceous breccia beyond the borders of the Cutler quadrangle remains to be learned. Any preliminary hypothesis must be derived solely from its local geologic relations and lithology. Clearly its origin must be found within the context of an explosive volcanic environment, largely submarine. Some of the breccia is made up solely of volcanic debris and all of it contains some volcanic materials. Those of wholly volcanic material, such as at Norse Pond or at the quarry near the head of Little River, may be vent breccias or thick sections of torrentially deposited unbedded tuffs. Much of it, however, resulted from disruption of lava and agglomerate and bedded sequences of limestone, argillite, shale, and lapilli tuff like those found still intact in other parts of the Little River formation; and some of these parent rocks may have been only partially consolidated when disrupted. Both the breccia and the blocks of bedded rocks it encloses contain the same kinds of mineral grains and the same size ranges, shapes, and types of rock fragments; and fossils found in the matrix of the coarse unbedded tuff conglomerate are the same species as those found in the limestone boulders of the conglomerate. The blocks and boulders are interpreted to be those parts of an original semi-consolidated sequence of rocks which were lithified enough to withstand the deformation that disaggregated the less consolidated surrounding rocks. This disruption must have been both penetrative and herculean for it destroyed the parent rocks on a grain to grain scale, largely homogenized previously heterogeneous bedded rocks, and tilted blocks acres in size to various angles.

The probability that the parent rocks were only partially consolidated when disrupted suggests that submarine landslides may have produced the breccia. Submarine volcanic eruptions might be expected to produce thick accumulations of volcanic debris which would be unstable, particularly if water-soaked and on steep constructional, volcanic slopes. Earthquakes accompanying the eruptions could undoubtedly trigger landslides of considerable magnitude. Great boulder conglomerates of Lower Paleozoic age near Quebec City have been interpreted as penecontemporaneous landslides because, among other reasons, of their "tumultuous and unbedded" internal arrangement, their gigantic boulders, and the fact that fossils found both in the boulders and in the matrix show a penecontemporaneous origin (Bailey, Collet, and Field, 1928, p. 603-604). Recently coarse limestone breccias of the Cow Head region in western Newfoundland were restudied, and a history of repeated penecontemporaneous chaotic landsliding from Middle Cambrian to Middle Ordovician time deduced (Kindle and Whittington, 1958). These submarine landslides were presumably initiated by earthquakes.

An alternative hypothesis is that the tuffaceous breccias are peperites – breccias formed by volcanic eruptions through unconsolidated or semi-consolidated sediments. Peperites have been studied extensively in the Auvergne region of Tertiary volcanism in France (Giraud, 1903, p. 299-367; Carozzi, 1960, p. 86-92). There, eruptions through lake beds formed breccias of mixed volcanic and sedimentary components, including much calcite and clay; some breccias are massive whereas others show faint bedding; there is a wide range of fragment sizes; and the complexity – within relatively small areas – of the geologic relations between breccias, intrusive rocks, and sedimentary rocks rivals that in the Cutler area. Presumably such eruptions might trigger landslides, and hence the two alternative hypotheses considered here are not mutually exclusive. However, the peperite hypothesis not only explains as adequately as does the landslide hypothesis the disruption of sediments, the mixture of sedimentary and volcanic fragments, and the tilting of large blocks of bedded rocks; it also explains the intrusive relations of some tuffaceous breccias and is in accord with the evidence that basaltic lava intruded unconsolidated mud and fine tuff.

Edmunds Formation

Age and correlation

The Edmunds formation was mapped by Bastin and Williams (1914) along the west shore of Cobscook River in the Eastport quadrangle and was named after good exposures near the town of Edmunds. Bastin and Williams (1914, p. 4-6) distinguish and map three lithologic units, rhyolite and andesite, diabase, and shale. The authors

present evidence that the contact with the underlying Dennys formation may be unconformable, but state that the contact is much faulted and locally appears to be concordant. A fairly abundant fauna in the shale member indicates a Silurian age.

The small area of Edmunds formation at the extreme northwest corner of the Cutler quadrangle is primarily an extension of the mapping of Bastin and Williams. There are only a few scattered outcrops and the rock itself is not notably different from the rhyolites of the Pembroke formation. If Bastin and Williams had not carried the contact between the Edmunds and Pembroke formations to the northwest corner of the Cutler quadrangle, the Edmunds formation in this report would probably have been mapped as part of the Pembroke formation.

Lithology

Flow-banded soda rhyolite tuff mapped as Edmunds formation crops out from under glacial drift in a few small patches in the extreme northwest corner of the Cutler quadrangle. The contact with Pembroke rocks to the southeast is not exposed. In outcrop the rhyolite weathers to a reddish-brown. Fresh surfaces show a dark gray dull glassy groundmass with many pink to white euhedral feldspar phenocrysts averaging about 2 mm. long. Dark gray to red fragments of rhyolite and very irregular flow banding can faintly be seen with a hand-lens.

Examination of thin-sections shows that the rhyolite consists primarily of flow-banded, devitrified glass in which are faint shadows of shards. This devitrified glass encloses scattered, euhedral plagioclase phenocrysts and small fragments of rhyolite and glassy, felsitic rocks.

The blocky, generally euhedral plagioclase phenocrysts are unzoned and albite twinned. Their composition is about An_{10}. They contain scattered minute flakes of white mica, probably sericite, and small patches of calcite. Many have irregular patches of very fine, slightly reddish dust, probably hematite staining, which gives them their pink color in hand specimen.

The glassy groundmass is devitrified to a banded fine-grained mosaic of chalcedony, quartz, and feldspar, and has a light reddish brown hematite stain. Flow-bands swirl around phenocrysts and the numerous irregular strung-out inclusions, and there are faint suggestions of

shards. Cognate inclusions form streaks and pods aligned parallel to the flow banding. They consist of reddish brown structureless glass now devitrified to feathery and dusty chalcedony, irregular patches of granular quartz, shreds of apple green chlorite, and a few euhedral oligoclase phenocrysts.

Origin

The Edmunds formation in the Cutler quadrangle is a devitrified soda rhyolite, probably a tuff. Whether or not it is marine is not known. It erupted as a mass of shards, tuff fragments, and plastic blebs of glassy rhyolite and was subsequently devitrified.

Pembroke Formation

Age and correlation

The Pembroke formation was originally defined by Bastin and Williams (1914, p. 6-7). They recognized and mapped four lithologies within it: rhyolite tuffs and flows; diabase tuffs and flows; the Hersey red shale member; and the Leighton gray shale member. They named the formation for the town of Pembroke on the north shore of Cobscook Bay. Fossils found in the shale members were assigned an Upper Ludlow Silurian age. According to Bastin and Williams, the Pembroke formation is conformable with the underlying Edmunds formation and the overlying Eastport formation. North of Cobscook Bay the Pembroke formation is chiefly Hershey red shale and Leighton gray shale, while south of the bay the formation is chiefly rhyolite and diabase flows and tuffs. The zone in which these two facies meet is beneath the waters of the bay.

The rocks labelled as Pembroke formation in the northwest corner of the Cutler quadrangle are so designated because they are a direct extension of the rhyolite member of the Pembroke formation as defined by Bastin and Williams. No fossils were found, but presumably these rocks are Silurian like their counterpart in the Eastport quadrangle.

Lithology

The Pembroke formation crops out on Great Ridge and in a low ridge to the northwest of the central valley. These ridges are locally free of glacial drift, and this combined with heavy timbering and a

network of timber roads leads to fairly numerous exposures compared to much of the Cutler quadrangle.

The rocks are tuffs and flows of generally a soda rhyolite composition, perhaps some rhyolite intrusive bodies, a few diabase flows and sills, and some poorly exposed argillite and tuffaceous shales. Excluding an intrusion of gabbro and assuming no folding, total thickness of the section is about 5,000 feet. The contact with the Edmunds formation is not exposed; and the southeast contact is presumably the Lubec shear zone covered by glacial drift in the central valley.

The bedding and banding in the formation strike north-northeast and are vertical or nearly so. Bedding dips steeply, and criteria for determining the top and bottom of beds were not found. The formation may be isoclinally folded, but there are no small order isoclinal folds to support this supposition. As the Edmunds formation is mapped in the Eastport quadrangle as being stratigraphically beneath the Pembroke, the steep northwest dips of the Pembroke in the Cutler quadrangle near Lively Brook may be of overturned beds.

The Pembroke formation is divided into three lithologic units: bedded tuffaceous sediments with minor shale and limestone (Sps), rhyolite flows and intrusives (Spr), and diabase flows and intrusives (Spd).

Sediments. In outcrop the tuffaceous rocks are generally gray; some are prophyritic and almost all show fragments of lapilli and inclusions of other rock types. In many exposures, the lapilli and inclusions are flattened parallel to bedding giving the tuffs a definite platy foliation.

Under the microscope it can be seen that the tuffaceous rocks consist predominantly of euhedral plagioclase, fragments of lapilli, and inclusions of felsitic rocks in a matrix of shards, more or less devitrified.

The phenocrysts of plagioclase are twinned albite, dusty with specks of sericite, chlorite, and calcite, and are commonly slightly stained by hematite. Some are euhedral; some are fragmentary; and others are rounded by irregular embayments of the groundmass. The lapilli inclusions have the same lithology as the rock as a whole. A few of the tuffs have irregular blebs consisting of granular quartz and albite, small euhedral laths of potash feldspar ($2\ V = 70$), and shreds of chlorite. These may be recrystallized inclusions.

The groundmass of shards and lapilli is commonly isotropic, devitrification being confined to the growth of small grains of quartz and albite and shreds of sericite and chlorite. The gray color of the tuffs apparently reflects many globulites and much fine magnetite dust throughout the shardy groundmass.

Gray, fine-grained bedded tuff, massive tuffaceous argillite, silty tuffaceous shale, and a few lens-like beds of limestone are exposed in several small outcrops on the low hill outlined by the 200 foot contour in the northwest corner of the Cutler quadrangle. Along the shear zone northwest of East Stream are a few outcrops of highly sheared black shale containing beds 1-2 inches thick of siltstone and tuff.

Rhyolite. The rhyolite of the Pembroke formation forms a wedge-like mass in the tuffaceous rocks south of Lively Brook and a more irregular mass on Great Ridge. Both masses are approximately concordant with the strike of the surrounding rocks; but as no contacts with the tuffaceous rocks were found, whether the rhyolite form flows or intrusive bodies is not known. Most outcrops display very intricately folded flow-banding, elongate pods of quartz around which the bands

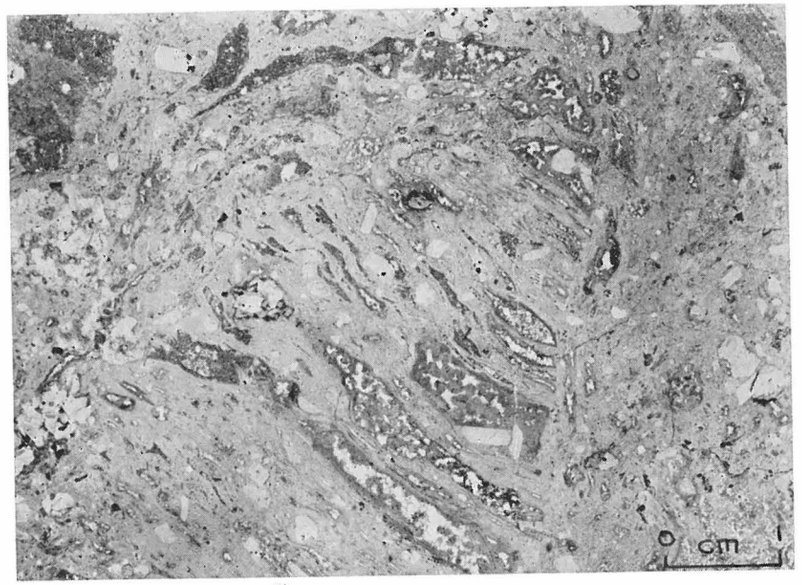

Fig. 13. Photomicrograph of flow-banded rhyolite from the Pembroke formation. Great Ridge.

swirl, scattered small phenocrysts of feldspar, and a pervading red color. Locally, particularly on Great Ridge, the rhyolite is a red to gray massive vitrophyre devoid of feldspar or quartz.

The microscope shows that the bands consist of minute hematite flakes scattered through fine-grained chalcedony. The intervening paler and thicker areas are of coarse chalcedony or fine-grained granular quartz enclosing a few subhedral to euhedral laths of albite and minute microlites of clear potash feldspar ($2V = 65$). Some faintly banded red rhyolites are composed of irregularly ellipsoidal amygdules of quartz, calcite, muscovite, and minor green chlorite in a once vitrophyric matrix now composed of chalcedony, sericite, very finely granular quartz and feldspar, and scattered small albite or sodic oligloclase phenocrysts.

The rhyolites of the Pembroke formation show varying degrees of alteration. Devitrification is widespread and patchy; plagioclase phenocrysts are dusty with hematite, sericite, chlorite, and calcite; and the minerals found in the amygdules are secondary. The fact that the potash feldspars have a large optic angle and are not sanidine, as would be expected in a volcanic rock, suggests either that they are secondary or have inverted to low temperature forms during a period of alteration. Furthermore, the optics of the albite phenocrysts indicate that these are also low temperature forms.

These rocks have been named soda rhyolite on the basis of their present mineral composition, predominantly quartz (or chalcedony), albite, and potash feldspar. However, it is possible that much of the quartz and particularly the potash feldspar were introduced hydrothermally rather than having been latent in the glassy groundmass. If so, these rocks when erupted may have had the composition of keratophyres or albite trachytes.

Diabase. Diabase and amygdular basalt form narrow sills and flows concordant with the bedding of the tuff or the banding of the rhyolite. Weathered outcrops of diabase or basalt are brown to dark gray and have rounded outlines in contrast to the jagged outcrops typical of the rhyolite. Some of the basalts have numerous white or green amygdules as much as a centimeter in diameter. The amygdular basalts are probably flows, and a few zones filled with ellipsoidal inclusions of highly amygdular basalt suggest flow tops.

The amygdular basalt consists of a swarm of small labradorite laths, giving a trachytic appearance, in a groundmass of minute magnetite

grains, small clots of granular pyroxene, and shreds of chlorite. The vesicles are filled with apple green chlorite and calcite, with chlorite and massive to granular epidote and a very small amount of calcite, or with quartz and epidote.

The microscope shows that massive diabase forming the sills has an ophitic texture. The plagioclase is euhedral to subhedral labradorite, with slight gradational zoning from a more calcic interior to a less calcic exterior, the total zoning being less than about 10% An. Much of the plagioclase is dusty with sericite, chlorite, and calcite, and along contacts with pyroxene there are small needles of actinolite. The anhedral pyroxene is augite with an optic angle of 48 degrees, untwinned and free of exsolution lamelli. In general, the pyroxene is less altered than the plagioclase but has patches and irregular stringers of chlorite and actinolite, particularly around the margins and in fractures and cleavages.

Origin

The Pembroke formation clearly reflects volcanic activity. Eruptions of siliceous and soda-rich magma formed the bedded tuffaceous rocks, and flows or intrusions at shallow depth produced the massive or flow-banded vitrophyric rocks. A lesser amount of basaltic volcanism led to the few flows and dikes of diabase and amygdular basalt. Bastin and Williams (1914, p. 14) concluded that the volcanic members in the Eastport quadrangle are marine as are the Leighten gray shale and the Hershey red shale members of the Pembroke. In the Cutler quadrangle, the few exposures of bedded tuff, argillite, and limestone suggest that the Pembroke formation here is also marine; but because they contain no fossils, confident correlation with the marine sedimentary members of the formation in the Eastport quadrangle is not possible.

INTRUSIVE ROCKS

Keratophyre

Field relations

Dikes and irregular plutons of a vitreous porphyry intrude the Little River formation at many places in the Cutler and Moose River quadrangles and to a minor extent the Pembroke formation. This porphyry forms blocky, hackly outcrops, is commonly laminated or banded, and contains conspicuous white phenocrysts of feldspar in a red, brown, or dark gray glassy groundmass. Locally, particularly at Almore Cove, phenocrysts of quartz or rounded blebs of granular quart accompany the feldspar. Study of thin-sections shows that the feldspar phenocrysts are albite and that the original vitreous groundmass has been altered to a fine-grained mosaic of quartz, albite, and a minor amount of chlorite. These rocks are therefore named keratophyre, although the accompanying suite of basaltic and diabasic rocks in the Cutler quadrangle are not considered to be typical spilites with which keratophyres are commonly associated.

Two principal centers of keratophyre have been mapped, one associated with the tuffaceous breccia at Norse Pond and extending south to Holmes Cove and Schooner Brook, the other on Little Machias Peninsula. In these areas, the keratophyre displays a considerable range in the size and number of phenocrysts, in degree of flow-banding, and in amount of brecciation. Likewise its relations to the Little River formation differ from place to place. Its general map pattern indicates that much of the keratophyre intrudes the Little River formation; and dikes with chilled and slightly vesicular margins, commonly with inclusions, cut the tuffaceous breccias or bedded rocks of the Little River formation at Bog Brook Cove, Holmes Cove, Almore Cove, and along the south shore of Little Machias peninsula. On the other hand, at Bog Brook Cove and Black Point Cove, the keratophyre is layered concordant with bedded tuffs of the Little River formation, and faint shadows suggesting shards can be seen in thin-sections. Furthermore, tuffaceous breccias at Norse Pond and on Little Machias peninsula contain fragments of keratophyre as well as many small dikes and irregular intrusions of keratophyre.

Thus much of the keratophyre appears to be both intrusive and extrusive and belong to the same volcanic episode that produced the Little River formation. However, some keratophyre intrudes the Pembroke formation and thus is not contemporaneous with the Little River volcanics. On the geologic map thin dike-like bodies of keratophyre appear to intrude the Cutler diabase, suggesting a post-diabase episode of intrusion. Actually, there are probably thin lenses or septa of keratophyre trapped between successive intrusions of diabase.

Lithology

Study of thin sections shows that the albite phenocrysts make up about 5 per cent of the rock either as discrete, commonly fragmentary crystals, or as somewhat fractured crystal clots. The phenocrysts average 1 to 2 millimeters in length, have no zoning, and are dusty with flecks and pin-points of chlorite, calcite, and less commonly epidote. Optic angle and extinction angle indicate a low temperature albite of composition An_4. In several samples, these albite phenocrysts have irregular veinlets and outside rim of potash feldspar. This rim encloses broken crystals as well as euhedral ones indicating that the rim grew after fracturing of the phenocryst. Universal stage determinations found optic angles ranging from about 60 to 75 degrees for the potash feldspar showing that it is a low temperature form, not sanidine.

The groundmass is a fine-grained mosaic of anhedral quartz, albite, and potash feldspar, with minor whisps of chlorite, patches of calcite, red blebs of hematite, a few euhedral crystals of sphene and zircon, and considerably clotty leucoxene. In some exposures the groundmass is fractured into polygonal fragments outlined by zones of coarsely devitrified glass that follow the cracks. At other places it shows considerable parallel or en-echelon shear fracturing in which chlorite and a little olive brown biotite along with hematite, magnetite, or leucoxene grains outline the fractures. The red and black or red and green mottling so distinctive in some outcrops appears in thin-section to reflect the state of oxidation of the magnetite or hematite dust in the groundmass.

Origin

The keratophyres are considerably altered rocks and their original mineralogy, particularly the composition of the plagioclase phenocrysts, can only be surmised. They appear to have undergone the same type of metamorphism which converted the surrounding Cutler diabase partially into greenstone. The phenocrysts of the keratophyre have the same composition and the same type of saussuritization as

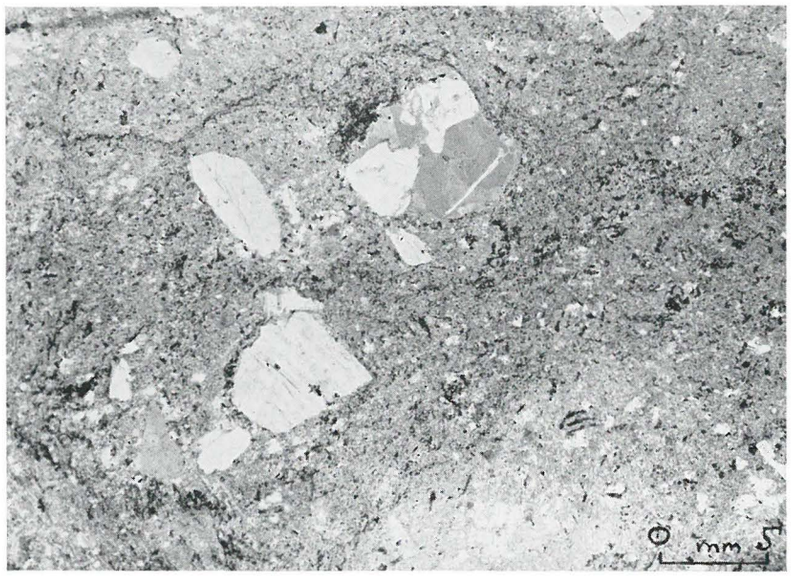

Fig. 14. Photomicrograph of typical keratophyre porphyry. Note broken albite crystals. Crossed nicols. Holmes Cove.

plagioclase crystals in the diabase. The latter were labradorite originally, hence the phenocrysts of the keratophyres might once have been a more calcic plagioclase, perhaps oligoclase or andesine. However, it is doubtful if the keratophyres were once typical andesites because they lack any evidence of the ferromagnesian mineral content typical of andesites.

The keratophyres were intruded at shallow depth into the Little River and Pembroke volcanics and also have formed some flows and tuffs interbedded within the Little River formation. The brecciation apparent in some exposures and their association with tuffaceous breccia at Norse Pond and on Little Machias peninsula, suggest an essentially volcanic origin, probably in part contemporaneous with formation of the Little River volcanic rocks and in part younger.

Cutler Diabase

More than half the bedrock in the Cutler and Moose River quadrangles is intrusive diabase, most of it the direct continuation of the diabase southeast of the Lubec shear zone in the Eastport quadrangle. The term diabase is used here to designate collectively a complex of

hypabyssal intrusive rocks consisting essentially of labradorite and pyroxene. The term dolerite might be equally applicable. The grain size and textures of the Cutler diabase range from those typical of basalt to those typical of coarse gabbro. Much of the Cutler diabase now consists of albite, chlorite, epidote and actinolite, minerals typical of greenstones. However, the original textures survive; and in many places so do the primary labradorite and pyroxene. Hence the term diabase is considered more descriptive of the rock genetically than the terms greenstone or metagabbro, which bear a connotation of regional metamorphism not strictly applicable here. Because of excellent exposures near the town of Cutler, the name Cutler diabase seems appropriate.

Field relations

Although the geologic map implies that the Cutler diabase is one large body apparently intruded at one time, it is actually an intricate network of many interlacing separate sills, dikes, small lenticular plugs, and irregular plutons. These cut not only the Silurian formations and keratophyre but also each other. The general pattern resembles that of the Karroo dolerites of South Africa (Du Toit, 1920, p. 4-14; Walker and Poldervaart, 1949, p. 604-611).

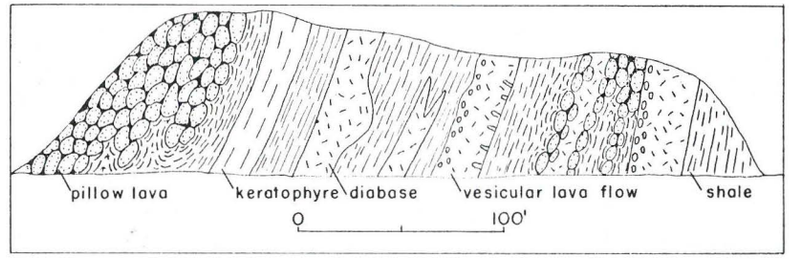

Fig. 15. A section of the Little River formation exposed in a cliff at Eastern Head. The beds dip south.

Generally concordant intrusions are well exposed along the shore north of Moose River and north of Eastern Head. Concordant contacts of gabbro with bedded argillite and tuff can be seen at survey station Hues west of Fairy Head and on the south shore of House Cove; and small sills of diabase cut bedded rocks at many places along the coast line. Some sills gradually pinch out, suggesting they were intruded as a wedge. Others end in an abrupt right-angled termination as if intrusion was dammed by a fault plane along which

the beds overlying or beneath the intrusion had moved to make way for the magma.

Cross-cutting dikes, some at right angles to the regional strike and others clearly oblique, are also apparent on the geologic map and exposed on a small scale in many outcrops along the shore and inland. Some of them make almost right-angled turns, suggesting stepping from one horizon to another. It is evident, however, that bedding was not the sole control of magma emplacement. Much original bedrock must have disappeared to make room for many of the discordant gabbro bodies.

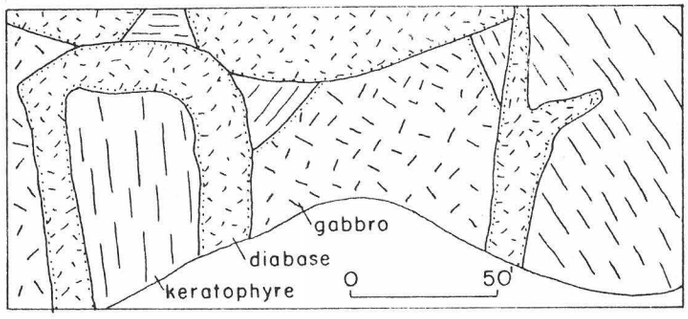

Fig. 16. Map of intrusive relations along the shoreline north of Bog Brook Cove to illustrate the complexity typical of the Cutler diabase.

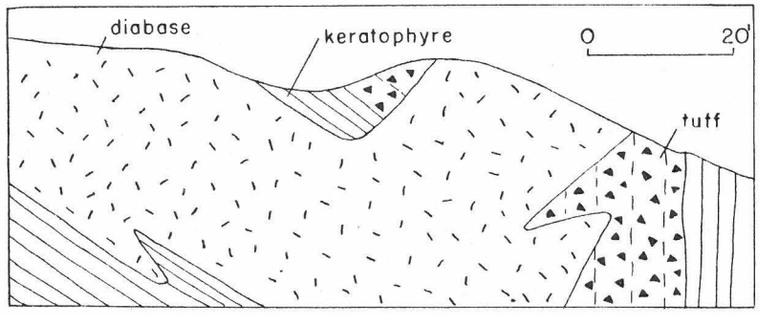

Fig. 17. A section exposed in a cliff along the shore north of Bog Brook Cove to illustrate tilting of blocks (lower right) of the host rocks.

Excellent exposures along the shore line commonly show contacts in which coarse-grained gabbro terminates abruptly against another gabbro along a fine-grained and commonly porphyritic contact zone from a few inches to a few feet wide. Many sheets of diabase with excellent columnar jointing cut older diabase or gabbro, and diabase

dikes ranging from a few feet to 20 feet wide are numerous. Unfortunately these contact zones cannot be traced inland through the timber and hence be mapped. It is probable, however, that many of the septa of keratophyre or sedimentary rocks shown on the geologic map are thin sheets trapped along the contact between two separate intrusions.

Intrusive contacts between the Cutler diabase and older rocks are invariably sharp, are generally free of inclusions, and show little evidence of contact metamorphism. A selvage from an inch to 3 or 4 inches wide of glassy diabase characteristically marks contacts of the diabase with bedded rocks, keratophyre or older diabase. This contact is sharp even on the scale of a thin section. The diabase within the contact zone generally consists of devitrified basaltic glass containing scattered plagioclase phenocrysts. Pyroxene phenocrysts are rare. Fine-grained tuffs or argillites commonly appear to be sheared and smeared out along the contacts; but coarse tuff or gabbro shows little evidence of deformation, the contact going around tuff fragments or mineral crystals.

Inclusions of host rocks in the diabase, even close to contacts, are the exception rather than the rule. In several places where gabbro invaded coarse breccia the contact is lined with breccia fragments, but contacts with fine-grained sedimentary rocks and tuffs or with other diabase bodies are normally free of inclusions. Even in thin-section, tuff fragments or individual mineral grains from the host rock are not common in the glassy contact selvage.

Evidence of chemical reaction, recrystallization, metasomatism, or melting of the host rock along contacts of the diabase is less prevalent than might be expected from the abundance of intrusive activity. Diabase or gabbro commonly has a zone several inches wide rich in vesicles filled with chlorite, carbonate, quartz, epidote, or pyrite along contacts with fine-grained tuffs, argillites, or shale. This suggests that the magma incorporated some water from the sedimentary rocks.

No zones of extensive hornfels or of porphyroblastic mineral growth were encountered along contacts with tuffaceous or argillaceous host rocks. Some contact zones show a slight change of color and a tendency to form a blocky or hackly fracture, and thin-sections of these zones show that some of the finer-grained sediments have recrystallized to a fine-grained mosaic of quartz, feldspar, and sericite. On the other hand original grain outlines and sedimentary structures are still

apparent in thin sections taken from many contact zones. The alteration that produced chlorite, epidote, albite, and calcite in many of the diabasic rocks, both intrusive and extrusive, also affected many of the fine-grained tuffs and argillites, and perhaps this common alteration has obscured some of the contact metamorphism once present. However, on the whole there appears to have been little chemical or thermal reaction between the Cutler diabase and intruded rocks.

Lithology

The Cutler diabase in most outcrops is readily recognized by its massiveness, its tendency to form rounded outcrops when weathered or steep-sided rounded hills, knobs, and roches moutonées where scoured by glaciers; its irregular blocky jointing, its dark gray to green color, and its conspicuous white feldspar laths in the coarse-grained varieties. Probably while mapping, I have mistaken some coarse-grained lava flows for intrusive diabase, some amygdular contact zones of intrusive diabase for lavas, and some sheared intrusive diabase for diabase tuff.

Much of the intrusive diabase has been metamorphosed to a greenstone. Chalky white feldspar laths are surrounded by dark green chlorite, some of which clearly has replaced pyroxene; and veins of chlorite or calcite ramify through the rock. A thin-section of a typical moderately coarse-grained gabbro shows laths of twinned but unzoned albite speckled with chlorite, epidote, and calcite; ragged remnants of pyroxene surrounded and veined by actinolite and chlorite; interstitial areas filled with chlorite, grains of epidote, and needles of amphibole; and irregular opaque masses of leucoxene containing thin laminae of magnetite.

However, this metamorphism is very unevenly distributed; and some of the diabase is sufficiently unaltered to permit identification of its primary mineralogy. This section will be concerned with this primary mineralogy and texture.

In a few thin-sections, the plagioclase feldspars, although veined and sutured by albite, still retain their original composition. They are albite and pericline twinned and gradationally zoned. The centers of large plagioclase crystals — up to 5 millimeters long in some coarse-grained gabbros — are calcic labradorite An_{65}, whereas smaller plagioclase crystals in finer-grained gabbro or in diabase are about An_{55}. These calcic centers grade outwards to crystal margins

ranging in composition from An_{40} to An_{50}. Commonly there is an abrupt change to a thin outside rim of about An_{20}. A few rims are of plagioclase as sodic as An_{12}. All the plagioclase crystals studied on the universal stage have optical properties which best fit those of low temperature plagioclase.

Pyroxene generally shows greater resistance to alteration than the plagioclase. Remnants of pyroxene occur in almost all the 40 thin sections studied of the Cutler diabase; and in many of the least altered samples at least 75 per cent of the original pyroxene has survived.

Clinopyroxene is the only pyroxene present in the diabase. The most common occurrence is in subophitic or ophitic clots which have a maximum diameter of 10 millimeters in the coarse-grained gabbros. Pyroxene also forms single subhedral to anhedral crystals or clots of three or four anhedral grains having irregular mutual contacts interstitial to plagioclase. The clinopyroxene present is augite, slightly richer in lime and ferrous iron than the center of the augite field (Hess, 1941, p. 518), according to tables of optical properties by Hess (1949, p. 634). Of the 18 samples in which the pyroxene optic angles were measured on a universal stage, 12 gave optic angles between 50 and 53 degrees, and the maximum range was from 40 to 60 degrees. ZΛC measured on 4 samples ranged from 40 to 43 degrees, and the Y index was about 1.700 in sodium light. A few pyroxene grains showed slight gradational zoning; but most extinguished uniformly under crossed nicols except when strained; and no ex-solution lamellae were seen.

Some of the Cutler diabase originally contained interstitial glass or chlorophaneite that has since been altered to chlorite. In a number of fairly fresh samples very pale green chlorite fills interstitial areas between plagioclase crystals with no indication that it has replaced an earlier pyroxene. This chlorite shows a very faint, very pale green pleochrism, is nearly isotropic, has a small positive optic angle, is fast parallel to the cleavage, and has an N_0 index of 1.620 in sodium light. These optical properties suggest ripidolite according to the data of Hey (1954).

Accessory minerals include apatite, which occurs as small euhedral crystals in the interstitial chlorite, and titaniferous magnetite commonly showing exsolution lamellae of magnetite and ilmenite. A few samples contain interstitial quartz, one as much as 10 per cent. These

quartz bearing diabases are also altered; hence this quartz may be either primary or an alteration mineral. No residual quartz-feldspathic interstitial material was seen in either fresh or altered diabases.

The original proportions of these primary minerals can only be estimated because in making modal analyses it was impossible to decide precisely which primary mineral was replaced by secondary minerals such as chlorite, epidote, or actinolite. If, in the 10 modes made by point count of the least altered diabases and gabbros, all the amphibole is assigned to original pyroxene and the chlorite and epidote is split evenly between plagioclase, pyroxene, and the interstitial chlorite residuum; then the average modal composition in volume per cent is: plagioclase 55.4, pyroxene 33.0, chlorite residuum 8.0, magnetite-ilmenite 3.6. A single modal analysis of the least altered specimen, a coarse-grained gabbro, gave the following volume percentages; plagioclase 52.0, pyroxene 27.3, actinolite 5.1, chlorite 9.6, epidote 1.4, opaque minerals 4.6.

The phenocrysts present in glassy border phases and the textural relations of the main minerals suggest that the order of beginning of crystallization was magnetite-ilmenite, plagioclase, pyroxene, apatite, and chloritic residuum.

The mineralogy of the diabase suggests that it crystallized from a subalkaline basaltic magma of the non-porphyritic central type of Mull (Bailey and others, 1924, pp. 16-18) probably somewhat differentiated towards iron enrichment as suggested by the ripidolite residuum and slightly iron-enriched augite.

The absence of a second pyroxene, normally pigeonite, is an exception to the generalizations on pyroxene crystallization made by Poldervaart and Hess (1951, p. 478-479). The basaltic rocks of the Deccan in India are another exception and carry only one pyroxene, augite. Sukheswala and Poldervaart (1958, p. 1491) suggest that if the magma is relatively rich in iron, pigeonite will not crystallize, but will be represented by a mesostasis of iron-rich chlorite. This explanation may also be applicable to the Cutler diabase.

Textural variations

The Cutler diabase displays a great range of grain size and textures. Much of this variation can be conventionally explained as the result of variations in cooling rate. Contacts typically are fine-grained and porphyritic compared to the central bulk of the intrusion. Most sills

and dikes from 1 to 5 feet wide are of fine-grained diabase in which the feldspar laths average approximately 1 millimeter in length. The thicker intrusions, 30 feet or so in width are gabbroic, and the feldspars average 3 to 4 millimeters in length. A few large gabbro bodies have feldspars and pyroxene crystals averaging about 1 centimeter.

However within many gabbro bodies, variations in grain size, differences in spacial relations of pyroxene or uralite to plagioclase, and in changes in the ratios of plagioclase to ferromagnesian minerals do not lend themselves to a simple explanation based on rate of cooling.

Banding. Banding is common in many exposures of the Cutler Diabase. On the south shore of Deer Island a section about 40 feet thick of banded gabbro is fully exposed (Fig. 18). The banding is so striking that from a distance the rocks were first identified as bedded sediments. The bands are distinguished by their color which in turn primarily reflects the ratio between plagioclase and uralitized pyroxene. Dark bands have a 1/1 ratio of plagioclase to uralitized pyroxene and the light bands a ratio of about 4/1. Commonly the contact between a dark band and the underlying light one is sharp whereas the contact with the overlying light band is gradational. However in some bands

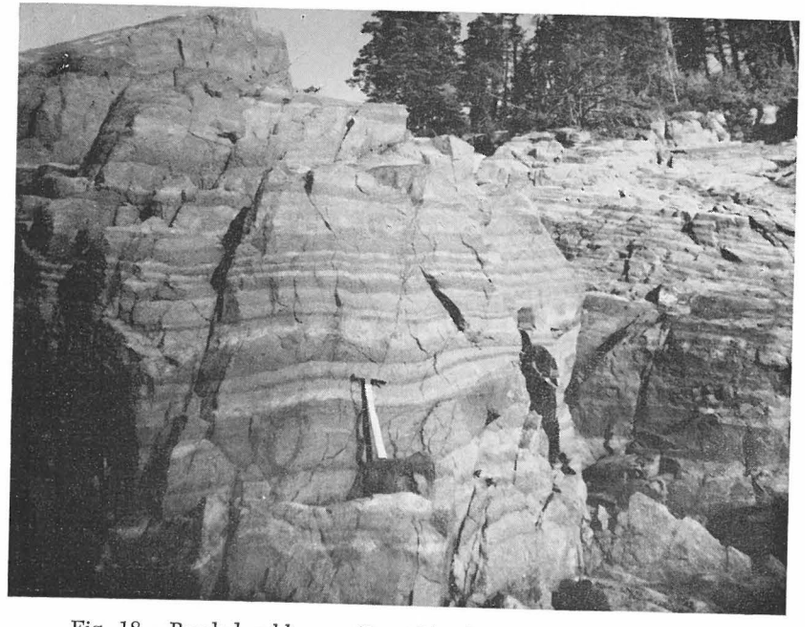

Fig. 18. Banded gabbro on Deer Island.

Fig. 19. Photomicrograph of plagioclase-rich band. The saussuritization of the plagioclase is typical of much of the Cutler diabase. Note the orientation of the plagioclase crystals. The black mineral is magnetite. At lower right is a hornblende crystal with a few remnants of pyroxene. Crossed nicols. Deer Island.

both contacts are sharp and in other bands both contacts are gradational. In thin-section the sharp contacts show a gradation through a zone about 1 centimeter thick.

Individual bands differ from each other considerably in thickness but maintain a fairly regular thickness throughout their exposure. Thicknesses range from an inch to about a foot and single bands extend as much as 60 to 100 feet, gradually lensing out. The zone of distinctly banded gabbro grades downward through increasingly thick and decreasingly differentiated bands into an essentially homogenous gabbro.

The lower and less well banded section of the sequence is a pyroxene-rich gabbro in which the ratio of pyroxene to plagioclase is about

1/1, the same ratio as in the dark bands of the upper well-banded section. The banding thus results from an increase in the amount of plagioclase relative to pyroxene which occurred periodically.

The dark bands generally have a smaller grain size (1-3 millimeters) for both the uralitized pyroxene and the plagioclase than the light bands in which the median grain size is 4-5 millimeters and the maximum as much as one centimeter. Thin-sections cut at right angles to the light bands show many plagioclase laths lying approximately parallel to the bands. This gives the impression of a good lineation. However when viewed perpendicular to the banding, the laths are seen to be randomly oriented in the plane of the banding. In other words, they are stacked like jackstraws rather than like cordwood to form a foliation. In the finer-grained uralite-rich dark layers, the comparatively stubby pseudomorphs of pyroxene crystals prevent the more lath-like plagioclase crystals from developing as good a foliation as is seen in the light layers.

The conventional hypothesis that crystal settling is primarily responsible for formation of bands in gabbro is applicable here, but the origin of the differences in mineral proportions and grain size from band to band is an open question. As both pyroxene and plagioclase settled together, both were probably crystallizing together, the magma being on the plagioclase-pyroxene boundary surface in the albite-anorthite-diopside-$FeSiO_3$ system (Bowen, 1928, p. 65). Yoder (1955, p. 1638) found that in the anorthite-diopside-water system, an increase in water vapor pressure moved the boundary curve between anorthite and diopside towards anorthite. This led him to suggest that variations in water vapor pressure on the magma, perhaps controlled by volcanic eruptions, might account for pyroxene-rich and plagioclase-rich bands in basic intrusions. The fact that crystals are larger in the plagioclase-rich layers of the Deer Island gabbro is in accord with Yoder's theory. Presumably a magma rich in volatiles would be less viscous, diffusion would be more rapid, and hence crystals would grow more quickly than in a drier magma. The main difficulty of Yoder's theory when applied to the narrow bands and the large variation in mineral proportions characteristic of the Deer Island gabbro is that numerous eruptions, large changes in vapor pressure, and rapid adjustments of equilibrium within the magma are required. Hess (1960, p. 133-137) has proposed an alternative theory based on his study of layering in the large Stillwater complex. He demonstrates that differences in settling velocities between pyroxene and plagio-

clase coupled with the up-or-down movement (primarily up) of convection currents could effectively control the relative proportions of these minerals in the bands. The difficulty with this theory when applied to the Deer Island gabbro is that the Cutler diabase is a complex of many small intrusions in which convective over-turn is not as likely to occur as in the large Stillwater pluton.

Pegmatite clots. Clots of gabbro pegmatite are another common textural variation in the Cutler gabbro (Fig. 20). They may occur sparsely scattered throughout a gabbro intrusion, or as irregular clusters, or as concentrations in rather vaguely defined zones. They have a great variety of shapes and sizes. Some are spherical, ellipsoidal, or lens-shaped; others are ameboid with rounded projections and indentations; others consist of irregular arms radiating octopus-like from a central body; still others are long, narrow, twisting dike-like masses. The clots of pegmatite that are approximately equidimensional range from less than a foot to 5 feet in diameter whereas some of the dike-like forms can be traced for 10 to 20 feet.

These clots consist of large laths of plagioclase ophitically to subophitically enclosed in large crystals of uralitized pyroxene or intergrown with long lath-like crystals of hornblende which may not be derived from pyroxene. The plagioclase is saussuritized albite with the same composition and habit as that in the surrounding normal gabbro. Some clots have pale green chlorite in the interstices between plagioclase crystals and others contain large interstitial patches of calcite replacing amphibole and plagioclase. Quartz-feldspathic mesostasis material is absent. Most, but not all, of the clots have a concentration along their borders of long narrow crystals of uralitized pyroxene or possibly primary hornblende. Within the clots themselves there is commonly a very inhomogenous distribution of plagioclase and uralite or pyroxene; some irregular areas are rich in plagioclase, others rich in hornblende.

Gabbro pegmatite clots resembling these in the Cutler diabase are common in other diabasic rocks. Typical examples are the Whin sill in England (Tomkeieff, 1929, pp. 102-114), Triassic diabase in Virginia (Shannon, 1926, p. 18-39), and pre-Cambrian diabase on the north shore of Lake Huron and in the Sudbury area (Collins, 1925, p. 80-82). Ideas concerning the origin of pegmatitic clots in diabase can be grouped into two main categories. (1) Many clots have considerable quartz-feldspathic mesostasis material or centers of recrystallized

quartzite inclusions. Collins (1925, p. 80) and Phemister (1934, p. 110-119) conclude that some clots thus result from reaction between the magma and inclusions. (2) Many gabbro pegmatite clots contain no evidence of derivation from inclusions, and these are generally attributed to concentrations of volatiles within a partially crystallized magma, the volatiles reducing viscosity and thus aiding rapid diffusion of constituents to make large crystals. (Tomkeieff, 1929, p. 114-120, Shannon, 1926, p. 21).

In the Cutler diabase, the absence of partially altered inclusions and the very irregular shapes of some of the clots argue against their being either recrystallized foreign inclusions or cognate inclusions of previously crystallized coarse gabbro. The hypothesis that these clots represent concentrations of volatiles in a crystallizing magma is more attractive. Presumably the more or less spherical clots were bubbles in a quiet magma. A magma in motion might knead, squeeze, stretch, or twist such a bubble into a great variety of shapes and sizes, such as many clots do assume.

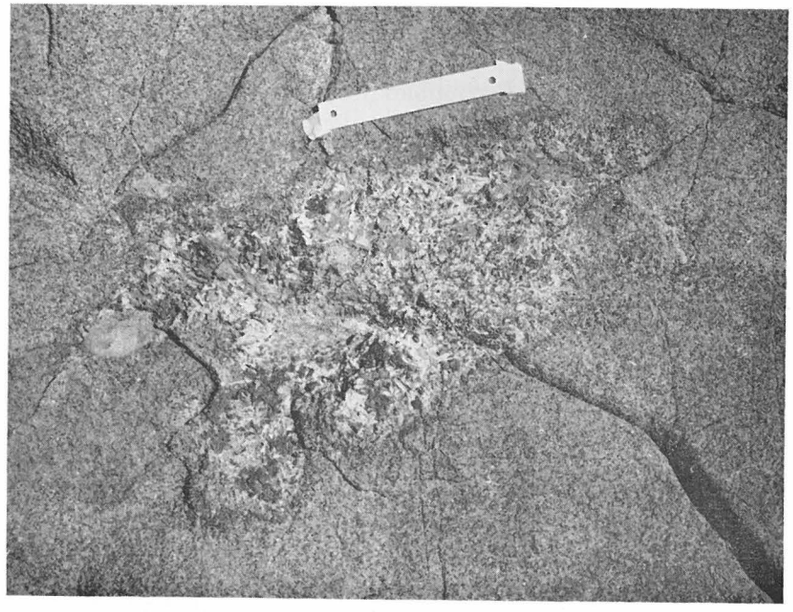

Fig. 20. A typical ameboid gabbro pegmatite clot in normal gabbro. Note the dark fringe of uralitized pyroxene. The rule is 6 inches long. Western Head.

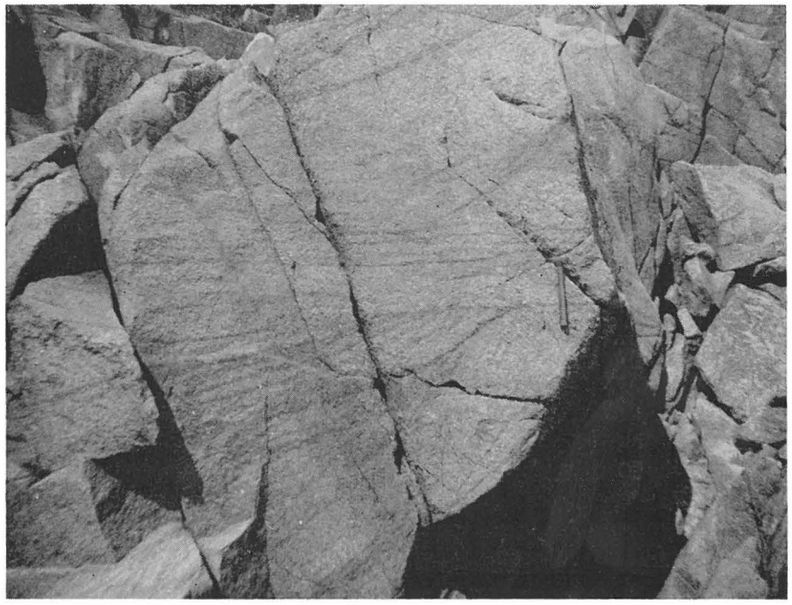

Fig. 21. "Cross-bedding" in gabbro. The dark "beds" are rich in uralitized pyroxene. Western Head.

It is unlikely that these bubbles were of magma saturated with volatiles, here assumed to be largely water, surrounded by unsaturated magma. This requires that there be no diffusion of water along a concentration gradient and that liquid immiscibility exist in the magma. The alternative is that these clots represent bubbles of a separate gas phase in the crystallizing magma. The bubbles formed because the magma became fully saturated with water. Saturation could arise from a number of causes: decrease in pressure as the magma works upward in the crust; decrease in vapor pressure due to gas eruptions at the surface; second boiling as crystallization of dry minerals concentrated volatiles in the remaining magma; absorption of water from the host rocks by an unsaturated magma; or most likely a combination of several of these. Crystallization of plagioclase and pyroxene in the bubbles as the magma crystallized would require multiphase equilibrium between magma, gas, plagioclase, pyroxene (or hornblende), and any other crystals also forming.

Evaluation of this bubble hypothesis requires that the composition of the main primary mineral phases in the clots and in the surrounding gabbro be known. Unfortunately unaltered clots in unaltered gabbro

were not found in the Cutler or Moose River quadrangles. Hence this possible solution to the problem of the origin of the clots remains only a working hypothesis.

The origin of the clots is of interest because they may provide a clue to conditions of crystallization of the Cutler diabase. Should the clots prove to have been gas bubbles in a fully saturated magma, and furthermore should the primary ferromagnesian mineral in the clot and the surrounding rock have been pyroxene – which is probable – then the vapor pressure at which the magma was saturated with volatiles must have been low. The melting curves of an Hawaiian olivine basalt under various water pressures determined by Yoder and Tilley (1956, p. 170) suggest that at a water vapor pressure above about 1,000 bars basaltic magma should crystallize to plagioclase and hornblende, below about 1,000 bars to plagioclase and pyroxene. A low vapor pressure of saturation would suggest either crystallization at a shallow depth in the earth's crust in a closed system where the vapor pressure equalled lithostatic pressure, or alternatively in an open presumably volcanic system in which the vapor pressure was essentially the hydrostatic pressure.

Miscellaneous. In addition to the banding and the pegmatite clots, the Cutler diabase displays a variety of textural differences many of which appear to reflect turbulence within magmas carrying plagioclase and pyroxene crystals when intruded. On Western Head, gabbro has cross-bedding (Fig. 21) so similar to the cross-bedding found in sedimentary rocks that deposition of crystals from a current of magma is immediately suggested. Also on Western Head irregular blocks of diabase are surrounded by gabbro in which the plagioclase laths have a lineation approximately concentric to the blocks indicating that either the blocks were sinking in the magma or the magma was flowing around them. Lineation of plagioclase laths parallel to contacts with other rocks is common, and a few diabase dikes and sills display this lineation throughout. Another characteristic of the Cutler diabase in many places, notably along the shoreline north of House Cove and north of Bog Brook Cove, is intimate mixing of coarse-grained gabbro with fine-grained in long, swirling, irregular streaks, lenses, or pods. Such concentrations of large crystals could indicate either incomplete mixing of two magmas carrying different amounts and sizes of phenocrysts or concentration of crystals by turbulence and eddies in the magma in the manner suggested by Balk (1931, p. 404-413) on a much larger scale for the Adirondack anorthosites.

Origin

In summary, the Cutler diabase crystallized from a basaltic magma – probably of the non-porphyritic central type slightly iron enriched – which repeatedly intruded a sequence of varied submarine volcanic rocks to form a complex of sills, dikes, and irregular plutons. Some intrusions apparently crystallized quietly with crystal settling, whereas others were turbulent. The magma recrystallized or metasomatized the host rocks only to a minor extent and in turn was little contaminated by assimilation of inclusions.

The Cutler diabase and the extrusive basaltic and diabasic rocks of the Little River formation have identical mineralogy, suggesting derivation from a common source magma. The extrusive rocks probably were the forerunners of intrusion at depth, and the two in fact may have been in part contemporaneous.

The abundance of igneous rocks, both extrusive and intrusive, of basaltic composition indicate that a large body of basaltic magma, which was cooling, crystallizing and differentiating very slowly, underlay the area. The abundance of intrusive diabase relative to intruded rocks, the lack of both structural and stratigraphic correlation from block to block of the Little River formation, and the mixture of concordant, discordant, and irregular intrusions suggest that the rocks of the Cutler and Moose River quadrangles once formed the fractured and perhaps subsiding roof zone of an underlying basaltic magma chamber.

The age of the Cutler diabase is given as Silurian(?) on the geologic map. Some of it may be approximately contemporaneous with the Little River formation; some clearly is younger than the Little River and Pembroke formations; and in the Eastport quadrangle diabase intrudes the Eastport formation, the youngest Silurian rocks, and to a minor extent the Perry formation of Upper Devonian age (Bastin and Williams, 1914, p. 12). The great bulk of the Cutler diabase, however, is older than the Perry formation as it has been sheared and folded during the deformation which preceded development of the unconformity beneath the Perry red sandstones and conglomerates. A few narrow dikes exposed along the shoreline in the Cutler quadrangle are unsheared and cut across the structure and hence may be post-Devonian in age. The conclusion of Bastin and Williams (1914, p. 12) that most of the diabase in the Eastport quadrangle may range from middle Silurian to late Devonian, but prior to deposition of the Perry

formation, is probably also applicable to the Cutler diabase of the Cutler and Moose River quadrangles.

METAMORPHISM

The original minerals of the Little River basaltic volcanic rocks and the Cutler diabase were principally labradorite and augite with a minor amount of titaniferous magnetite and some interstitial residual chlorite. These rocks now have a much more complex mineralogy. Albite, epidote, chlorite, and calcite have replaced labradorite; chlorite, actinolite, hornblende, and epidote have replaced augite and the interstitial chlorite; and the titaniferous magnetite is now largely leucoxene. Over 90 per cent of the metamorphosed rocks studied in thin-section (about 50) contain albite, chlorite, and leucoxene or sphene; about 75 per cent have epidote in substantial amounts; about 50 per cent have actinolite, and about 20 per cent have hornblende. Mineral assemblages include albite-epidote-chlorite, albite-epidote-chlorite-actinolite, albite-epidote-actinolite-hornblende, and albite-hornblende. A few rocks contain minor amounts of quartz or biotite; calcite is common; and pyrite and pyrrhotite are widespread in minute amounts. Except in highly sheared rocks and those massively replaced by granular epidote, original textures remain.

Neither the degree of alteration, measured by the amount of unaltered labradorite and pyroxene, nor the occurrence of particular mineral assemblages shows a pattern that can be correlated with geographical position, size of any particular rock body, grain-size or texture, degree of shearing, or intrusive versus extrusive origin of the basaltic and diabasic rocks. Variations in alteration may occur from outcrop to outcrop not 100 yards apart. Within the 32 square mile land area of the two quadrangles, may be found mineral assemblages typical of either the greenschist metamorphic facies or albite-epidote-amphibolite facies, as well as little altered basic rocks whose primary magmatic minerals essentially represent the pyroxene hornfels facies.

The metamorphism appears to have resulted from a two stage process. The first stage was the development of albite-chlorite-amphibole plus disseminated epidote assemblages irregularly distributed through the basalts and diabases. The second was an epidotization concentrated locally in clots, veins, and dikes both in basaltic rocks and keratophyres and to a minor extent in bedded tuffs and breccias.

First Stage

The course of the first stage of metamorphism can be traced from thin-section to thin-section of increasingly metamorphosed rocks. In labradorite, metamorphism began with the growth of sericite and chlorite in cleavages and fractures, a slight dusting of the entire crystal, and the growth of actinolite needles along contacts with augite. Next came formation of narrow irregular veinlets of albite in the labradorite and growth of albite outward from cracks and cleavages. Even at this early stage, the albite has erased all zoning, although those parts of the crystal still labradorite retain their original zoning. Replacement appears to have been controlled largely by fractures rather than composition of the zones in the original labradorite. Finally the crystal is all albite, commonly cloudy with specks of chlorite, minute actinolite needles, hematite(?) flakes, and pin-heads of calcite. In many rocks small anhedral grains of epidote are also present, but completely albitized plagioclase without a grain of epidote is not uncommon. All trace of zoning has disappeared, although the twinning, size, and shape of the original labradorite remain. The albite has optical properties suggesting a low temperature form of composition An_4 to An_6.

Alteration of labradorite to albite merely by subtraction of calcium and aluminum would greatly reduce the volume of each crystal, which clearly has not occurred. In only a few rocks does epidote replace a significant volume of plagioclase, and some albites are only slightly dusty with alteration products. Albitization must also involve addition of sodium and silicon. The presence of quartz and small twinned euhedral crystals of clear albite in many of the amygdules in the basaltic tuffs and lavas indicates that sodium, silicon, and aluminum, were mobile.

Augite has withstood metamorphism better than labradorite. In many rocks in which all the labradorite has gone to albite, more than one-half the original pyroxene remains. In very few samples studied are there no remnants of augite. In comparatively unaltered rocks, the pyroxene has a fringe of actinolite along contacts with labradorite; chlorite follows cracks and cleavages within the crystal; and contacts with interstitial chlorite are a feathery intergrowth of pyroxene and chlorite. In the more altered rocks, chlorite extensively replaces augite, and epidote may be present. In a few specimens, however, augite alters to a blue-green calcic hornblende, either directly or through an

intermediate stage in which any actinolite and chlorite present assumes a dark green color prior to becoming hornblende.

The very pale interstitial ripidolite found in the least altered diabases commonly, on metamorphism, turns a slightly darker green, assumes the anomolous blue interference colors of penninite (although the optic angle and N_0 index remain the same as in the ripidolite), and becomes indistinguishable from the chlorite replacing plagioclase and pyroxene. In most of the altered rocks chlorite is ubiquitous, occurring as vesicle fillings, as massive replacements of pyroxene and albite, following shears and cracks in mineral grains, and ramifying through the rocks in branching irregular veinlets. In those rocks, however, in which augite has altered to hornblende, needles of actinolite have grown in the chlorite and both may be replaced by hornblende.

The epidote that is disseminated through a rock, in contrast to that concentrated in clots and veins, presumably grew during the first stage of metamorphism. Its distribution is erratic. It most commonly replaces albite as small anhedral grains whose distribution does not reflect the zoning in anorthite content of the original labradorite; yet in many samples albite is devoid of epidote. It is next most common as replacement of chlorite, particularly in interstitial areas and in rocks in which there is little or no hornblende. It is least common in pyroxene. Some rocks thoroughly altered to chlorite and albite contain no epidote; others have epidote in chlorite, pyroxene, and albite; and in still others, epidote is restricted largely to just one of these minerals. In addition it fills vesicles in basaltic tuffs and lavas.

Biotite occurs sparingly in many of the altered rocks and in quantity in a very few. It forms small flakes with a pleochroism that ranges from sample to sample from a very pale green to a dark olive green maximum absorption. Some biotite, especially in the tuffaceous breccias, is pale to dark brown. It most commonly replaces chlorite, showing a gradation from typical, very pale green, almost isotropic chlorite to a slightly darker green pleochroic and moderately birefringent mica to typical olive green biotite. Almost all samples carrying biotite also have considerable calcite, but the reverse is not true. In general, biotite is most common in the altered basaltic tuffs of the Little River formation and in intrusive diabase rocks near contacts with the Little River formation, but there are many exceptions to this generalization.

Calcite is common in the more altered rocks, generally as irregular patches in albite, chlorite, or pyroxene. In a few of the coarse peg-

matitic clots it fills interstices between plagioclase crystals and has spread outward into the margins of adjacent plagioclase and pyroxene or hornblende crystals. Many vesicles in basalt flows and agglomerates are filled by calcite.

Although most quartz is clearly associated with the second epidotization stage, in a few samples it forms small irregular patches in albite or chlorite, particularly in the much altered rocks with much epidote. It also is a common vesicle filling.

Pyrite and pyrrhotite are widespread throughout the Cutler and Moose River quadrangles. Pyrite forms euhedral crystals in vesicles or in the matrix of volcanic and intrusive rocks. It commonly is most abundant along contacts between intrusive diabase and the Little River formation. Pyrrhotite occurs as small anhedral granular masses usually strung out parallel to foliation in the tuffaceous breccias.

The patchy distribution of alteration, its local variations in degree of intensity, the lack of uniformity in mineral assemblages even within short distances, and the presence of calcite and sulfides indicate that this first stage of metamorphism stemmed from hydrothermal activity (defined here as solutions in which the solvent is primarily H_2O regardless of whether in a liquid, gaseous, or supercritical state.) The presence of various combinations of albite, chlorite, quartz, epidote, calcite, biotite, and iron-sulfides in vesicles suggests that Na, Ca, Si, Al, Fe, S, K and CO_2 were in the hydrothermal solutions. Of these only H_2O, K, CO_2, S, and probably additional Na could not have come from the original labradorite, augite, chlorite, and magnetite in the parent magmatic rocks. Possible sources for these excess elements are the marine sedimentary rocks (including interstitial sea water) intruded by the diabase, or residual solutions from the hypothetical underlying basaltic magma chamber.

The origin of hornblende-bearing rocks locally in a hydrothermal environment which produced dominantly chloritic rocks is a puzzling problem. In regional thermal metamorphism, conversion of chlorite to hornblende is generally attributed to an increase in temperature in a system open to the escape of water or to a decrease in water vapor pressure in a closed system. The rocks of the Cutler area appear to have been metamorphosed in an open system allowing migration of hydrothermal solutions; and it is unlikely that strictly local areas of significantly higher temperatures could be maintained for long in the rocks. The alternative is that the presence of hornblende in primarily

chloritic rocks stems from local variations in the composition of the hydrothermal solutions. One of these variations appears to have been ferric iron, judging by the very erratic distribution of epidote. Solutions saturated in calcium with respect to hornblende and low in ferric iron might alter chlorite to hornblende rather than epidote and at the same time convert augite to hornblende merely by the addition of water without the leaching of calcium from augite that would lead to chlorite.

Second Stage

The second stage of metamorphism in the Cutler and Moose River quadrangle consisted of an influx of epidote in dikes, veinlets, ellipsoidal clots, and irregular patches. In general there was blanket replacement of all or almost all the older minerals by coarse granular epidote with no apparent regard for the composition of the replaced minerals. The epidote is commonly accompanied by considerable granular quartz.

Dikes. Narrow dikes and veinlets of epidote take two forms, replacement and fracture filling. In the replacement type there is a gradational boundary between the epidote-rich dikes and the surrounding rocks; the dike is rather variable in width and in concentration of epidote; and some pyroxene crystals and remnants of plagioclase can be seen. Thin sections of these diffuse dikes show that epidote has replaced other minerals – either pyroxene, amphibole, chlorite, or albite – to a varying but substantial degree, usually without preference for any one mineral (Fig. 22). In some of these dikes, previous minerals have been completely replaced, leaving the original textures still discernable in the pattern of the epidote.

The fracture-filling dikes have sharp boundaries and generally are more uniform in width and concentration of epidote, although they may be more irregular and branching than the replacement type. When seen as veinlets in thin section they appear to have followed cracks cutting sharply through pyroxene, chlorite, or plagioclase. In some slides epidote is confined to the veinlet, in others it is also scattered throughout the slide. Most veinlets consist of granular epidote plus a little quartz and calcite; but the mineral assemblage of the veinlet may differ depending on the mineral cut. Where the veinlet crosses albite, the vein minerals are epidote, calcite, clear albite and quartz; where it crosses pyroxene, chlorite and actinolite are the vein minerals.

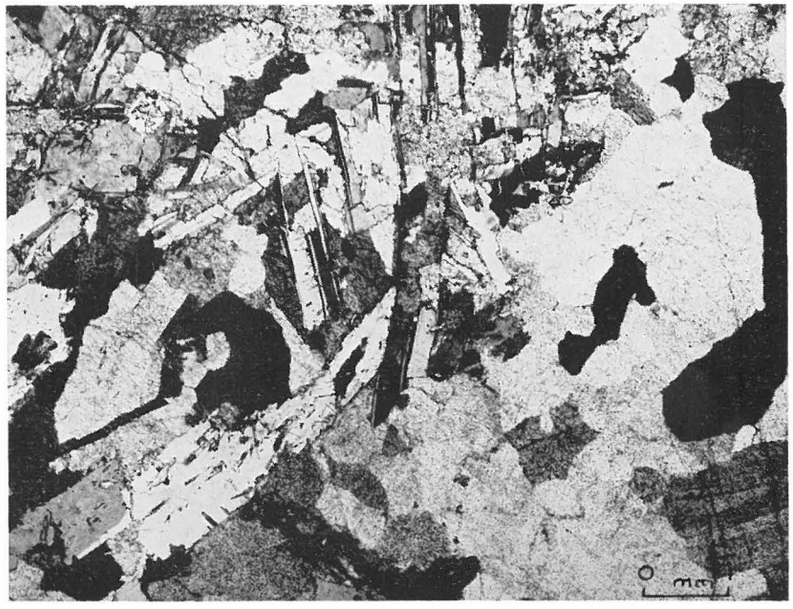

Fig. 22. Photomicrograph of epidote (lower right) replacing gabbro. Crossed nicols. Dennison Point.

Clots. Clots of epidote are common, particularly in the Cutler diabase and the basaltic lavas and agglomerates of the Little River formation, but they also occur in the keratophyres, and rarely in some of the breccias and tuffs of the Little River formation.

The clots tend to occur in swarms, commonly but not invariably scattered through a zone up to 20 or 30 feet wide along contacts (Fig. 23). Some of these clots have a coarse pegmatitic texture and except for their high epidote content are indistinguishable from the more normal pegmatitic clots already described. The pegmatitic clots apparently have acted as centers for epidote replacement. Other clots, however, either are of massive granular epidote or display typical gabbroic or diabase textures in which the epidote has completely replaced the plagioclase. Pyroxene may be partially altered to actinolite, chlorite, or epidote in some clots, but in others it is remarkably fresh. Some of these clots contain considerable granular quartz along with the epidote. The edges of these epidote clots are marked by a change from epidotized to unepidotized plagioclase and may be irregular on a scale of an inch or so, although these clots at a distance appear to be

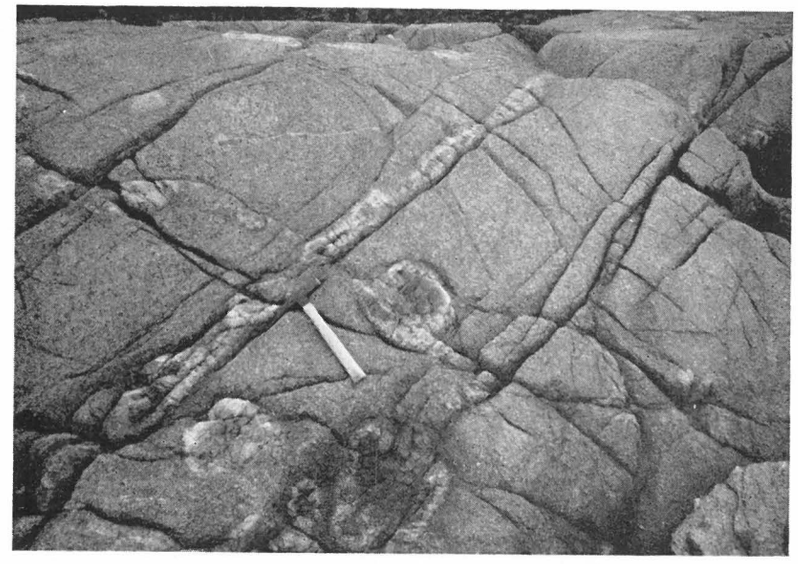

Fig. 23. Epidote clots and dikes in gabbro. Dennison Point.

rather sharply defined. The areas rich in clots generally have epidote dikes or veins, but the veins do not feed the clots.

In addition to dikes dominantly of epidote, there are widely scattered dikes of white bull quartz, commonly with a little epidote. These are much more plentiful in the acidic volcanic rocks of the Pembroke formation than in the basaltic volcanics of the Little River formation and of the Cutler diabase. Several of the major shear zones contain veins of calcite, and one intersecting the coast between Bog Brook Cove and Moose River has a mass of ankerite with quartz veins which must weigh at least several tons.

A puzzling problem is that there appears to be no relation between the degree of alteration and the porosity or permeability of the rocks as controlled either by grain size or the amount of shearing. Although some of the highly sheared rocks are much altered, there are many others equally sheared which are altered only to a minor degree; and conversely many unsheared rocks are highly altered. Dikelets and veinlets of epidote and calcite clearly follow cracks, both in outcrop and in thin section, but the diffuse dikes of epidote, the clots of epidote, and the pervasive alteration of many rocks do not correlate with visible paths for the introduction of hydrothermal solutions.

Age of Metamorphism

The metamorphism was approximately contemporaneous with the shearing and deformation of the rocks. Some of the chlorite and actinolite in the least altered diabases and gabbros may be late magmatic or deuteric, but to differentiate such alteration from the later complex metamorphism can not confidently be done. Some alteration clearly post-dates the shearing. Dikes and veins of epidote or calcite cut across the regional cataclastic foliation, and in thin-section veinlets of epidote and calcite cut microshears and fractured crystals. Vesicles which have been flattened parallel to the shearing are filled by unstrained chlorite, calcite, quartz, or epidote. Cracks in fractured crystals clearly guided the alteration in its early stages. On the other hand, many of the microshears seen in thin section contain distorted and strained chlorites or actinolite and in many rocks fracturing and granulation show no correlation with the degree or location of the alteration. Apparently metamorphism occurred before, during, and after deformation.

STRUCTURAL GEOLOGY

Tectonic deformation of the bedrocks in the Cutler and Moose River quadrangles has been intense. The most prominent and consistent structure is a northeast to east cataclastic shear foliation in both volcanic and intrusive rocks which in some places is sufficiently concentrated to constitute mappable shear zones. In addition there is a multitude of minor shear zones, small faults, local breccia zones, and joints that are visible along the shoreline but cannot be traced inland through the woods. A major fault mapped by Bastin and Williams in the Eastport quadrangle, the Lubec shear zone, presumably lies buried beneath the glacial debris of East Stream and Eastern Marsh Brook valleys.

The comparatively few outcrops of bedded rocks generally show moderate to steep dips; and small folds occur in several places along the shoreline and inland. However a system of consistent and numerous drag-folds is absent. The general map pattern of the diabase and the attitudes of the bedding in the scattered blocks of sedimentary rocks suggest that a large anticline and syncline may be present; but the lack of mappable stratigraphic units within the Little River formation makes the existence of these large folds tentative.

Cataclastic Shear

A foliation, apparently the result of cataclastic shearing, is common in almost all rock types of the Cutler and Moose River quadrangles. It occurs both concentrated in shear zones – a few of the most pronounced are shown on the map – and scattered with varying degrees of intensity throughout the two quadrangles. In the Pembroke formation it strikes north-northeast, parallel to bedding. In the Little River formation and Cutler diabase it strikes generally northeast with a range from about north to about east. Most of this foliation ranges in dip from 70 to 90 degrees, but locally it may have a dip as low as 45 degrees.

Shear Zones. The shear zones shown on the geologic map display closely spaced cleavage planes; flattened and commonly elongated inclusions, lava pillows, or boudons; and much slickensiding usually indicated by smeared-out chlorite and calcite. In some of these shear zones, coarse gabbro has been broken down to a mixture of chlorite and fractured or granulated plagioclase crystals, the original texture

completely destroyed. In many places, the zones of intense shearing lie along contacts of diabase or gabbro with other rocks, but they also may cut through intrusions.

The geologic map does not indicate displacement in a consistent direction along the shear zones, although the absence of displacement may be more apparent than real owing to lack of stratigraphic control. Slickensides in some shear zones suggest vertical movement and in others oblique movement with a plunge as low as 45 degrees. Very few slickensides indicating horizontal strike slip movement were found.

Foliation

The distribution of foliation symbols on the geologic map shows the widespread occurrences in all rock types of the cataclastic foliation in addition to the highly sheared zones. It is prevalent along contacts of gabbro or diabase, particularly where the contact trends approximately northeast. Probably at least one half the contacts of the Cutler diabase seen in the field were sheared to some degree. Shearing is less

Fig. 24. Cleaved argillite on the north shore of Moose River. Bedding dips south about 55 degrees and the cleavage dips south about 75 degrees.

prevalent within large intrusive gabbro bodies, but even there it may appear as a faint northeast-trending grain on well exposed outcrops.

The various rock types of the Little River formation and Cutler diabase responded to the deformation in several ways. The development of cataclastic foliation in gabbro can be clearly seen in thin-sections. In slightly deformed gabbro, fracturing is restricted to small crystals of plagioclase and pyroxene in areas rich in chlorite. The weak chlorite readily yielded, causing fracturing of the discrete small plagioclase and pyroxene crystals within it; but the large ophitic to subophitic patches of pyroxene and plagioclase with little chlorite were strong enough to remain untouched. As the amount of deformation increased, as traced from thin-section to thin-section, the chlorite patches of fractured crystals gradually coalesced and became elongated parallel to the shearing; and small cracks branched off from the main zones into the surviving masses of plagioclase and pyroxene. In some slides, the chlorite appears to have been forced into these cracks. The most completely brecciated rocks consist of broad, more or less parallel zones of brecciated crystals in distorted, sheared, and strained chlorite. These zones wrap around and separate from each other irregular spindle-shaped clots of highly fractured pyroxene and plagioclase which however still retain their original ophitic or subophitic texture.

In many of the basaltic or diabasic lava flows, amygdules are flattened parallel to the shearing; and foliation planes tend to wrap around inclusions or the rubble fragments in flow tops. Pillows in pillow lavas are likewise deformed; shear planes wrap around the pillows and weave in and out between them. In a number of places, cross-sections of columnar jointing, show the columns thinned at right angles to the foliation and lengthened parallel to it. Under the microscope, deformation of the matrix of the basaltic volcanics and the tuffaceous breccia show the same general features on a small scale. Rock fragments are flattened; amygdules, once spherical, are now ellipsoids; and lenses and streaks of chlorite along with fine fibrous actinolite and small fragments of pumice or of plagioclase crystals ramify through the rock, enclosing larger rock fragments or filling cracks within them, all oriented parallel to or at a small angle to the megascopic foliation. In a few of the most sheared tuffaceous breccias, even the large pumice lapilli or small bombs have been ground down to a mixture of angular rock fragments, plagioclase chips, epidote grains, strained chlorite shreds, and actinolite needles.

In the bedded argillaceous and tuffaceous rocks of the Little River formation, the spacing and the excellence of the cleavage correlate with the grain size of the rock. The argillites along the north shore of the Moose River have a true slatey cleavage, closely spaced and with the sheen of oriented mica flakes on the cleavage surface. In fine-grained siltstones and tuffs, the foliation is marked by a parallel alignment of minute mica flakes which wrap around small quartz or feldspar grains. In sandstones and medium-grained crystal tuffs, the cleavage derives from the appproximately parallel orientation of streaks and thin septa of chlorite or muscovite between crystal grains or tuff fragments. These streaks may join one another to give foliation planes at least as long as the slide.

The siliceous banded sediments of the Little River formation and the devitrified groundmass of the keratophyre porphyries locally show brecciated zones filled with chlorite or an alignment of scattered muscovite or chlorite shreds in micro-shears. In general, however, these rocks form blocky or irregularly fractured or jointed outcrops rather than cleaved ones.

In many thin-sections, particularly of gabbro, the shear zones are commonly branching and show no rotated crystal fragments, no common direction of displacement along fractures in crystals, and no en echelon pattern of fractures. On the other hand, some dikes show consistent offsets along the shears and some bedding planes in cleaved shales, siltstones, or fine tuffs have a definite and consistent shearing offset. Some rocks apparently sheared like a deck of cards. Others were merely flattened, extrusion at right angles to the flattening occurring along brecciation planes with no consistent shear displacement.

Lineation

The regional cataclastic shear foliation is accompanied in some places by a lineation in the shear planes. This lineation consists of slickensides, streaked-out chlorite blebs or feldspar crystals, and pillows, amygdules, epidote clots and pebbles stretched into triaxial ellipsoids. The lineation generally plunges steeply, commonly within 20 degrees of vertical, and the short axis of the ellipsoid is perpendicular to the plane of shearing. At a few places the lineation was found horizontal or nearly so; and locally amygdules, epidote clots, and pebbles form uniaxial ellipsoids rather than triaxial ones.

In addition to the northeast trending cataclastic foliation, the rocks of the Cutler and Moose River quadrangles have been broken up by innumerable joints, by small faults, and by shear zones with a wide range of orientations. They are unmappable away from the shoreline and their great number and complexity require a statistical examination not within the scope of this study.

Lubec Shear Zone

The strike of the Edmunds and Pembroke formations northwest of the central valley is divergent from that of the Little River formation and the Cutler diabase southeast of the valley. This discontinuity, buried beneath the glacial sediments of East Stream valley, is on strike of the Lubec fault zone in the Eastport quadrangle, which Bastin and Williams (1914, p. 13) describe as a zone of shearing several hundred feet wide with a border zone of less concentrated shearing several miles wide. They conclude that displacement on the fault was "certainly great" and was a combination of dip slip and strike slip, the block to the northwest moving down and to the northeast. When projected northeast, the Lubec fault zone coincides with the passage between Deer and Campobello Islands. The geologic map of Campobello Island (Alcock, 1948) shows that on several small islands in this passage there is a fault that brings Pre-Devonian rocks on the northwest against Devonian rocks on the southeast. This displacement is opposite to that proposed by Bastin and Williams. However, the only evidence that this post-Devonian fault on the islands and the Lubec fault zone are the same is that they line up with each other, and this is not compelling.

Minor Folds

Minor folds were found at several places along the shoreline and on French Ridge and the hills about a mile southwest of Moose River in the interior. They range in form from open to isoclinal (Fig. 25), but all have several common characteristics. Their axes are horizontal or plunge at low angles, generally less than 20 degrees; their axial planes parallel the steep to vertical regional cataclastic foliation; and many show a good lineation formed by stretched volcanic bombs, pumice lapilli, or pebbles oriented vertically or nearly so in the axial plane (cataclastic) foliation. In the fold on French Ridge, thin bedding laminations are consistently offset along the foliation vertically, suggesting a shear fold. Small drag-folds on limbs and down-dip slickensides on bedding planes were not found.

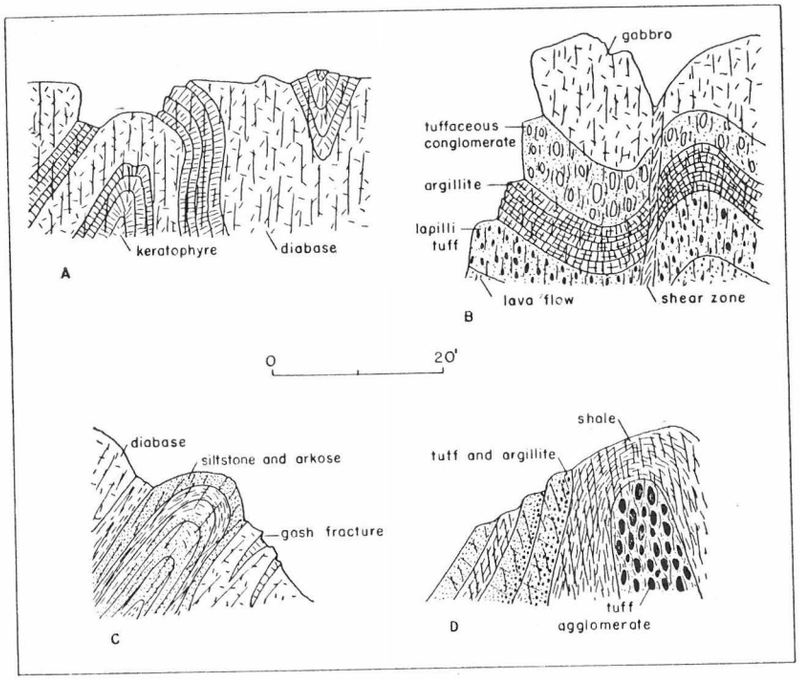

Fig. 25. Typical minor folds. Note that the regional cataclastic shear forms an axial plane cleavage in these folds and that rock fragments, bombs, and boulders are elongated in the plane of the shear cleavage. A — Sandy Point; B — Almore Cove; C — south shore of Moose River; D — north shore of Moose River.

In addition to well-defined minor folds, some blocks of bedded rocks show horizontal or gently warped beds whereas in others dips are steep to vertical. On Deer Island the bands in the banded gabbro dip south on the south side of the island and north on the north side; and along the coast from Long Point to Moose Cove, local masses of faintly banded gabbro dip as much as 30 degrees. With few exceptions, the strikes of tilted beds or bands range from northeast to east, parallel or at a narrow angle to the strike of the cataclastic foliation.

Major Folds

Southeast of the East Stream valley, rather meager evidence suggests that the Little River formation, keratophyre, and Cutler diabase have been regionally folded to form a large northeast trending anticline and syncline. The sedimentary rocks, tuffs, and lavas along the north shore of Moose River dip to the south and are right side up,

judging from cross-bedding, graded bedding, pipe amygdules, and the fact that the cataclastic shear — which appears to be axial plane cleavage — dips more steeply than bedding. These rocks could thus be on the north limb of a syncline. Around Cutler, the sedimentary rocks along the north shore of Almore Cove dip to the south, and the general dip, though variable, along the south shore of Little Machias peninsula is also to the south. Graded bedding in tuffs at these localities indicate that the beds are right side up. On the ridge west of Ackley Pond, bedded coarse basaltic tuff dips northwest. The only evidence that these beds are right side up is that the cataclastic shear cleavage dips more steeply than the bedding. Along the shoreline from Long Point to Bog Brook Cove, the dips of bedded rocks range from steep south to steep north, with a consistent northeasterly strike.

These data, plus the very general S-shaped pattern of the diabase, suggest the presence of a northeast trending syncline south of Moose River and a companion anticline near the head of Little River in which the axial planes parallel the regional cataclastic shear foliation and the major fold axes parallel the axes of the scattered minor folds. However, the scarcity of bedded rocks and the absence of mappable stratigraphic units in the Little River formation prevent confident delineation of these major folds. The alternative, judging from the prevalence of south dips, is that the rocks southeast of the central valley are on the north limb of a regional northeast to east trending syncline.

Northwest of the presumed Lubec fault zone, rocks of the Edmunds and Pembroke formations strike north-northeast, and dips of banding, laminations, bedding and basaltic flows or sills are vertical or nearly so. The few outcrops of shale tuff and argillite near Lively Brook reveal a northwest dip, but whether or not these beds are overturned is not known. Perhaps these rocks are more highly folded than the map indicates.

Origin of Deformation

Southeast of the central valley the northeast to east strike of the shear foliation, shear zones, and bedding plus the steep dips of foliation, shear zones, and axial planes of minor folds indicate that these structural features reflect the same tectonic deformation. The minor folds are primarily shear folds, and a system of numerous assymetrical drag-folds is absent. The low plunges of fold axes, and the steep lineation formed by distorted bombs, tuff fragments, and pebbles, both

in axial planes of the minor folds and in the rocks, suggest that the shearing developed in response to stress that produced a maximum compressive strain at right angles to the plane of foliation and a minimum compressive strain, (or extension) vertically in the foliation plane. The rocks thus appear to have been distorted by a compressive stress that acted horizontally with a general north northwest–south southwest orientation. The large, massive, competent gabbro intrusions channeled and locally distorted the shearing around their margins and undoubtedly prevented formation of more numerous minor folds in the bedded rocks.

The significance of the discordance in structural trends between the rocks northwest and those southeast of the Lubec shear zone – presumably present beneath the glacial debris of the central valley – is not known. Whether this discordance is due to differently oriented original stresses or is due to deformation of originally parallel trends is a question that requires mapping of adjacent quadrangles (now in progress) to answer.

The age of deformation is probably late Silurian or the lower half of the Devonian according to structural relations in the Eastport quadrangle. There Silurian volcanic rocks and shales are overlain along a pronounced angular unconformity by the Perry formation red conglomerates of Upper Devonian age. Bastin and Williams (1914, p. 13) conclude that the major deformation in the area occurred during this interval.

GEOLOGIC HISTORY

The record of geologic history in the Cutler and Moose River quadrangles began during middle Silurian time with explosive volcanic eruptions, at first largely basaltic to form the Little River formation, and later more acidic to form the Pembroke and Edmunds formations. Many of these eruptions were marine. During and after this volcanism there were intrusions and perhaps some extrusive flows and tuffs of glassy keratophyres. Repeated intrusions of basaltic magma then produced a complex of basaltic, diabasic, and gabbroic dikes, sills, and small plutons known collectively as the Cutler diabase, and some of the volcanic rocks may have foundered in the rising magma. The entire area was then deformed, cataclastically sheared, and permeated by hydrothermal solutions which changed the original pyroxene-plagioclase-titaniferous magnetite assemblage of the basic volcanics and intrusives to complex assemblages of albite, chlorite, epidote, actinolite, hornblende, biotite, calcite, quartz, sphene, and iron sulphides.

Post or late Silurian deformation, uplift, and erosion, and deposition of overlying Devonian red sandstones and conglomerates must be inferred from the geology of the neighboring Eastport quadrangle. If the Carboniferous sedimentation and the Appalachian folding and faulting of the Maritime provinces occurred in the Cutler area, all record of these events have been destroyed by Mesozoic and Tertiary erosion. By the time Pleistocene glaciers advanced over the area, it had been eroded to a land of rolling hills and subdued relief, possibly an upland bounding the Triassic basin of the Bay of Fundy. With retreat of the last glacier the sea advanced over most of the area, in fact there may have been several advances and withdrawals of both sea and ice. The sea finally withdrew to leave the Cutler and Moose River quadrangles essentially as they are today except for some later minor erosion by streams and the cutting back of glacial debris by waves to form local sea cliffs, sandy and gravelly beaches, and boulder-strewn tidal mud flats.

ECONOMIC GEOLOGY

Although sulfides are widely distributed throughout the Cutler and Moose River quadrangles, no zones of sufficiently high concentration to suggest possible ore-bearing veins were found. Several of the main shear zones, notably the one north of Stone Hill, are strongly iron-stained and contain considerable quartz, pyrite, and ankerite, but no metallic ore minerals. An old digging on Eastern Head, once a copper prospect 50 to 60 years ago according to local knowledge, consists of a quartz vein about 3 feet wide in a vertical shear zone cutting gabbro. The vein strikes north and can be traced by a notch in the topography and patches of white quartz float almost to Money Cove. The ore mineral is chalcopyrite associated with pyrite and siderite.

A survey by scintillometer of all rock specimens from the two quadrangles showed no radioactivity above the background count.

Fine-grained sheared basaltic tuffaceous breccia and some diabase were being quarried near the head of Little River during 1959 for use as cement aggregate and road-metal. In general, the diabase and gabbro in the area are too sheared and jointed for use as dimension stone.

The outwash sands and gravels of Pleistocene age have been extensively used for gravel and general fill, but for specialized use they require much washing and screening because of pockets of clay and a great range of sizes. Leavitt and Perkins (1934, I, p. 113, map 51) give results of sampling of 5 locations, most of them part of the Pond Ridge moraine. An estimated content of as much as 20 to 25 per cent of glassy fine-grained acidic porphyries limits the use of these glacial gravels for cement aggregate because of potash content.

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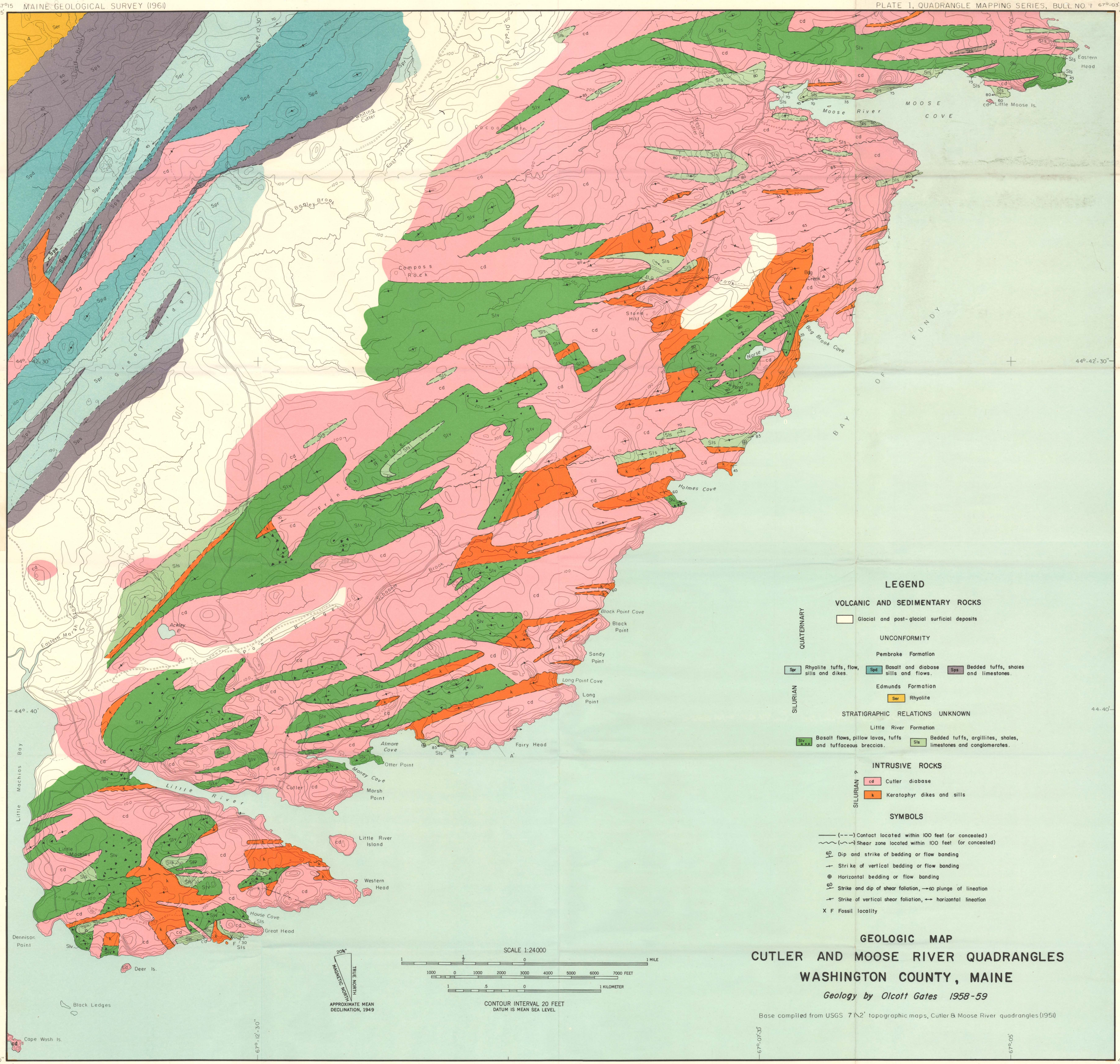
67°-15'
44°-45'
67°-03'-45"
44°-45'
67°-12'-30"
67°-10'
67°-07'-30"
67°-05'
44°-42'-30"
44°-40'
44°-37'-30"
67°-15'
67°-03'-45"
44°-37'-30"
Great Ridge
Pond Ridge
French Ridge
Little Machias Bay
Little River
MOOSE COVE
BAY OF FUNDY
Moose River
Eastern Head
Little Moose Is.
Cocoa Mtn.
Compass Rock
Stone Hill
Bog Brook
Bog Brook Cove
Holmes Cove
Black Point Cove
Black Point
Sandy Point
Long Point Cove
Long Point
Fairy Head
Almore Cove
Otter Point
Money Cove
Marsh Point
Cutler
Little Machias
Little River Island
Western Head
House Cove
Great Head
Dennison Point
Deer Is.
Black Ledges
Cape Wash Is.
Ackley P.
Eastern Marsh
Whiting Cutter
Easy Stream
Bagley Brook
Schooner Brook
Norse P.
Bog Creek
Ser
Sps
Spd
Spr
cd
Slv
Sls
k
A
A'
F
X
SCALE 1:24000
1 MILE
7000 FEET
1 KILOMETER
CONTOUR INTERVAL 20 FEET
DATUM IS MEAN SEA LEVEL
MAGNETIC NORTH
TRUE NORTH
20½°
APPROXIMATE MEAN DECLINATION, 1949
LEGEND
VOLCANIC AND SEDIMENTARY ROCKS
QUATERNARY
Glacial and post-glacial surficial deposits
UNCONFORMITY
SILURIAN
Pembroke Formation
Spr Rhyolite tuffs, flow, sills and dikes.
Spd Basalt and diabase sills and flows.
Sps Bedded tuffs, shales and limestones.
Edmunds Formation
Ser Rhyolite
STRATIGRAPHIC RELATIONS UNKNOWN
Little River Formation
Slv Basalt flows, pillow lavas, tuffs and tuffaceous breccias.
Sls Bedded tuffs, argillites, shales, limestones and conglomerates.
INTRUSIVE ROCKS
SILURIAN ?
cd Cutler diabase
k Keratophyr dikes and sills
SYMBOLS
(- - -) Contact located within 100 feet (or concealed)
(∿∿) Shear zone located within 100 feet (or concealed)
60 Dip and strike of bedding or flow banding
Strike of vertical bedding or flow banding
Horizontal bedding or flow banding
60 Strike and dip of shear foliation, →60 plunge of lineation
Strike of vertical shear foliation, ↔ horizontal lineation
X F Fossil locality
GEOLOGIC MAP
CUTLER AND MOOSE RIVER QUADRANGLES
WASHINGTON COUNTY, MAINE
Geology by Olcott Gates 1958-59
Base compiled from USGS 7½' topographic maps, Cutler & Moose River quadrangles (1951)
A
A'
Sea Level
2000'