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The Geology and Geophysics of the Passamaquoddy Bay Area, Maine and New Brunswick, and their Bearing on Local Subsidence

Olcott Gates
P.O. Box 234
Wiscasset, Maine 04578

ABSTRACT

The basement rocks of the Passamaquoddy Bay area are Precambrian volcanics and Lower Ordovician shales, perhaps part of the Avalonian terrane. They are overlain by a thick section of Silurian-Lower Devonian marine volcanic rocks which are bimodal and accumulated in an extensional, subsiding tectonic regime. During the late Early Devonian Acadian orogeny, the volcanic pile was folded and faulted and intruded by gabbro, diorite, and granite. In the Late Devonian, coarse alluvial red-bed conglomerates and arkoses derived from erosion of the underlying volcanic pile and plutonic rocks accumulated in a fault-bounded basin apparently largely confined to the Passamaquoddy Bay area. Carboniferous faulting divided the rocks into separate fault blocks of differing stratigraphy and structural complexity. Intrusion of Lower Jurassic diabase dikes and development of the Fundian fault along the western margin of the Triassic-Jurassic basin beneath the Bay of Fundy marks the end of the bedrock record. The bedrock geologic record suggests that the Passamaquoddy Bay area has been one of crustal weakness prone to subsidence and faulting at least as far back as the Silurian. A concentration of historical seismicity and an abrupt change in magnetic patterns suggesting a crustal discontinuity further imply a weak underlying crust.

In coastal Maine, postglacial rebound between about 13,000 and 9,000 yr BP was about 100 to 120 meters, ending with a stillstand at -65 meters. Since then the coast has submerged, rapidly at first but then slowing to a rate of 1.44 mm/yr by 5,000 yr BP. In Passamaquoddy Bay the differential downwarp of about 12 meters relative to the coast farther west, as shown by the reentrant of glaciomarine delta elevation contours, was largely completed by 5,000 yr BP. Limited archaeological and radiocarbon data suggest that from then to perhaps historical times subsidence was not much greater than that to the west along the Maine coast. Currently, subsidence appears to be faster than to the west, but the data from different sources give conflicting rates.

Most of the 12 meter differential subsidence of the Passamaquoddy Bay area was the response of a weak crust to the large stresses resulting from the large and rapid postglacial warping ending some time prior to 5,000 yr BP, augmented perhaps by isostatic disequilibriums due to the contrasting densities of neighboring rock bodies and to hydroisostatic loading of the crust beneath the Bay of Fundy as the sea rapidly deepened following the low stillstand. Now that rebound warping has ceased, crustal stresses derived from plate motion remain the only identified cause of the current downwarping and seismicity besides the unknown gravitational stresses.

INTRODUCTION

This paper summarizes the bedrock geology, geophysics, and geologic history of the Passamaquoddy Bay area, including postglacial rebound, as basic data with which to construct a model explaining the history of anomalous warping in that region. I have drawn on my detailed and reconnaissance mapping and that of Canadian geologists for the bedrock geology.

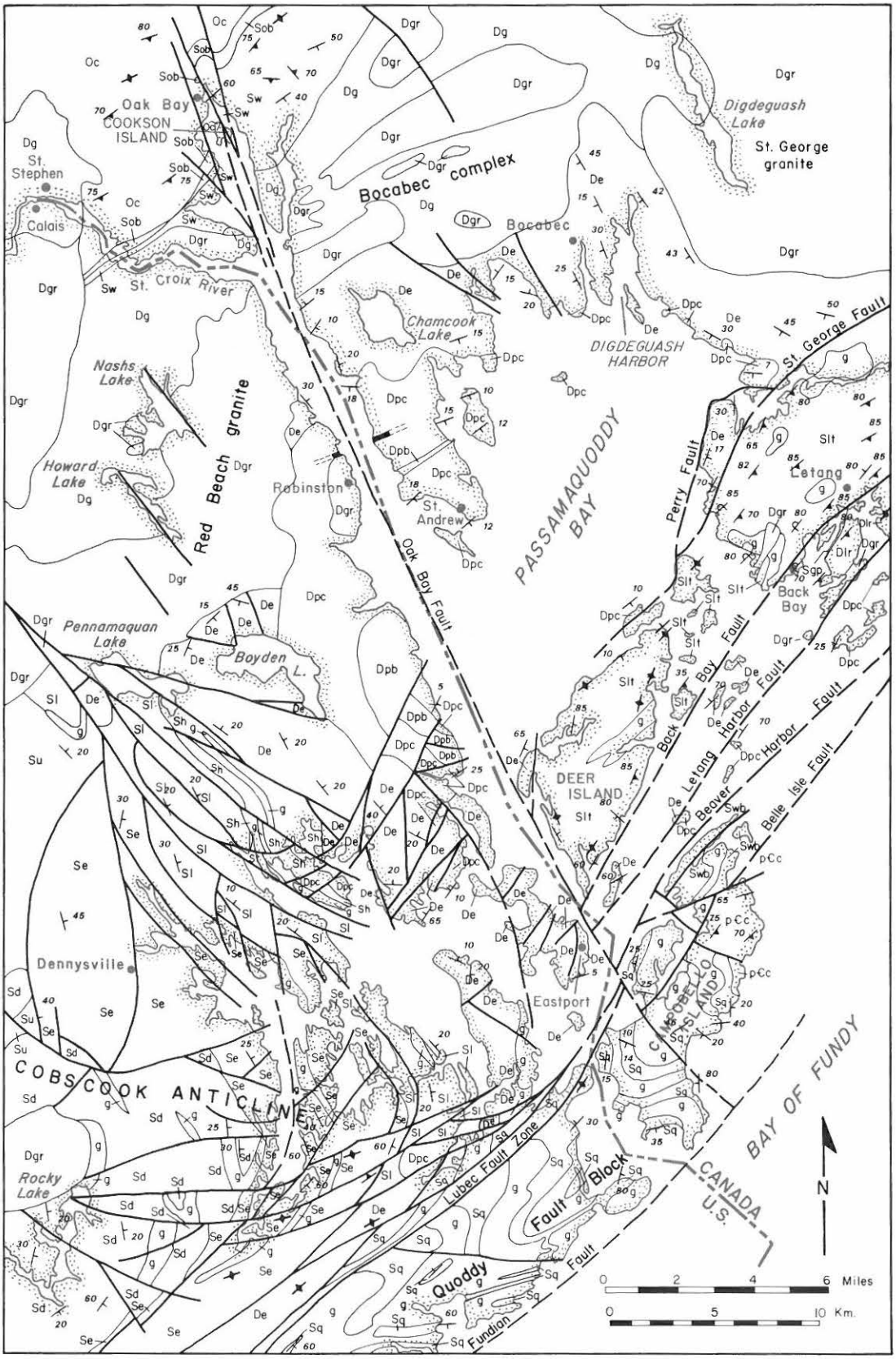

Figure 1. Geologic map of the Passamaquoddy Bay area, Maine and New Brunswick. (Sources: Amos 1963; Ruitenburg 1967, 1968; Gates 1975; Pajari 1976; McLeod 1976; Donohoe 1978; Ludman 1978)

Most of the information regarding postglacial warping and seismicity has been gleaned from companion papers in this volume, to which the reader is referred for additional data and references to original sources.

STRATIGRAPHY

Stratified rocks of Precambrian, Ordovician, Silurian, and Devonian age are exposed in the Passamaquoddy Bay area (Figure 1). They can be divided on the basis of major hiatuses into three groups: pre-Silurian, Silurian and Lower Devonian, and Upper Devonian. Figure 2 summarizes the pre-Upper Devonian stratigraphy.

PRE-SILURIAN STRATIGRAPHY

Precambrian

The Coldbrook Formation crops out in the northeastern part of Campobello Island (McLeod, 1976). The Belle Isle fault separates the Coldbrook Formation from the Lower Silurian Quoddy formation except for one small area (see Figure 1) where McLeod (1976) reports that the Quoddy Formation unconformably overlies the Coldbrook. The Coldbrook Formation continues northeastward along the south side of the Belle Isle fault to the Saint John, New Brunswick area, where it is unconformably overlain by fossiliferous Cambrian rocks.

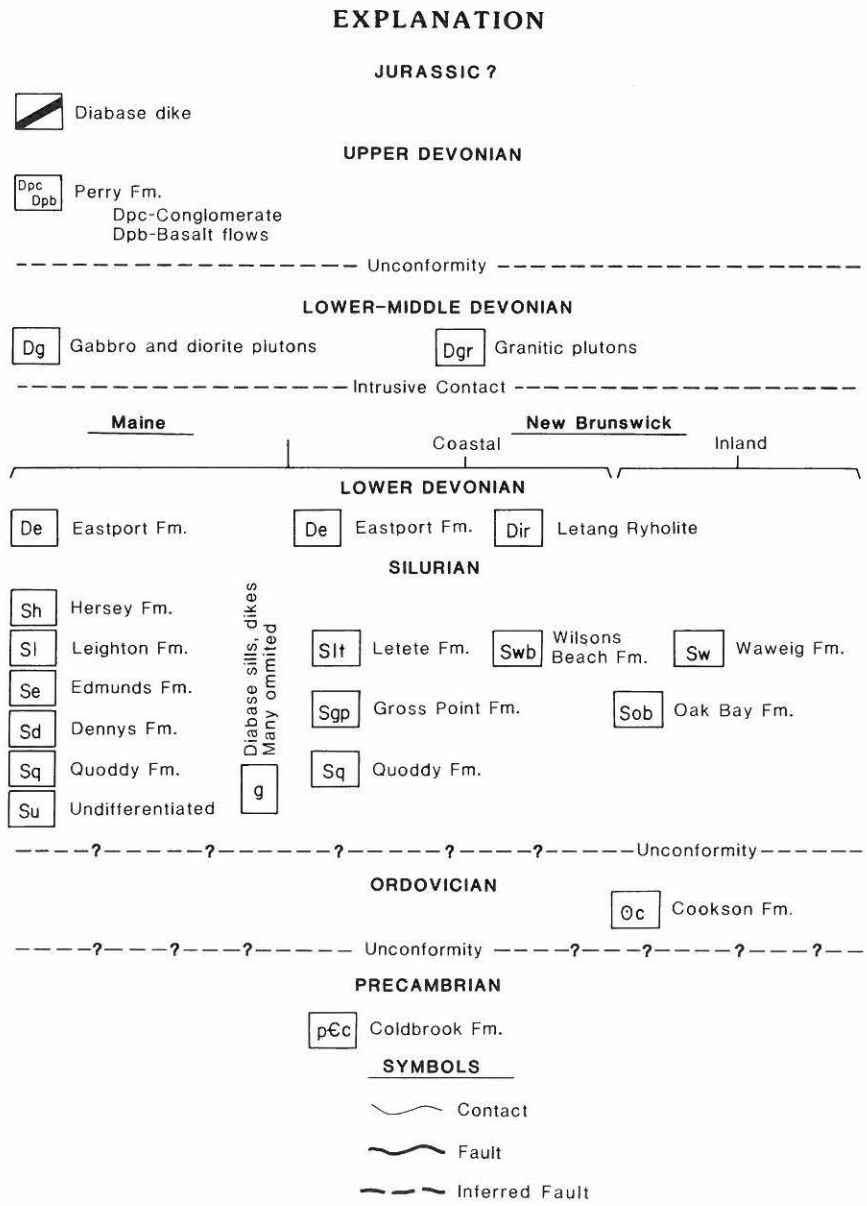

Figure 1. Continued.

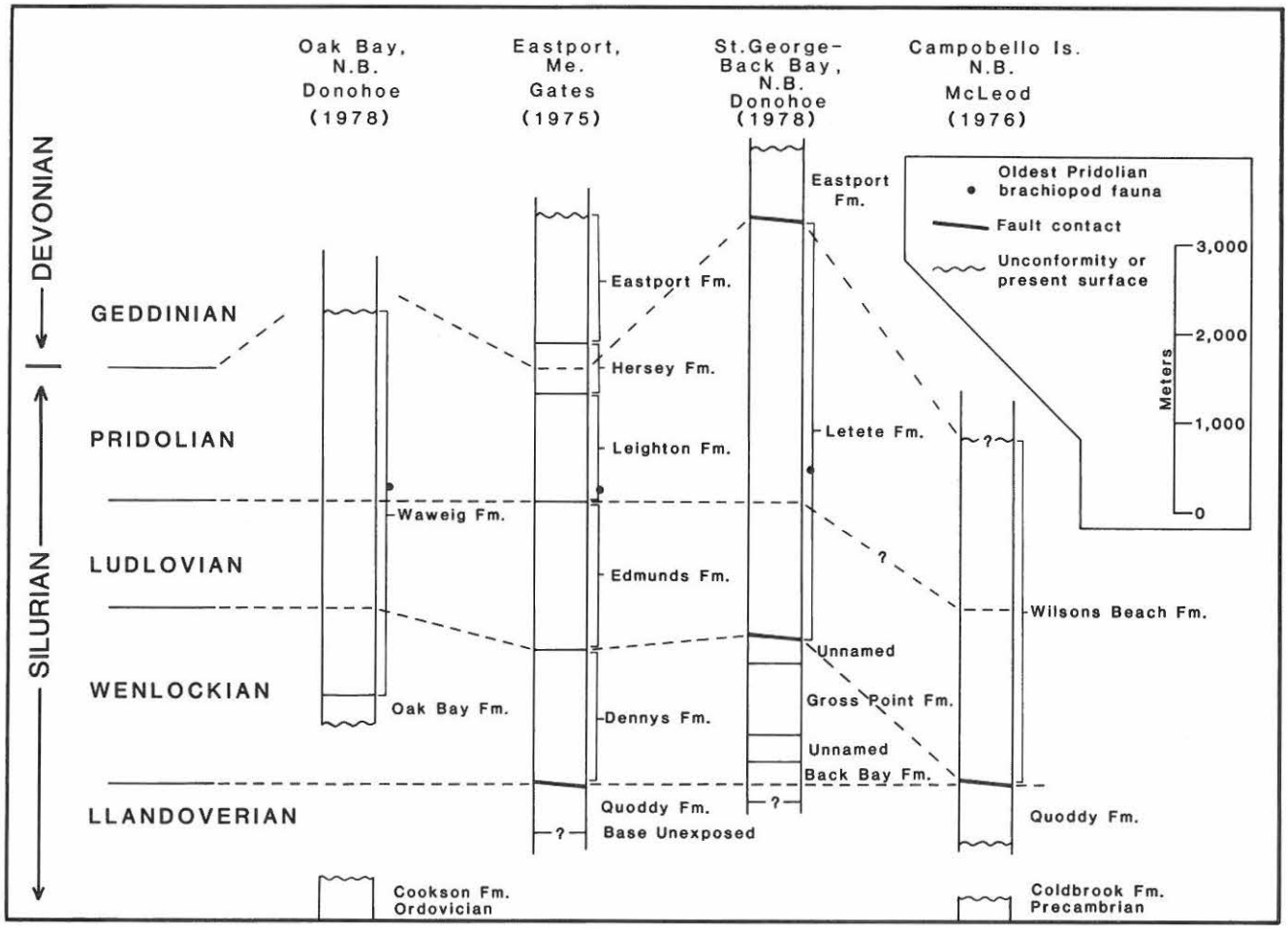

Figure 2. Correlation of the Silurian-Lower Devonian stratified rocks in the Passamaquoddy Bay area. (Adapted from Gates (1975), McLeod (1976), and Donohoe (1978)).

The Coldbrook Formation consists of completely recrystallized quartz-feldspar-biotite volcanic rocks and thinly-bedded fine-grained tuffaceous sediments (McLeod, 1976). The unit has undergone two episodes of metamorphism and three phases of deformation of probable pre-Silurian age.

Ordovician

The Cookson Formation occurs in the Calais-St. Stephen area north of Passamaquoddy Bay, and it or its equivalent extends westward to at least Penobscot Bay. Graptolites found on Cookson Island in Oak Bay establish a Lower Ordovician age. The base of the formation is not exposed, but on Cookson Island it is overlain in a complex fault zone by the Silurian Oak Bay conglomerate (Ruitenberg, 1967). It is composed of black slate, phyllite, quartzite, and minor basalt flows (Ruitenberg, 1967; Ludman, 1978). Regionally it is in the greenschist metamorphic facies, but in contact metamorphic zones around Acadian plutons the rocks are converted to staurolite-andalusite schists.

SILURIAN-LOWER DEVONIAN STRATIGRAPHY

The Silurian-Lower Devonian marine volcanic sequence is exposed on both sides of Passamaquoddy Bay in Maine and New Brunswick. It is part of the Coastal Volcanic Belt (Boucot, 1968) which also crops out to the west in the Mt. Desert Island area, Penobscot Bay, and near Newbury, Massachusetts. Bastin and Williams (1914) initially mapped the Eastport Quadrangle and named the formations there. The following brief descriptions of the various formations are organized according to age from oldest to youngest. Figure 2 illustrates the probable correlations.

Lower Silurian

The Quoddy Formation underlies the Quoddy fault block south of the Lubec fault zone in Maine (Gates, 1975) and south of the Beaver Harbor fault on Campobello Island (McLeod, 1976). Graptolites establish an Upper Llandovery age (Gates,

1975). Neither the top nor the bottom of the formation is exposed in Maine, but McLeod (1976) reports that the Quoddy Formation unconformably overlies the Coldbrook Formation on Campobello. In Maine, the Quoddy Formation is composed dominantly of pyrite-bearing and rusty-weathering black siltstone and siliceous argillite with feldspathic ash beds. Volcanic rocks, including basaltic and rhyolitic flows and tuff-breccias, become predominant to the southwest and upwards in the section. On Campobello Island turbidite beds are interlayered with volcanic rocks (McLeod, 1976). Rocks similar to those of the Quoddy Formation ("Su" in Figure 1) occur inland north of Dennysville, Maine, but are too poorly exposed to make lithologic correlation certain.

Middle Silurian

The Dennys and Edmunds Formations in Maine crop out in the large Cobscook anticline north of the Lubec fault zone. A diverse marine fauna of brachiopods, gastropods, trilobites, and corals suggest Wenlockian and Ludlovian ages respectively (Berry and Boucot, 1970; Gates, 1975). The two formations consist of a varied suite of marine volcanic rocks including basalt flows, rhyolite flows and domes, coarse tuff-breccias, and bedded tuffs together with interbedded fossiliferous shales and siltstones. The proportion of basaltic rocks is higher in the Dennys Formation than in the Edmunds.

Between the Back Bay and Letang Harbor faults in New Brunswick, Donohoe has mapped four units, the Back Bay and Gross Point Formation and two unnamed volcanic units, all of which have been combined under the Gross Point Formation symbol on Figure 1 because of the small map scale. They contain volcanic rocks, both flows and tuffs, interbedded with shales, siltstones, and limestone. A very sparse brachiopod fauna suggests a Lower to Middle Silurian age (Donohoe, 1978). Compared to the Quoddy and Dennys Formations, with which Donohoe correlates them, the four units have less volcanic material and more detrital and carbonate rocks. It is probable that the New Brunswick section is the distal equivalent of the Quoddy and Dennys Formations, deposited farther from the volcanic source region.

Upper Silurian

The rocks assigned to the Upper Silurian on Figure 1 have a distinctive shallow-water brachiopod Salopina fauna that Berry and Boucot (1970) assign to the Pridoli stage. The first appearance of that fauna is shown on Figure 2. In Maine the Leighton Formation rests conformably on the Edmunds in the Cobscook anticline and is overlain conformably by the Hersey Formation. The Leighton Formation is composed of gray shales and siltstones with intercalated basalt flows and tuff-breccias, and rhyolitic flows and tuff-breccias, all deposited in shallow water. The Hersey Formation consists of maroon siltstones and shales, locally calcareous, and has a restricted gastropod-pelecypod-ostracode fauna. It probably was deposited in tidal flats and lagoons. Ostracodes suggest that the Silurian-Devonian boundary lies within the Hersey Formation (Berdan, 1971).

In New Brunswick, on the peninsula south of St. George, and on Deer Island between the St. George and Back Bay faults, Donohoe (1978) has assigned a Pridoli age to the Letete Formation on the basis of a small brachiopod collection. The formation is fault-bounded without stratigraphic bottom or top. It consists of gray siltstones and shales, some local red siltstones, volcaniclastic conglomerates, quartz and feldspathic wackes, and flows and tuffs (Donohoe, 1978).

The Wilsons Beach Formation on Campobello Island, between the Belle Isle and Beaver Harbor faults, lithologically and structurally resembles the Letete Formation and has been assigned an Upper Silurian age (McLeod, 1976).

Middle to Upper Silurian rocks crop out in the Oak Bay, New Brunswick, area north of and intruded by the belt of Acadian plutons. On Cookson Island in Oak Bay, the Oak Bay Formation, a polymict conglomerate containing round pebbles and cobbles, overlies the Ordovician Cookson Formation along a complexly faulted, difficult to find, unconformity and is overlain conformably by the Waweig Formation. The Waweig Formation has Middle to Upper Silurian brachiopods in its lower part and the Salopina brachiopod fauna in its middle and upper parts (Pickerill, 1976), and thus is partly equivalent to the Leighton, Letete, and Wilsons Beach Formations. It consists largely of feldspathic and lithic graywackes, siltstones, and shale with some volcaniclastic units near its base. Near contacts with the intrusive rocks of the plutonic belt it is metamorphosed to mica schists with minor cordierite, andalusite and staurolite (Ruitenberg, 1967).

Lower Devonian

In Maine, the Eastport Formation is exposed in a fault block in the Lubec fault zone and in the east limb of the Cobscook anticline. It is cut off to the north of Boyden Lake and up the St. Croix River by the Devonian plutonic belt. In New Brunswick, the Eastport Formation makes up several islands between the Back Bay and Letang Harbor faults. It also crops out north of the St. George fault and along the north shore of Passamaquoddy Bay northward to the plutonic belt (Pickerill and Pajarie, 1976). Its ostracode fauna suggests a Geddinian age (Berdan, 1971). Throughout the Passamaquoddy Bay area, the Eastport Formation consists of flows and tuff-breccias of basaltic andesite; shallow intrusions, flows, domes, and ash-flows of rhyolitic composition; and variegated gray, green, and maroon siltstones, shales, and pebble conglomerates. It carries a brackish water assemblage of ostracodes, pelecypods, gastropods, and linguloids. Some of the volcanic rocks are subaerial, and the

fossiliferous sedimentary rocks were deposited in tidal flats and shallow lagoons.

The Silurian-Lower Devonian volcanic rocks of the Passamaquoddy Bay area are perhaps as much as 7 kilometers thick. They thin rapidly to the east and west. The volcanic components are dominantly of basaltic and rhyolitic mineralogy and chemical composition, with very few andesitic rocks. Gates and Moench (1981) have proposed an extensional tectonic regime for these bimodal volcanic rocks, perhaps one of normal faulting and subsidence permitting the accumulation of a locally thick section (Figure 3). In general, the lithologies and fauna of successive formations from the Quoddy to the Eastport indicate a gradual shallowing of the marine waters in which the rocks were deposited as the volcanic pile thickened, ending with subaerial and brackish water conditions in the Early Devonian.

Upper Devonian

The Perry Formation is exposed around the margins of Passamaquoddy Bay and probably underlies most of it. Plant fossils indicate an Upper Devonian age. It rests along an angular unconformity on the Silurian-Lower Devonian volcanic section and on the Acadian plutonic rocks which intrude the latter. It consists of maroon alluvial conglomerates, arkosic sandstone, red shales, and lacustrine beds (Schluger, 1973). It also contains two basalt flows. The Perry Formation is a post-tectonic molasse deposited in a fault basin during uplift following the Acadian orogeny.

INTRUSIVE ROCKS

On the geologic map (Figure 1) intrusive igneous rocks are divided into two groups. Within the Silurian-Lower Devonian sequence are numerous dikes, sills, and small irregular plutons consisting largely of diabase, including the diabase that intrudes the Quoddy Formation south of the Lubec fault zone. These intrusions have chemical and mineralogical compositions similar to those of the volcanic formations they intrude and are thus believed to be about the same age as their host formations.

A flow-banded bulbous rhyolite, the Letang rhyolite, intrudes the Gross Point Formation between the Back Bay and Letang Harbor faults and has been assigned a Lower Devonian age by Donohoe (1978).

The belt of large plutons of gabbroic to granitic rocks inland across the head of Passamaquoddy Bay intrude the Silurian-Lower Devonian stratified rocks and represent perhaps in part hypabyssal phases of the volcanism as well as the plutonic phase of the Acadian orogeny (Amos, 1963; Abbott, 1986). The older mafic magmatism included some layered plutons, gabbro, diorite, and associated granodiorite together with a few comingled diabase-granite dikes and mixed gabbro-granite breccias along plutonic contacts (Abbott, 1986). The younger and

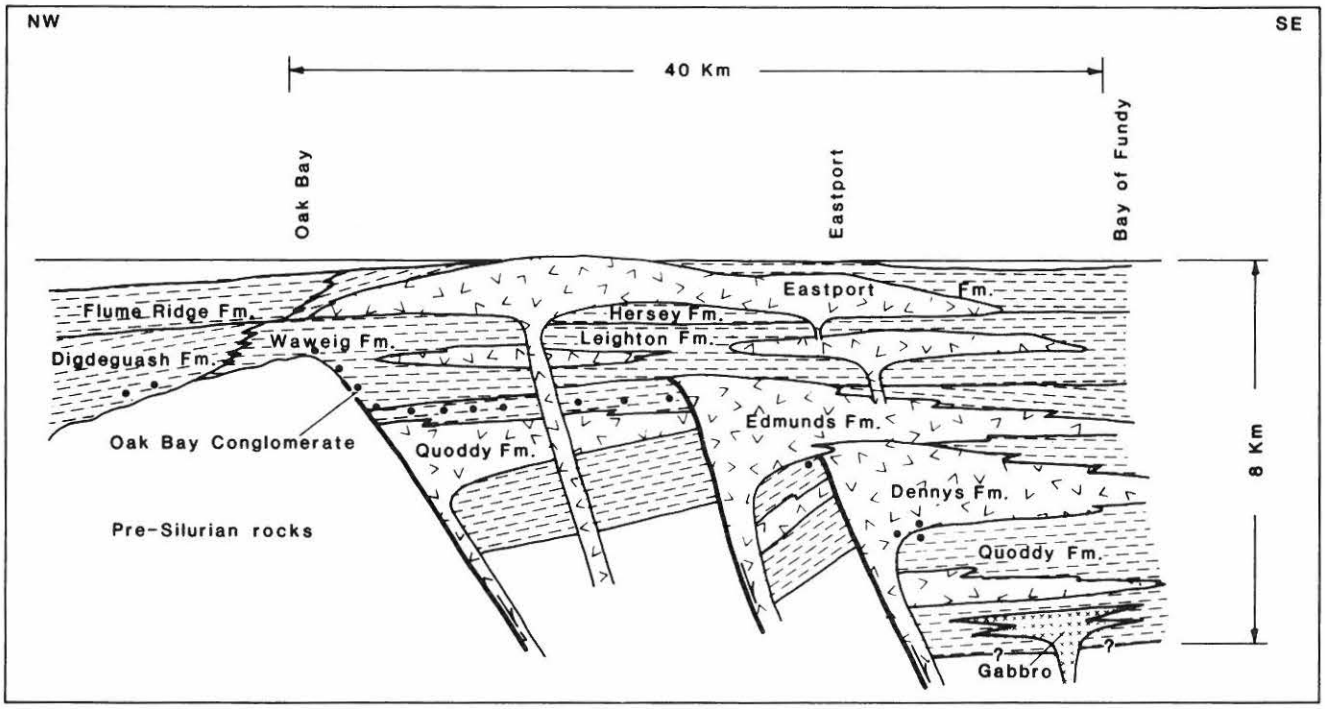

Figure 3. Schematic cross-section of the Eastport region illustrating the relationship of Silurian faults to the volcanic stratigraphy. (Gates and Moench, 1981)

probably overlapping granitic rocks consist of a varied suite of hornblende granophyre and biotite granite (Abbott, 1986).

In New Brunswick, the corresponding mafic rocks and later felsic intrusions have been named the Bocabec complex (Pajarie, 1976). The mafic rocks consist largely of coarse ophitic gabbro plus a few local lenses of layered gabbro with cumulate textures. The felsic intrusions include hornblende and biotite granodiorite, rapakivi granite, and red biotite granite (Pajarie, 1976). Hybrid intermediate rocks along some of the gabbro-granitic rock contacts reflect chemical and mineralogical reactions between the intrusive granitic rocks and the host gabbro. The Bocabec complex is intruded by the St. George pluton which extends eastward from Digdeguash Lake on Figure 1. It is a coarse-grained granite with phases of porphyritic granite, megacrystic granite, and aplite (Pajarie, 1976).

Compilations of isotopic dates by Donohoe and Pajarie (1973) and Pajarie (1976) give ages of 393 ± 6 Ma for the Red Beach granite near Red Beach in Maine and 396 ± 20 Ma for the felsic rocks of the Bocabec complex. The St. George pluton, however, may be of Late Devonian to Early Mississippian age (Pajarie, 1976).

The youngest intrusion in the Passamaquoddy Bay area is a diabase dike that strikes east-northeast, crops out along the shore on both sides of the St. Croix River north of Robbinston and St. Andrews, and has yielded a K-Ar age of 189 ± 8 Ma (Early Jurassic) (Stringer and Burke, 1985, p. 8).

STRUCTURE

Faulting

Five episodes of faulting -- Silurian, Middle Devonian (Acadian), Late Devonian, Carboniferous, and Triassic -- have been identified in the Eastport 2-degree quadrangle (Gates, 1978, 1982), although not every fault has sufficient criteria for establishing its age. On the geologic map (Figure 1), the arcuate north to northeast-trending fault north and west of Dennysville, which separates the Edmunds Formation from the Silurian undifferentiated (Su) unit, may be contemporaneous or nearly so with deposition of those units.

Faults that cut Acadian folds but in turn are cut by Acadian plutons, such as those that form a fan-like pattern cutting the Cobscook anticline, are assigned a late Early Devonian age. Some of the northeast-striking faults that drop the Perry Formation down against the Eastport Formation are bordered by wedges of coarse boulder conglomerate extending into the Perry Formation, the clasts of which are of the rocks on the upthrown side. Such faults probably formed during development of the Perry fault basin during the Late Devonian.

Faults that displace Acadian plutons or cut off the Lower Devonian Perry faults are probably of Carboniferous age. The most conspicuous Carboniferous faults are the vertical or steeply dipping, northeast-striking ones such as the St. George, Back Bay, Letang Harbor, Beaver Harbor, and Belle Isle faults. They displace the Perry Formation, cut Acadian plutons to the northeast in New Brunswick, or bound Carboniferous fault basins in New Brunswick (Potter and others, 1979). Lack of post-Carboniferous rocks along the faults prevents determination of post-Carboniferous fault movement, although the parallelism of Triassic faults bordering the Bay of Fundy suggests the possibility of such movement on the Carboniferous faults. Detailed study of small structures along some of the faults in the Passamaquoddy Bay area suggests that the latest movement was dip-slip (Donohoe, 1978).

The Lubec fault belongs to this northeast-striking regional Carboniferous fault system, but its correlation with companion faults in New Brunswick is uncertain. Donohoe (1978) favors correlation with the St. George fault offset along the Oak Bay fault. On the other hand, correlation with the Beaver Harbor or Belle Isle fault maintains continuity of the block of Quoddy Formation south of the fault.

The strike of the Oak Bay fault is normal to the regional structural grain and the regional Carboniferous faults. The Oak Bay fault appears to have determined the trend of the St. Croix River and the west side of Passamaquoddy Bay. It is exposed only on Cookson Island and the adjacent shores of Oak Bay, where it is a zone of sub-parallel faults nearly a kilometer wide. It is likely that the fault's continuation beneath Passamaquoddy Bay is also a fault zone rather than the single fault schematically mapped. The fault zone offsets the Oak Bay Formation, intrusive contacts of the plutonic belt, a lava flow in the Perry Formation, and the Lubec fault and its probable extension, the Beaver Harbor fault. The Letete Formation on Deer Island terminates at the fault zone and has not been identified to the west. Map-view displacement of the fault zone is left lateral, but a descriptive geometric construction suggests a large vertical component of net slip. If the average dip of 60° on the base of the Oak Bay Formation on each side of the fault zone is extended downwards in the plane of the fault zone to the intersection with the contact between the Waweig Formation and the plutonic belt (a contact that appears to be vertical according to its topographic trace), the net slip can be determined. This construction gives a net slip plunging about 45° to the north-northwest in the vertical fault plane, with a horizontal slip component of about 2 km and a vertical component of about 2 km, east side down.

Most of the displacement on the Oak Bay fault zone occurred before emplacement of the Lower Jurassic dike and is thus of Carboniferous to Early Jurassic age. Seismic profiles between Eastport and Deer Island across the strike of the Oak Bay fault show well-bedded unfaulted sediments, indicating no postglacial displacement (Kelley, 1986, personal communications).

The Fundian fault (Johnson, 1925) parallels the shoreline beneath the Bay of Fundy. It forms the northwest margin of the Late Triassic to Early Jurassic fault basin that underlies the Bay of Fundy (Ballard and Uchupi, 1975) and is the youngest fault identified in the Passamaquoddy Bay area.

Folding

The regional Carboniferous faults just described divide the stratified rocks of the Passamaquoddy Bay area into several fault blocks each containing its own style of folding and in some cases distinctive stratigraphy (Mcleod, 1976). In the following description of folding, the designations F_1, F_2, etc. apply only to the sequence of folding within a fault block, not regionally.

North of the plutonic belt and inland from the area shown in Figure 1, the Ordovician Cookson and the Silurian Waweig Formations have upright, east-northeast striking, gently plunging F_1 folds with a slaty axial-plane cleavage. These in turn have been folded by recumbent F_2 folds that have gently plunging hinge lines and gently dipping axial-plane cleavage (Ruitenberg, 1968; Ruitenberg and Ludman, 1978). Steeply plunging S-Z shaped F_3 folds occur near faults and contacts with granitic plutons. F_4 folds are kink bands. F_1 and F_2 folds are attributed to the Acadian orogeny; F_3 and F_4 folds may be Devonian to Carboniferous (Ruitenberg and Ludman, 1978). Pre-Silurian deformation of the Cookson Formation has been identified on Cookson Island and the shores of Oak Bay (Stringer and Burke, 1985).

South of the plutonic belt and west of the Oak Bay fault, the Cobscook anticline, an F_1 structure containing the Dennys, Edmunds, Leighton, Hersey, and Eastport Formations, is the major structure (Gates, 1975). It is a broad open fold plunging to the east. Dips on the south limb near the Lubec fault zone are vertical to overturned, and a very strong cleavage accompanied by isoclinal F_2 folds with gently plunging hinge lines parallels the Lubec fault zone and occurs within it. The faulting appears to have steepened and sheared out the south limb, presumably during the Carboniferous. The east limb dips from 20° to 60° east-northeast and strikes into the plutonic belt which intrudes it. F_2 folds and accompanying local cleavage occur along some of the faults that cut the anticline. A regional northeast-striking and steeply dipping cleavage cuts obliquely across the Cobscook anticline and parallels the Lubec fault zone, suggesting association with the Carboniferous regional northeast-trending faults.

The east limb of the Cobscook anticline strikes almost at a right angle to the east-northeast strike of the folding in the Cookson and Waweig Formations north of the plutonic belt. There is also a marked contrast in the style and complexity of the folding north and south of the plutonic belt. Perhaps a major regional fault separating two regional structural styles guided the ascending magmas of the plutonic belt.

A single open folding, like that of the Cobscook anticline and of the Eastport Formation of Passamaquoddy Bay, also occurs in the block of Quoddy Formation south of the Lubec fault zone and the Beaver Harbor fault of Campobello Island (Gates, 1975; McLeod, 1976), and is accompanied locally by an axial-plane cleavage and by minor folds and strong shearing in the Lubec fault zone.

In contrast to the relatively simple open folding of the syncline of Passamaquoddy Bay, the Cobscook anticline, and the Quoddy fault block, the rocks of the Letete Formation in the St. George-Back Bay fault block and of the Gross Point Formation between the Back Bay and Letang Harbor faults have a history of multiple deformation. Ruitenberg (1968), McLeod (1976), and Donohoe (1978) have all commented on this contrast. Donohoe (1978) has mapped upright isoclinal F_1 folds plunging steeply to the northeast with an accompanying steeply dipping axial plane cleavage marked by metamorphic chlorite, muscovite, and biotite and a northeast plunging lineation of elongated clasts and granulated minerals. F_2 folds are also upright with hinge lines plunging northeast to southwest and a local crenulation cleavage. F_3 and F_4 are successive episodes of kink-banding. Donohoe attributes all four phases of folding to the Acadian orogeny and has not identified any penetrative deformation associated with the major Carboniferous faults that bound the blocks. No correlation between F_1 and F_2 folding of the Cookson and Waweig Formations and those of the Letete and Gross Point Formations can be made because the configurations and plunges of the fold systems are not the same.

The Wilsons Beach Formation in the Beaver Harbor-Belle Isle Fault block forms an overturned gently plunging F1 syncline with minor tight folds on the limbs and an axial-plane cleavage marked by metamorphic muscovite and granulated minerals (McLeod, 1976). This was followed by an F_2 crenulation of the cleavage and an episode of kink-banding.

These contrasts in structure from fault block to fault block and the confinement of certain stratigraphic units to certain fault blocks suggest that the fundamental regional structure and stratigraphy of the Passamaquoddy Bay area has not yet been determined. Much more work clearly needs to be done. Meanwhile, I hazard the very tentative working hypothesis that the Passamaquoddy Bay area is underlain by two plates of a thrust (Acadian?), one above the other, each with its own stratigraphy, one with a relatively simple Acadian deformation plan, the other with a complex one. One or the other of these plates has been brought to the present level of exposure in the different Carboniferous fault blocks depending on the amount and sense of vertical displacement on the bounding faults.

The Silurian-Lower Devonian stratified rocks of the Cobscook anticline and the syncline of Passamaquoddy Bay are intruded by the Acadian plutonic belt and overlain along an angular unconformity by the Upper Devonian Perry Formation whose conglomerates carry clasts of the underlying volcanic rocks and granitic rocks of the plutonic belt. The folding is thus a part of the Acadian orogeny.

GEOPHYSICS

Three geophysical anomalies set the Passamaquoddy Bay area apart from surrounding coastal Maine and New Brunswick. First, the area has a concentration of small earthquakes which can be traced back through the historical record and are now more accurately recorded instrumentally (Figure 4). In his discussion of seismicity in Maine, Ebel (this volume) notes that only

in the Passamaquoddy Bay area where earthquakes are spread throughout the subsiding area is there an obvious relation between seismicity and geologic or tectonic structures, although no direct relationship between specific epicenters and known faults is apparent. The concentration of continuing seismicity indicates that the crust beneath Passamaquoddy Bay and vicinity is not strong enough to accumulate much strain.

The Bouguer gravity anomaly map (Figure 5) shows a northwest-trending embayment of relatively flat gravity ranging from slightly positive to slightly negative beneath Passamaquoddy Bay and adjacent Maine. The embayment corresponds to the thick accumulation of the Silurian-Lower Devonian volcanic rocks and to the distribution of the Perry Formation. The granitic and gabbro-diorite plutons that underlie the gravity embayment south and southwest of Calais have little gravitational expression and hence are probably relatively thin. The large negative anomalies to the north and west of the gravity embayment coincide in part with surface exposures of granitic plutons, which presumably extend to considerable depth.

The steep gravity gradient, which becomes increasingly positive to the southeast and extends from Campobello Island southwestward beneath the Quoddy fault block, coincides with large masses of diabase and gabbro that intrude and presumably underlie the Quoddy Formation. This steep gradient continues 175 km to the southwest, diverging from the coastline. Southeast of Mt. Desert Island, it is 25 km offshore in the Gulf of Maine. Kane and others (1972) conclude that it marks a major boundary between two crustal blocks, one of comparatively dense Precambrian rock beneath the Gulf of Maine, the other of less dense Paleozoic rocks overlying the Precambrian basement in coastal Maine. Superimposed on this regional gravity gradient and within the crustal blocks are numerous large plutons of gabbroic and granitic rocks which show as large closed or partly closed positive or negative anomalies respectively. The volume of gabbro and diabase intruding and probably underlying the Quoddy Formation may have been emplaced along this crustal boundary, thus dating the boundary as at least Early Silurian in age.

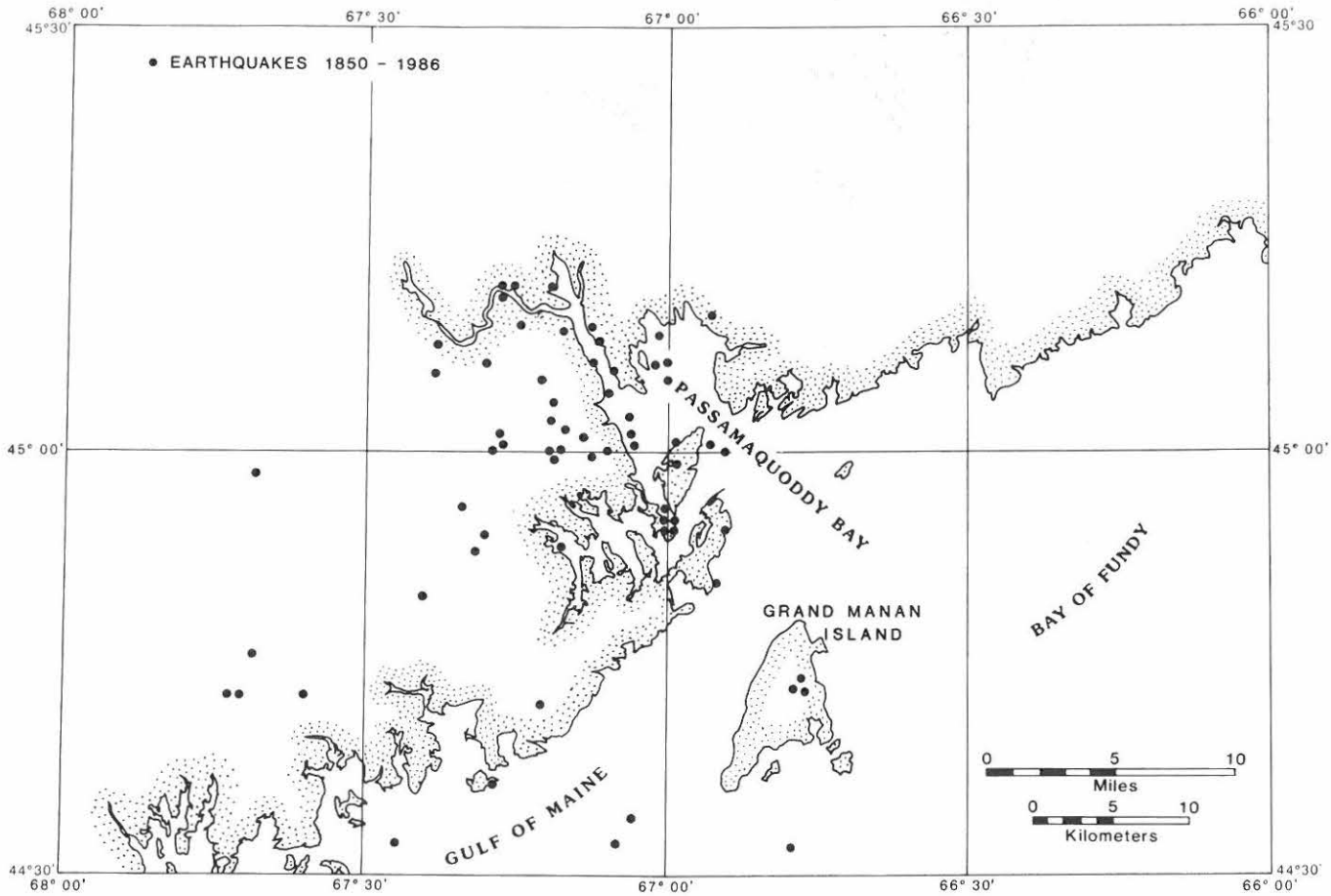

Figure 4. Epicenters of earthquakes from 1850 to 1986, Passamaquoddy Bay area, Maine and New Brunswick. (compiled from Rand (1977), Chiburis (1981), Johnston and Foley (1987), and Seismicity of the Northeastern United States, various editors, 1981-1986, Northeastern U.S. Seismic Network, Bulletins 22-37)

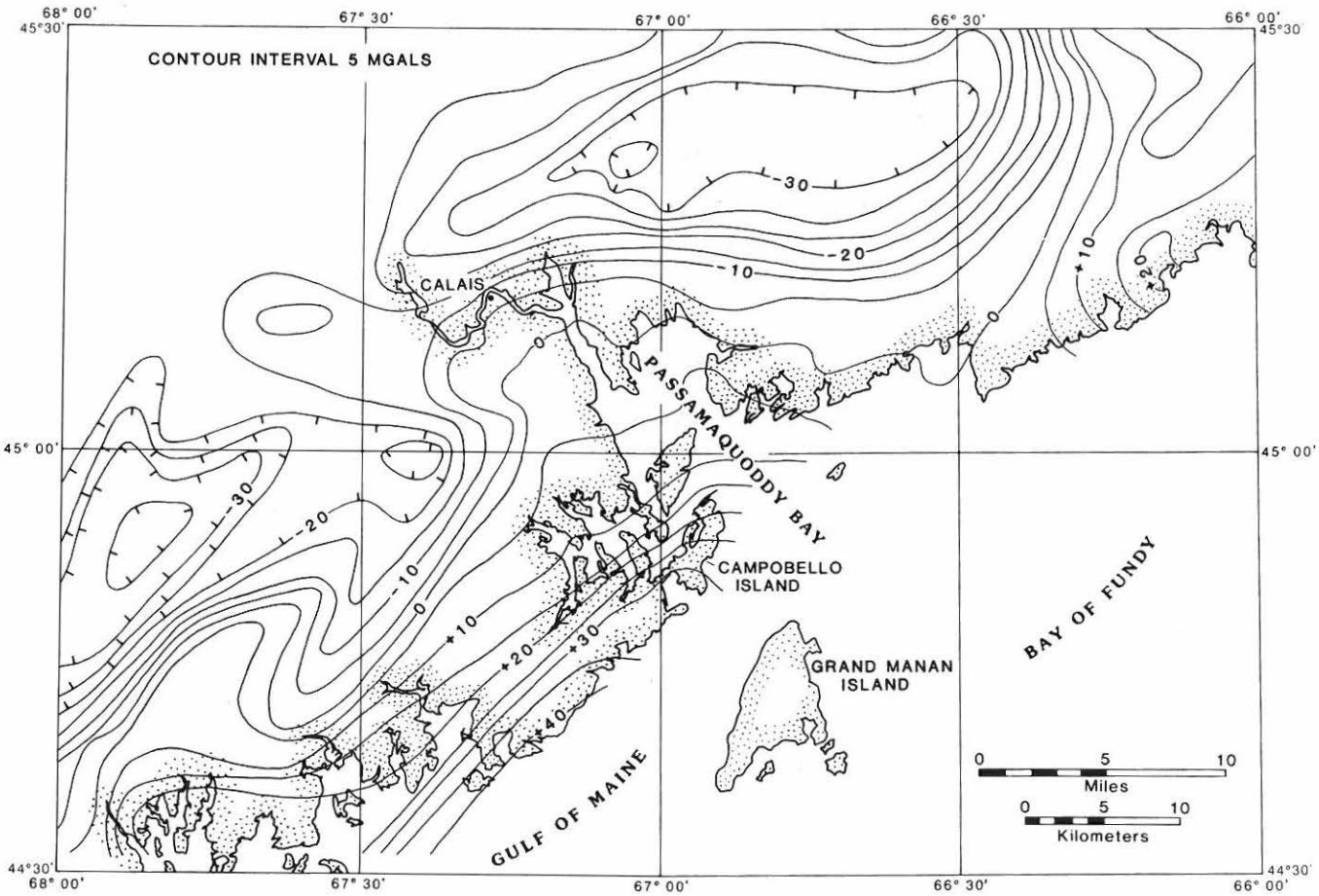

Figure 5. Bouguer gravity anomaly map of coastal Maine and New Brunswick adjacent to Passamaquoddy Bay. (Compiled from Kane and others (1972) and Weng (1981))

The magnetic anomaly map (Figure 6) shows a discontinuity between a linear pattern in New Brunswick, especially on Deer and Campobello Islands, and a flat pattern in Maine just across Passamaquoddy Bay. The linear pattern of magnetic highs corresponds to the late Precambrian Coldbrook Formation of volcanic rocks on Campobello Island, on Deer Island presumably underlying the Silurian rocks, and on the Kingston peninsula north of Saint John. The more diffuse linear pattern of magnetic lows northeast of Saint John corresponds to the Precambrian Greenhead Group of quartzites, limestones, and gneisses in the Saint John area. The linearity of magnetic values abruptly terminates at about the location of the Oak Bay fault. The Greenhead linear pattern appears again on the east side of Grand Manan Island where rocks correlated with the Greenhead are faulted against Triassic rocks on the west side of the island (Potter and others, 1979). Kane and others (1972) trace the linear pattern of the Precambrian rocks to the southwest of Grand Manan beneath the Gulf of Maine.

If the strike of the Greenhead rocks and their magnetic signature are followed southwestward they appear to be offset 10 to 15 km eastward to Grand Manan. Presumably the parallel pattern of the Coldbrook rocks is also offset but obscured by the overlying Triassic rocks on Grand Manan. This offset is far greater than that ascribed to the Oak Bay fault. No corresponding displacement can be found in the pattern of Ordovician, Silurian and Devonian sedimentary rocks inland from Calais and Passamaquoddy Bay on the maps of Ruittenberg and Ludman (1978), and Potter and others (1979). If the offset of magnetic patterns and corresponding rocks does represent a northwest trending discontinuity in the crust, it must date back to the late Precambrian or Cambrian.

GEOLOGIC HISTORY

The late Precambrian and early Paleozoic history of the Passamaquoddy Bay area is tied to that of the Avalonian terrane. In the Saint John area (Figure 6), the Precambrian Coldbrook Formation and the overlying Saint John Group of fossiliferous Cambrian to Lower Ordovician sandstones and shales (Alcock, 1940) have been assigned to the Avalonian terrane on the basis of similar Precambrian volcanic rocks and fossiliferous

Cambrian rocks on the Avalon Peninsula of Newfoundland (Williams, 1978; Williams and Hatcher, 1983). The graptolite-bearing Cookson Formation underlying the Oak Bay conglomerate on Cookson Island correlates in age with Lower Ordovician graptolitic shales at the top of the Saint John Group, but whether the Cookson Formation belongs to the same geological tract as the Saint John Group is uncertain (Fyffe and Pickerill, 1986; Ludman, 1986b). Whether Avalonia accreted to North America during the Devonian Acadian orogeny or prior to the Silurian is controversial (for example Wilson, 1966; Dewey and Kidd, 1974; Osberg, 1978; Gates and Moench, 1981; Williams and Hatcher, 1983; Ludman, 1986a).

During the Silurian and Early Devonian, a thick pile of marine volcanic rocks accumulated in a subsiding, extensional tectonic regime. The volcanic rocks probably rest unconformably on a composite basement of Avalonian terrane and perhaps other terranes (Williams and Hatcher, 1983), but can be considered part of Avalonia only if Avalonia accreted to North America during the Devonian Acadian orogeny or later. Ludman (1986a) marshals evidence that Avalonia, and perhaps other exotic terranes, joined North America prior to the Silurian period and the suture or sutures were subsequently buried beneath the Silurian and Devonian sedimentary rocks of the central Maine flysch belt and the Fredericton trough. The Silurian-Lower Devonian volcanic pile thus may be autochthonous, deposited on an already docked Avalonian basement and contiguous with the clastic rocks of the same age in central Maine and southwestern New Brunswick.

Acadian deformation, not well understood, produced folds of differing complexity, attitude, and degree of foliation. Intrusion of Devonian mafic to felsic magmas was largely confined to a wide linear plutonic belt. During the Late Devonian, a local fault basin filled with coarse alluvial sediments derived from erosion of the volcanic pile and Acadian plutonic rocks. Regional northeast-striking faults and the north-northwest striking Oak Bay fault of Carboniferous age divided the rocks of the Passamaquoddy Bay area into discrete structural blocks having different stratigraphies and structural styles. A Lower Jurassic diabase dike is the youngest pre-Quaternary rock in the area. The last recorded fault, the Fundian, delineated the northwestern margin of the Late Triassic to Early Jurassic fault basin now largely beneath the Bay of Fundy and the Gulf of Maine.

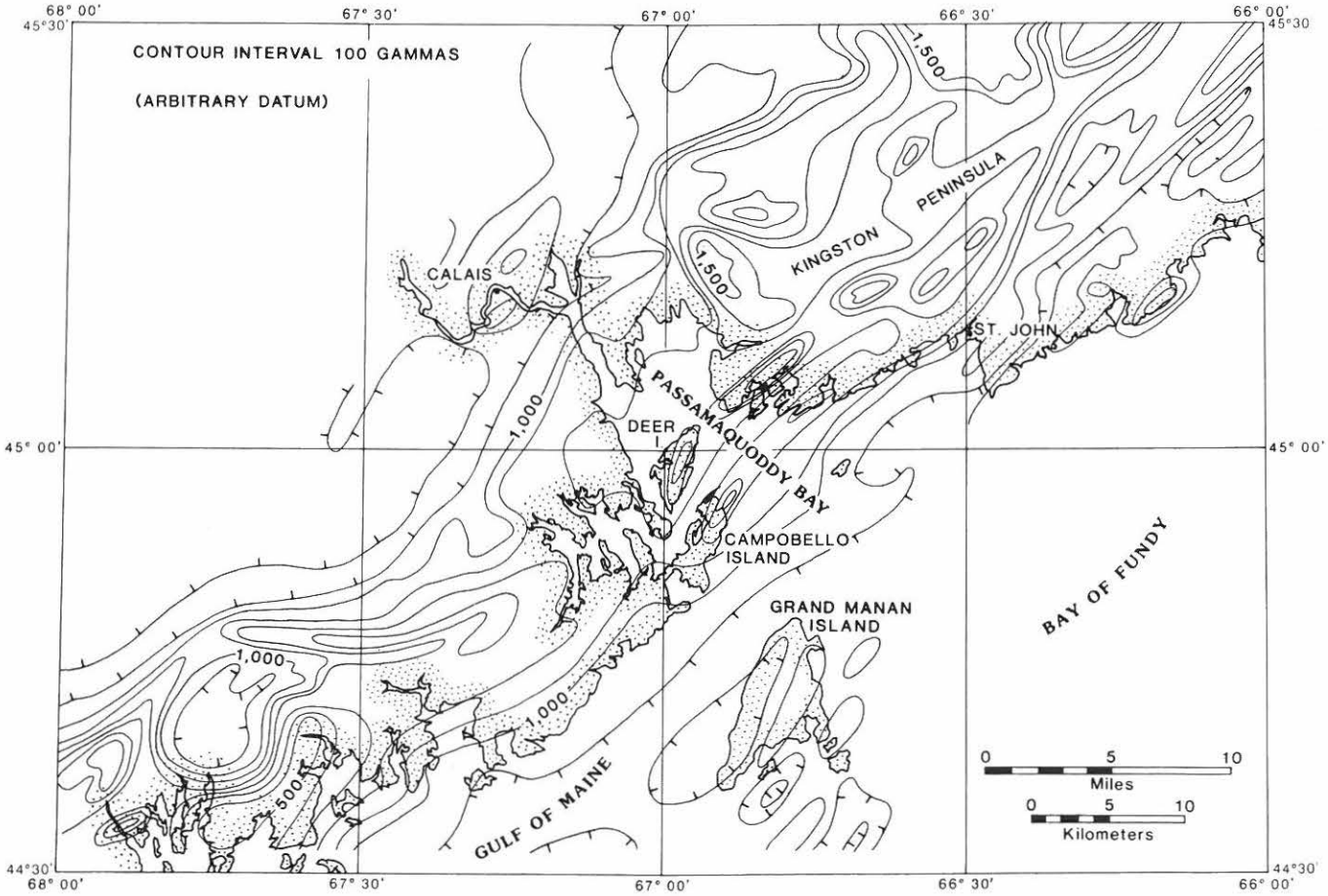

Figure 6. Aeromagnetic anomaly map of coastal Maine and New Brunswick adjacent to Passamaquoddy Bay. (Adapted from Kane and others (1972))

The final chapter in the geologic history of Passamaquoddy Bay and coastal Maine involves large changes in the direction, magnitude, and rate of crustal rebound following withdrawal of the Pleistocene ice sheets. Stuiver and Borns (1975) bracket submergence by the Presumpscot sea and its withdrawal to about the present coast within a 1,000 year period beginning about 13,000 yr BP. The uplift of glaciomarine deltas since maximum extent of the Presumpscot sea (Thompson and others, this volume) indicates total emergence along the outer Maine coast of about 61 meters. The sea, however, originally withdrew to an apparent stillstand at about -65 meters based on depth profiles and seismic lines in the adjacent Gulf of Maine (Schnitker, 1974; Kelley and others, this volume; Belknap and others, this volume) and dated uncertainly at about 9,000 yr BP. Thus within about 4,000 years, the present coastline area underwent an emergence of perhaps as much as 120 meters depending on the elevation of the hypothetical offshore isobase, a rate of 25-30 mm/yr. Between the maximum emergence and the beginning of coastal salt marsh data at about 5,000 yr BP at an elevation of -5.5 meters the outer coast submerged about 60 meters in about 4,000 years, a rate of 15 mm/yr averaged over a curve that progressively flattens with time to join that of the salt marsh data (Belknap and others, this volume). From 5,000 to 1,500 yr BP, salt marsh studies (Belknap and others, this volume) show a relatively constant submergence rate of 1.44 mm/yr with little evidence of differential crustal warping between Addison and Wells, about 15 km north of the border with New Hampshire. After 1,500 yr BP, the data are more scattered but Belkap and coauthors postulate a rate of .5 mm/yr.

There are much less data with which to reconstruct the history of sea level changes for Passamaquoddy Bay. Depths on the navigation chart show that at the -65 meter stillstand, several narrow lakes probably occupied the deep channels between Eastport and Deer Island and between Deer Island and Campobello. The subsequent subsidence history rests on evidence from Indian clam shell middens along the present shore and of radiocarbon dates from Digdeguash Harbor. Sanger (1985, this volume) describes Indian middens which provide artifact and radiocarbon dates for occupation from about 2,500-2,000 yr BP to historical times. Sanger points out that their preservation suggests a fairly stable sea level for almost 2,000 years. He notes, however, that the absence of older middens, common in the Penobscot Bay area, suggests that greater submergence than that farther west may have drowned or eroded older middens in Passamaquoddy Bay prior to 2,500 yr BP. He also comments on the current rapid erosion of existing middens, indicating a recent rise of sea level in Passamaquoddy Bay.

Cores taken from the bottom of Digdeguash Harbor contain clam shells that yield radiocarbon dates from about 5,300 to 2,700 yr BP (Sanger, 1985). Although sampling and coring problems preclude use of these cores for detailed quantitative analysis of sea level rise, the clams do prove that the sea occupied Digdeguash Harbor at least 5,000 radiocarbon years ago.

Straight-line extrapolation backwards of some proposed subsidence rates permit a reasonable guess of the rate of subsidence for Passamaquoddy Bay. The 1.44 -0.5 mm/yr curve of Belknap and others (this volume) for coastal Maine between Addison and Wells puts sea level 1 meter lower 2,000 years ago at the midden sites and 5 meters lower 5,000 years ago in Digdeguash Harbor. These figures seem reasonable. The 3.2 mm/yr current sea level rise rate from the Eastport tide gauge places sea level 6.4 meters lower at the middens 2,000 years ago, drying up present and likely past clam flats. At Digdeguash Harbor 5,000 years ago sea level would have been 16 meters lower, completely draining the harbor. Thus, the subsidence rate for Passamaquoddy Bay averaged over the last 5,000 years appears to be closer to that for the Maine coast from Addison to Wells than to the current tide gauge rate at Eastport.

Present rates of subsidence in the Eastport area are described by Anderson and others (1984), Reilinger (1987), and by accompanying papers in this volume. Rates of subsidence derived by using different methods are not in agreement, and there is some evidence for very short term fluctuations. Nevertheless, in balance, subsidence seems to be faster than along the coast of Maine to the west and in New Brunswick to the east (Vanicek, 1976). Survival of the middens for 2,000 years coupled with their present rapid erosion suggests that the downwarping has increased recently, perhaps within the last few centuries.

The glaciomarine deltas (Thompson and others, this volume) provide the long term record of postglacial crustal rebound warping in Maine. Extension of the delta elevation contours farther into New Brunswick would strengthen geologic interpretation of the reentrant in the Passamaquoddy Bay area by confirming its areal extent and allowing comparison with the long term crustal warping record in coastal New Brunswick. The reentrant as presently drawn suggests a crustal downwarping, or lag in uplift, of at least 12 meters compared to the rebound record along coastal Maine to the west. Most of this differential downwarping must have developed prior to 5,000 yr BP if the limited data from middens and Digdeguash Harbor are anywhere near the mark.

The history of crustal warping in the Passamaquoddy Bay region relative to that of coastal Maine to the west thus can be divided into three chapters: 1) differential downwarping relative to most of coastal Maine of at least 12 meters prior to 5,000 yr BP as suggested by the delta elevations; 2) a period of relative stability from then to perhaps historical times implied by the archaeology and the salt marsh data; and 3) a recent increase of uncertain magnitude in the rate of subsidence indicated by modern data.

CONCLUSIONS

A weak fractured crust is responsible for localizing past and present differential warping and seismicity in the Passamaquoddy Bay region. Part of the evidence is historical, beginning

perhaps with the northwest-trending Precambrian to Cambrian crustal discontinuity vaguely suggested by the offset of the linear magnetic patterns. The thickness of the Silurian-Lower Devonian volcanic pile is greater in the Eastport region than elsewhere in the coastal volcanic belt, implying greater local subsidence. The development of the Upper Devonian Perry fault basin and its subsequent survival from erosion suggest a continuing local tendency towards subsidence. The numerous faults of many ages show repeated crustal fracturing; and the concentration of present seismicity demonstrates that the local crust is not sufficiently strong to sustain much strain without movement along fractures.

A rough coincidence in time between the changing magnitudes and rates of the crustal warping history of coastal Maine as a whole and that of Passamaquoddy Bay suggests that strains due to crustal rebound acting on the locally weak crust are responsible for most of the 12 meter differential subsidence indicated by the reentrant in glaciomarine delta elevation contours. Judging from the midden and Digdeguash Harbor data most of that differential subsidence occurred sometime prior to 5,000 yr BP when crustal warping of large magnitudes was proceeding at high rates along coastal Maine and presumably New Brunswick. By 5,000 yr BP, submergence of coastal Maine had slowed to 1.44 mm/yr; and from then on differential submergence of the Passamaquoddy Bay area appears to have been slight, at least until recently.

The large and rapid rebound and the following rapid submergence of coastal Maine ending sometime prior to beginning of the salt marsh data must have imposed much stress on the regional crust. Local isostatic disequilibriums due to large contrasts in the density of the rock bodies adjacent to Passamaquoddy Bay, coupled with hydroisostatic loading of the crust beneath the Bay of Fundy and the Gulf of Maine as the sea deepened rapidly early in the submergence following the -65 meter stillstand, must have imposed additional stresses.

The straight delta elevation contours in coastal Maine indicate that the crust of the coastal region reacted to the stresses by uniform warping; but the weak crust of the Passamaquoddy Bay area failed by differential downwarping. Rapid postglacial crustal warping had run its course by 5,000 yr BP as indicated by the slow submergence recorded in the salt marsh data. Kane and others (1972), citing the data of Yellin (1968), concluded that the slightly positive free air gravity anomaly over coastal Maine and the Gulf of Maine indicates that the crustal blocks of different densities beneath the two regions are now in isostatic equilibrium.

This model implies that the crustal stresses which produced most of the 12 meter differential warping need not, and probably are not, the same as those powering the present subsidence and seismicity in the Passamaquoddy Bay area. The only connection may be the area of crustal weakness (due to repeated fracturing perhaps since the Cambrian) beneath the Bay common to both warping regimes. Now that postglacial rebound has ceased, the only identified remaining stresses in the crust are those that throughout the northeastern United States have measured orientations and magnitudes consistent with derivation from motion of the North American plate (Ebel, this volume; Lee, this volume) plus local gravitational ones. Presumably these stresses are constant through time but were once masked or overridden by the larger stresses of postglacial rebound. The postulated recent increase in subsidence and the possible recent short-term fluctuations in the Eastport area might reflect an underlying crust which, though fractured, has sufficient cohesion along the fractures to accumulate strains produced over thousands of years by the present regional stress field up to some low threshold amount. The crust then fails along fractures, releasing the strain energy by subsidence and seismicity over a period of time perhaps measured within a century or less.

The above very tentative model explaining the past and present differential downwarping of the Passamaquoddy Bay area rests on a weak data base. Needed to strengthen (or demolish) the model are extension of the glaciomarine delta elevation data farther into New Brunswick, acquisitions of more definitive quantitative data on the warping history of Passamaquoddy Bay and farther east in New Brunswick, and resolution of the conflicting data on the rate of the current subsidence in the Eastport area.

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