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A Stratigraphic Synthesis of Eastern Maine and Western New Brunswick

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ABSTRACT

Lower Devonian and older strata from Portland, Maine to Saint John, New Brunswick and 150 miles inland are divided into nine northeast-trending lithotectonic belts based on their internal stratigraphy. Provincial Cambrian fauna of Old World affinity are present in the Saint John belt; Arenigian to Llanvirnian fauna of the Celtic province are present in the northern Maine and Miramichi belts; Caradocian brachiopods of the Benner Hill belt are disputedly either Old World or undiagnostic; and Pridolian to Gedinnian fauna have Old World affinity in the Ellsworth belt and North American affinity in the Miramichi and northern Maine belts. For all belts and ages other than these, fossils are either absent or not diagnostic of biogeographic province.

The following stratigraphic interpretations are proposed. Ellsworth belt: the Ellsworth and Queen Brook Formations are correlative and pre-Silurian. The Llandoverian Quoddy Formation is equivalent to unnamed units near Sagwa, at Ketchum Brook, and at the eastern end of the belt. All rest disconformably across the Queen Brook Formation, Cambrian sandstones, and volcanic rocks possibly equivalent to the Precambrian Coldbrook Group. Islesboro-Rockport belt: the three sequences, separated by unconformities, are (1) retrograded Precambrian schist, marble, and amphibolite, (2) the Islesboro sequence, and (3) polymictic conglomerate, gray sandstone and phyllite. St. Croix belt: the upper Jam Brook Formation, with basal polymictic conglomerate, may represent the base of the Megunticook Formation, both of which are faulted. The lower Jam Brook Formation may be Precambrian. Casco Bay - south-central Maine belt: Cushing Formation and Casco Bay Group contain pre-Ashgillian rocks in a faulted anticlinal inlier. The Passagassawakeag Gneiss, at the northeastern end, may contain high-grade equivalents of the Casco Bay Group. The inlier is unconformably overlain by the Macworth and Hutchins Corner Formations at the base of the Ashgillian(?)-Gedinnian central Maine cover sequence. The Appleton Ridge Formation may be Devonian. Fredericton belt: this belt includes Silurian strata on both sides of the Fredericton fault. The Pocomoonshine Lake Formation, against the St. Croix belt, may be Lower Devonian rather than Cambrian(?). Miramichi belt: widespread Cambrian(?) sandstone and slate is capped by Lower Ordovician black slate, which is unconformably overlain by Middle Ordovician volcanic rocks, slate, limestone, chert, and iron formation with complex facies relationships. Northern Maine belt: black slate and mélange above the Grand Pitch Formation is equated with the Lower Ordovician Chase Brook Formation to the northwest. They are unconformably beneath rocks ranging from the Arenigian Shin Brook Formation to the Caradocian Chase Lake Formation. Other interpretations for these belts and the remaining Saint John and Benner Hill belts generally follow previous workers.

Many age assignments are uncertain, but the following relationships are inferred. Precambrian: there are no reasonable equivalents for the middle Proterozoic Green Head Group of the Saint John belt. The lower Jam Brook Formation is similar to the Islesboro sequence. Other correlations are uncertain. Cambrian: Cambrian strata of the Saint John, Ellsworth, Islesboro, and St. Croix belts are correlated, and Cambrian rocks of the northern Maine, Miramichi, and less confidently, Casco Bay belts are correlated. Tremadocian-Caradocian: Lower Ordovician black shales, many graptolitic, are present in the Saint John, St. Croix, Miramichi, and northern Maine belts, suggesting that they were covered by a continuous black sheet in the Early Ordovician. Heterogeneous, volcanic-rich Llanvirnian-Caradocian sections in the northern Maine, Miramichi, and Casco Bay - south-central Maine belts are correlated, but contemporaneous(?), volcanic-poor, Ellsworth belt rocks might not correlate. Ashgillian-Ludlovian: thin, relatively shallow-water, clastic sections present around the pre-Ashgillian inliers in the northern Maine belt, and in the Miramichi, St. Croix, and Ellsworth belts, are areally restricted and cannot be correlated between belts. Thick, deeper-water sections in the northern Maine, Casco Bay - south-central Maine, and Fredericton belts, however, are similar and correlation is proposed. Pridolian-Gedinnian: volcanic and sedimentary rocks of the northern Maine and Miramichi belts correlate, and are North American. Similar rocks in the Ellsworth belt have Old World affinity and correlation is precluded. Siegenian-Emsian: proposed correlation of the Carrabassett, Seboomook, Appleton Ridge, and Pocomoonshine Lake Formations suggests Lower Devonian flysch overspread at least the Fredericton, Casco Bay - south-central Maine, and northern Maine belts.

Two equally plausible tectonic models arise. In one, all the belts belonged in the Cambrian to the Celtic archipelago at the western edge of Gondwana. In the late Cambrian-earliest Arenigian, the Eastern Iapetus, between the Celtic archipelago and the Chain Lakes terrane, closed along a west-dipping subduction zone, culminating in the Penobscottian orogeny. From Llanvirnian to Caradocian time, the Western Iapetus, between the Chain Lakes-Celtic archipelago terrane and North America, closed along an east-dipping subduction zone. Taconian back-arc spreading, recorded by Caradocian volcanic rocks in the Miramichi belt, continued rifting into the Silurian, creating the Kronos Ocean. In the late Silurian, the Kronos began to close by subduction, predominantly westward to produce the volcanic rocks of the northern Maine belt, but intermittently eastward beneath the Ellsworth belt of the Avalon terrane. A blanket of Siegenian-Emsian flysch was deposited westward in front of the approaching Avalon until the late Emsian-Givetian Acadian orogeny.

In the alternative tectonic model, the Kronos Ocean existed from the Cambrian to the Early Devonian, between the northern Maine, Miramichi, Casco Bay - south-central Maine, and Fredericton belts of the Celtic archipelago on one side and the eastern belts of Gondwana on the other.

INTRODUCTION

Stratified Lower Devonian and older rocks of eastern Maine and western New Brunswick can be divided into lithotectonic belts as shown in Figure 1. We have identified the Saint John, Ellsworth, Islesboro-Rockport, Benner Hill, St. Croix, Casco Bay - south-central Maine, Fredericton, Miramichi, and northern Maine belts. In general, the lithostratigraphy within each belt is reasonably well known, and in most belts a few fossils roughly constrain the chronostratigraphy. But how the belts are related to each other is an important question. Possible answers range between two extreme models: (1) every belt represents a different terrane which evolved for at least part of its history independently of the other belts, and (2) all belts are stratigraphically related, and differences reflect sedimentary facies and deformational, metamorphic, and igneous effects superimposed on the belts during orogenesis.

We have taken the view that a lithotectonic belt contains a coherent stratigraphy and encompasses only those regions for which the local stratigraphic names are applicable or to which they can be confidently extended. This procedure maximizes the number of lithotectonic belts, and our task is to use the techniques of terrane analysis as outlined by Jones et al. (1983) to investigate the connections between them. In this regard, because of the importance of faunal diversity, we will first discuss faunal provinciality. We will then define the boundaries of the lithotectonic belts and describe their internal stratigraphic relationships. Finally, we will discuss permissible connections between belts and tectonic events that can explain the juxtaposition of lithotectonic units.

All of the lithotectonic belts were deformed and metamorphosed during the Acadian orogeny, although some of the belts may also contain tectonic elements produced in earlier events (Penobscottian, Taconian) and later events (Alleghenian). Post-tectonic Devonian plutons seal most boundaries, and slightly deformed or undeformed Middle Devonian to Carboniferous sedimentary rocks cover many belts (Fig. 1), so the major tectonic elements appear to have been in place since the

Devonian. Even so, the faults that bound most lithotectonic belts at the present level of erosion have uncertain ages and amounts of motion across them. Their role in the assembly of the region is at issue. We believe that most faults presently mapped were active late in the tectonic history, and that, by offsetting unconformities or preexisting tectonic boudaries, they obscure the original relationships between adjacent belts. Because of the possibly repeated and complex motion along these faults, as well as in many cases their unknown three-dimensional geometry, it is difficult to restore the rocks to their pre-fault relationships. Instead, we have chosen to assess the significance of these boundaries by comparing the rocks of the belts themselves. No rock unit or stratigraphic sequence can be mapped continuously across the region. Therefore, proposed correlations and facies relationships between strata exposed in different belts depend on similarities in lithologies, stratigraphic sequence, and faunal elements.

FAUNAL PROVINCIALITY

Faunal provinciality is shown by symbols in Figure 2. North American, Celtic, Old World, and cosmopolitan faunal associations have been specified. Symbol type indicates the provinciality, and symbol position in the stratigraphic column indicates the age of the fossils.

Cambrian fossils have been reported from the Saint John, Ellsworth, and northern Maine belts. Fossils of the Saint John belt have an Old World affinity (Palmer, 1971; North, 1971), and because the fossils in the Cambrian of the Ellsworth belt are similar (Yoon, 1970), they are also regarded as Old World. *Oldhamia* in the northern Maine belt is cosmopolitan.

Fossils recovered from Tremadocian rocks in the region, primarily graptolites, are found in the Saint John, St. Croix, and Miramichi belts. In addition, *Caryocaris* in the northern Maine belt suggests a Tremadocian age, and conodonts from the same formation suggest a slightly younger age. In the Tremadocian, North America is thought to have been equatorial and the Old World was at high latitudes (Cocks and Fortey, 1982). Tremadocian graptolites from the Saint John belt have not been studied from the point of view of provinciality. Graptolites from the St. Croix and Miramichi belts are identical (Fyffe et al., 1983), and John Riva (pers. commun., 1989) has indicated that these forms are known in Scandinavia, but not in North America outside of the Appalachians. These graptolites are deep-water forms, however, and should have equal opportunity to appear on both sides of an ocean, if it were not for latitudinal differences. The provinciality of *Caryocaris* is unknown, but the conodonts, like the graptolites, are deep-water forms (J. E. Repetski, pers. commun. to B. A. Hall, 1983). None of these Tremadocian fossils differentiate continents.

Arenigian to Llanvirnian fossiliferous rocks have been reported in the Saint John, Miramichi, and northern Maine belts. Arenigian fossils in the northern Maine and Miramichi belts have been assigned to the Celtic province (Neuman, 1984), which is characterized by having stronger affinities to Old World than to North American fauna, yet is distinct from both. Llanvirnian fossils in the northern Maine and Miramichi belts, and Arenigian fossils in the uppermost part of the Saint John Group are not diagnostic with regard to province.

Fossiliferous rocks of Llandeilian to Ashgillian age are found in the northern Maine, Miramichi, and Benner Hill belts. Brachiopod assemblages from the northern Maine belt have a North American aspect (Neuman, 1984), but conodonts and graptolites of this age in the northern Maine and Miramichi belts have not been used to identify faunal province. The provinciality of deformed brachiopods from the Benner Hill belt is disputed. Boucot et al. (1972) and Boucot (1973) interpreted them to be Old World, but Neuman (1973) thought their deformed condition, along with a possibly incomplete assemblage, invalidated such a conclusion. Because of the regional importance of this fossil locality, we believe a concerted effort should be made to find additional fossils in the Benner Hill belt which would establish its faunal affinity.

Ashgillian fossils from the northern Maine belt have been interpreted to indicate environmental diversity within the North American province (Neuman et al., in press).

Fossil assemblages of Llandoverian through Wenlockian age are found in the Ellsworth, Casco Bay - south-central Maine, Miramichi, and northern Maine belts. These fossils are cosmopolitan within the Appalachian-Caledonide mountain belt (Cocks and McKerrow, 1977).

Provincial communities were re-established in latest Ludlovian to Pridolian time and persisted well into the Early Devonian. Fossiliferous rocks of this age occur in the Ellsworth, St. Croix, Miramichi, and northern Maine lithotectonic belts. Both brachiopods (Johnson, 1979) and ostracodes (Berdan, 1970, 1971; Martinsson, 1970; Brookins et al., 1973; Copeland and Berdan, 1976) indicate that the Ellsworth belt contains Old World assemblages. Brachiopods from the Waweig Formation in the St. Croix belt (Pickerill, 1976) have not been studied from the point of view of provinciality, but because this assemblage is almost identical to those of the Leighton and Hersey Formations in the Ellsworth belt, it implies that the St. Croix belt also belongs to the Old World province. On the other hand, brachiopod assemblages from the Hartin Formation in the Miramichi belt (A. J. Boucot, pers. commun. to D. V. Venugopal, 1978 and 1979) and from the Seboomook Formation in the northern Maine belt (Boucot, 1968) have North American affinities.

LITHOTECTONIC BELTS

General Statement

The geology and the distribution of lithotectonic belts shown in Figure 1 agree largely with maps presented by Ruitenberg et al. (1977), Ruitenberg and McCutcheon (1978, 1982),

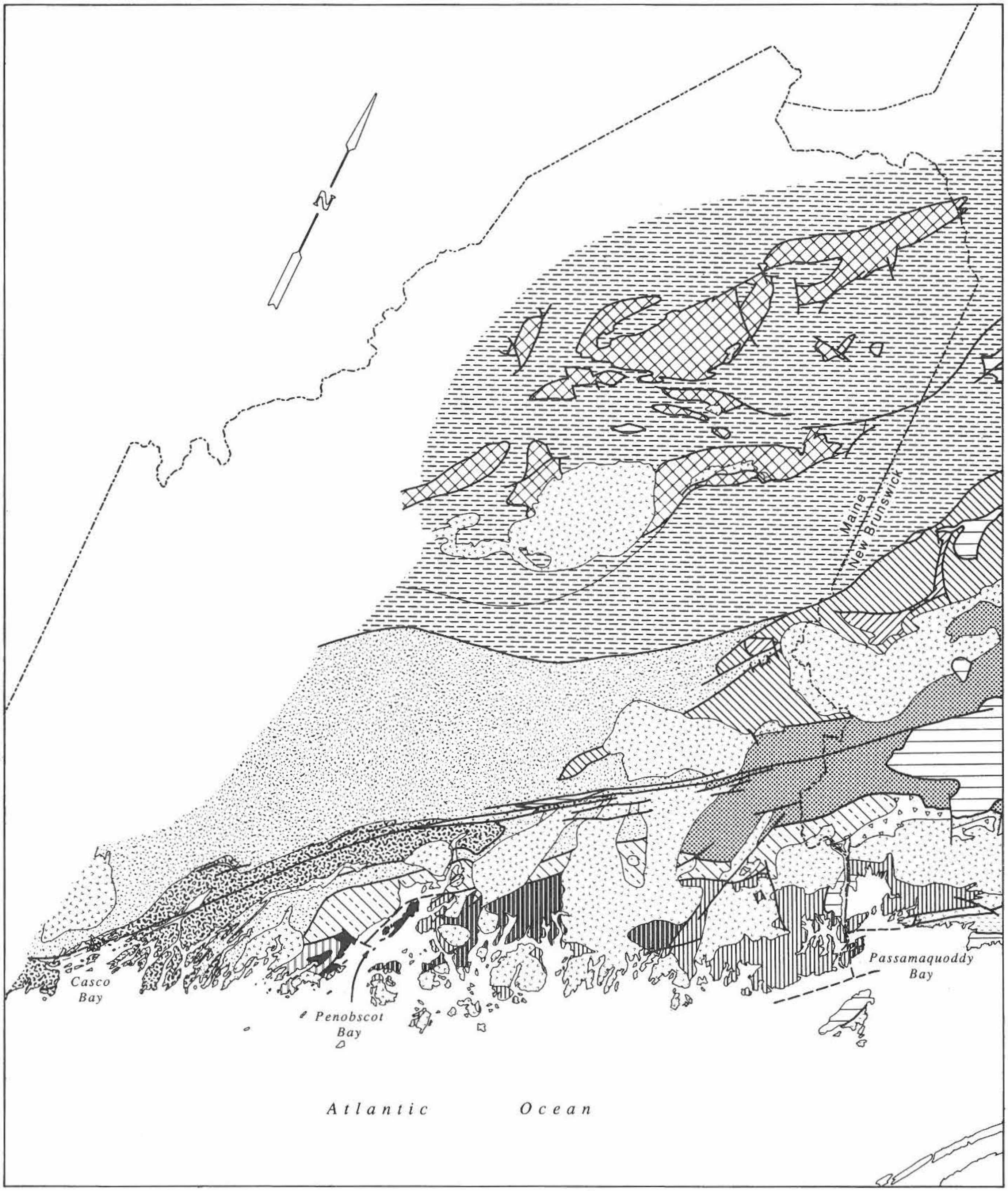
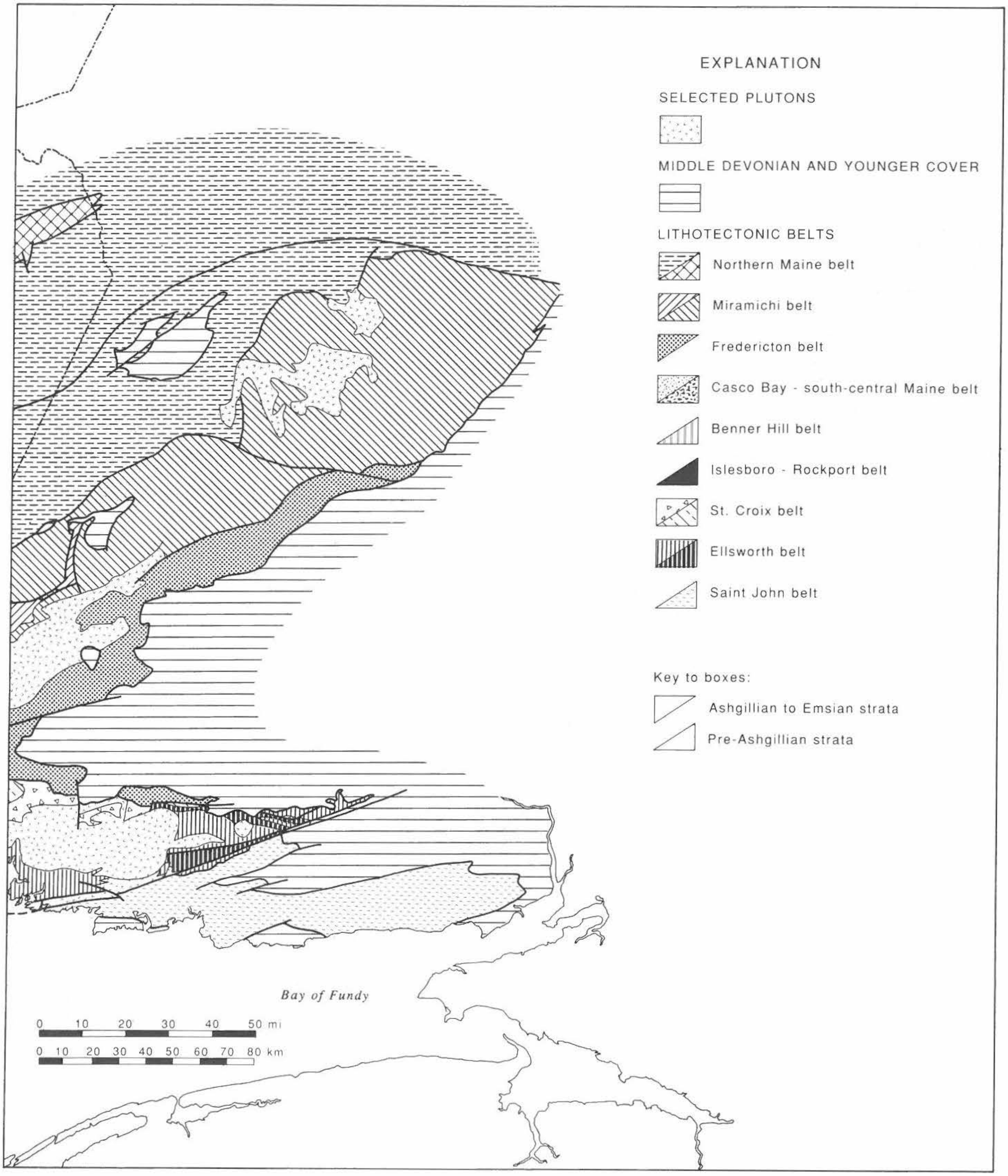


Figure 1. Lithotectonic map of eastern Maine and western New Brunswick.



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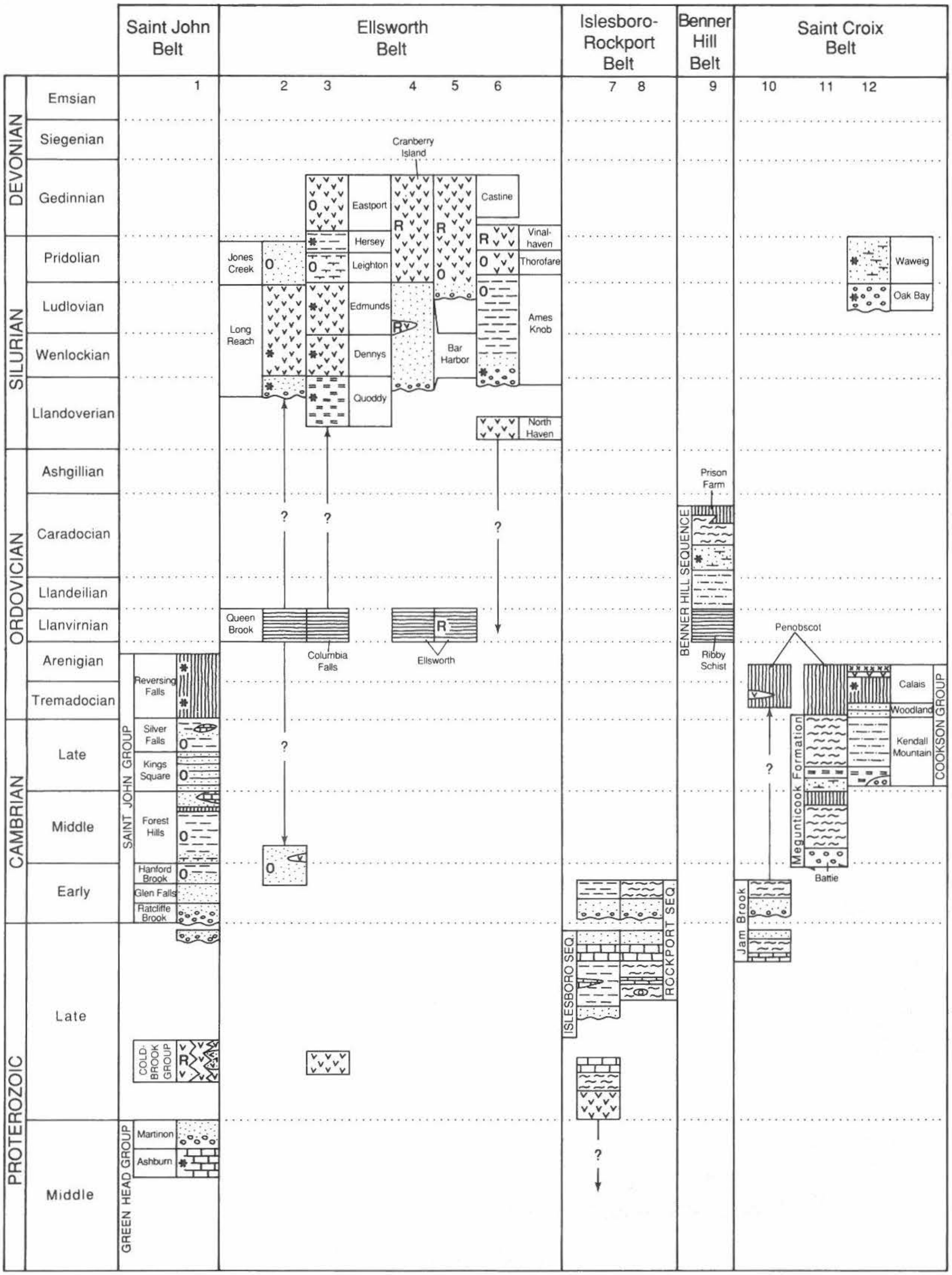

Figure 2. Interpreted stratigraphic columns discussed in text, showing ages of units and predominant rock type. Column locations shown in Figure 3. Vertical arrows suggest alternative ages for some units. For discussion of correlations and age control, see text. Strict adherence to the Stratigraphic Code for unit designations has not been attempted here nor in text. Abbreviations: SEQ. = Sequence, R = Radiometric age control. Faunal provinciality: O = Old World, N = North American, C = Celtic, * = Cosmopolitan. See Editors' Note regarding stratigraphic order of the Cookson Group presented in column 12.

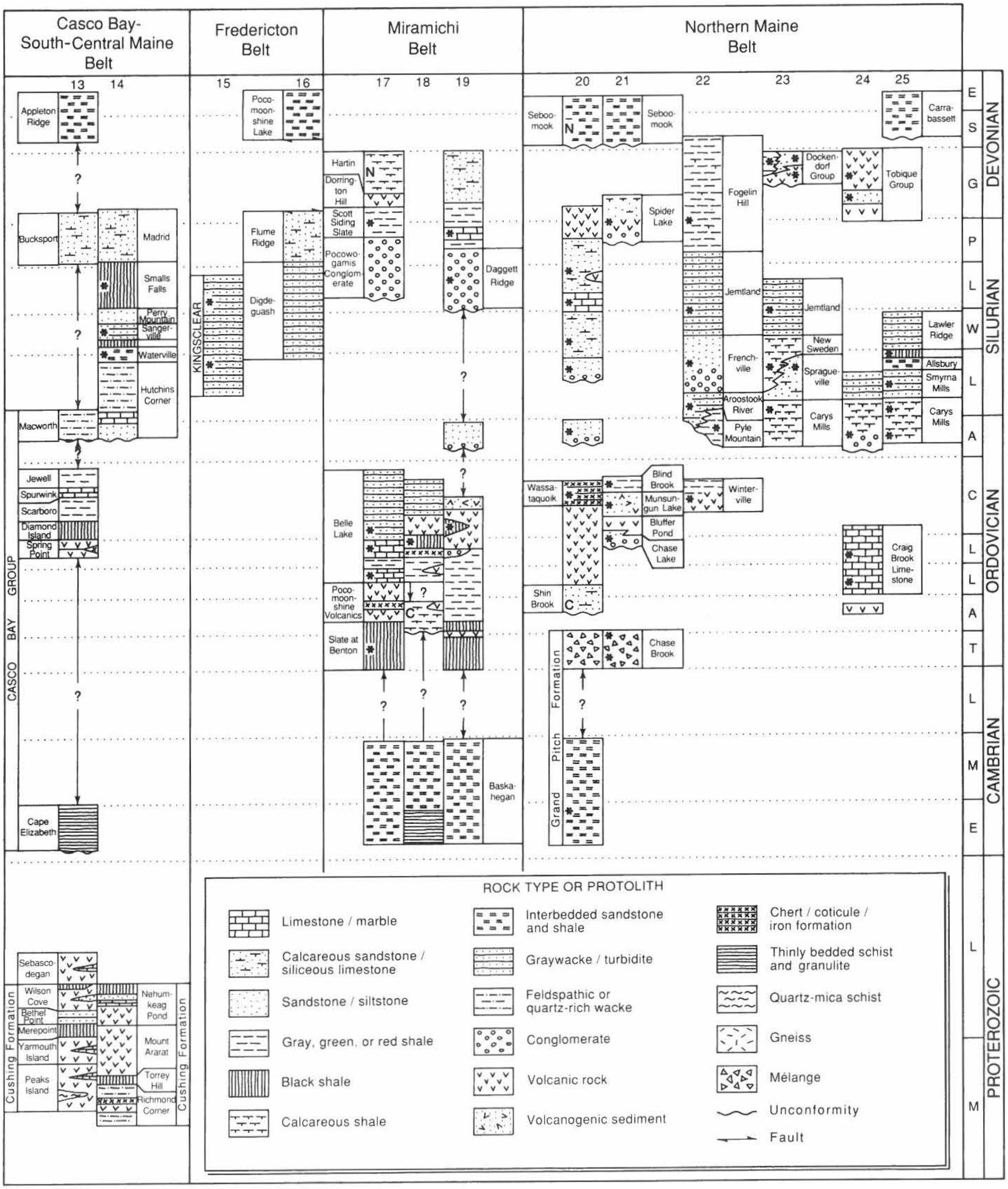

Figure 2. Continued.

McCutcheon (1981), Osberg et al. (1985), Fyffe and Fricker (1987), and Ludman (1987). The Saint John, Islesboro-Rockport, and Benner Hill belts contain only pre-Silurian strata, while the Ellsworth, St. Croix, Casco Bay - south-central Maine, Miramichi, and northern Maine belts include pre-Silurian rocks together with Late Ordovician to Early Devonian cover sequences.

The metamorphic grade over most of the region is greenschist, with lower grades to the north and east. West of Penobscot Bay, in the Islesboro-Rockport, Benner Hill, and Casco Bay - south-central Maine belts, the intensity increases to the upper amphibolite facies (see Guidotti, 1989). To emphasize stratigraphic relationships, some metamorphosed rocks are described here in terms of their sedimentary protoliths where they are easily inferred.

Figure 2, which graphically portrays the stratigraphic columns, is designed to accompany the following descriptions. Names of places and structural features cited in the text are shown in Figure 3.

Saint John Belt

Location. The Saint John belt lies in southern New Brunswick along the Bay of Fundy coast (Fig. 1). Carboniferous conglomerates and sandstones rest unconformably along its eastern margin. Conglomerates, sandstones, and shale of Carboniferous age, and volcanic rocks of uncertain age are in thrust contact with part of its southern border. On the northwest, it is bounded partly by the Belleisle fault, and partly by a fault slice containing Silurian volcanic rocks southeast of the Belleisle fault (Fig. 3).

Stratigraphic Relationships. Precambrian platformal sediments (Fig. 2, col. 1) of the Green Head Group (Hayes and Howell, 1937; Alcock, 1938) are divided into the Ashburn Formation, dominated by carbonates with lesser quartzites and argillites, and the Martinon Formation, consisting of thick calcareous and quartzose conglomerates overlain by interbedded argillite and quartzite (Ruitenberg et al., 1979). Stromatolites from the Ashburn Formation suggest an approximate age of Neohelikian (Hofmann, 1974). Recent investigations suggest an unconformity may separate the Martinon from the Ashburn (McCutcheon and Ruitenberg, 1987). The Green Head Group is more deformed and considered older than the Coldbrook Group (Alcock, 1938), although the inferred unconformity is everywhere faulted and has not been observed (Ruitenberg et al., 1979).

The Coldbrook Group (Fig. 2, col. 1) can be divided into three northeast-trending belts, each with its own internal lateral and vertical facies changes, described in detail by Ruitenberg et al. (1979). A bimodal suite of mafic and felsic volcanic flows, crystal and lithic tuffs, and volcanic breccias dominates the western and central belts, whereas the eastern belt is composed mostly of volcanogenic sedimentary rocks and water-laid tuffs with lesser intercalated flows. The Coldbrook Group is thought to be Hadrynian, from field relations and a few wide-ranging isotopic age determinations (Cormier, 1969; Stukas, 1977; Olszewski and Gaudette, 1982).

Fossiliferous Cambrian through Early Ordovician rocks of the Saint John area have been studied for over a century (Matthew, 1863, 1897, 1898; Bailey and Matthew, 1872; McLearn, 1915; and many others). Stratigraphy of the Saint John Group was presented by Hayes and Howell (1937) and Alcock (1938) and revised by Pickerill and Tanoli (1985). An Eocambrian sequence of conglomeratic redbeds containing feldspathic and volcanic clasts has recently been identified at the bottom of the section (Currie, 1984; Tanoli et al., 1985). The Saint John Group rests disconformably across the Eocambrian sediments and the Coldbrook Group. The Lower Cambrian Ratcliffe Brook Formation (Fig. 2, col. 1) consists of a basal pebble conglomerate overlain by interbedded maroon and green sandstones, shales, and conglomerate. Next, the Glen Falls Formation is divided into a lower coarse grained, locally conglomeratic white sandstone and an upper, thin, coarse grained, black sandstone. Above the Glen Falls Formation are thickly bedded gray sandstones and brown-weathering gray shale of the Hanford Brook Formation.

At the base of the Middle Cambrian Forest Hills Formation (Fig. 2, col. 1) is a one- to two-meter thick siliceous limestone overlain by gray shales, most of which are "literally filled with fossils" (Hayes and Howell, 1937, p. 73). Thinly bedded, gritty black shale is higher in the formation, and the upper part consists of dark gray shales, thin sandstone beds, and thinly bedded black shales with dark gray limestone lenses.

The Upper Cambrian Kings Square Formation (Fig. 2, col. 1) is a thick succession of interbedded gray shales and massive sandstones, commonly with scattered limestone lenses and nodules. The Silver Falls Formation, also Upper Cambrian, consists of dark gray to black shales with some interbedded thin sandstones and limestone concretions.

Early Ordovician rocks of the Reversing Falls Formation (Fig. 2, col. 1) are thinly bedded, sulfidic, carbonaceous, black shales with Tremadocian and Arenigian graptolites.

No strata between Arenigian and Carboniferous age are known from the Saint John belt.

Ellsworth Belt

Location. In New Brunswick the southeastern boundary of the Ellsworth belt (Fig. 1) corresponds to the Belleisle fault except near Long Reach (Fig. 3), where a slice of Silurian volcanic rocks assigned to the Ellsworth belt lies east of the Belleisle fault. To the southwest, the Ellsworth belt extends under Penobscot Bay. Its northwestern boundary is the west contact of the Queen Brook Formation in New Brunswick and several faults to the southwest which have been connected as the Graham Lake-Rockland line (Osberg et al., in press). The New

Brunswick portion of the belt corresponds approximately with the Mascarene terrane of Fyffe and Fricker (1987).

Stratigraphic Relationships. Felsic volcanic rocks without associated sedimentary rocks, located west of Sagwa, New Brunswick (Fig. 3), have been assigned a Precambrian age by MacKenzie (1964) and McCutcheon (1981), and similar rocks on Campobello Island (Fig. 3) have been equated to the Coldbrook Group (Fig. 2, col. 3) (McLeod and Rast, 1988). The relationships of these volcanic rocks to adjacent rocks is uncertain, and their assignment to the Precambrian is based solely on lithic similarity to volcanic rocks of the Coldbrook Group in the neighboring Saint John belt.

The North Haven Formation (Smith et al., 1907; Dow, 1965), exposed on North Haven Island in Penobscot Bay, consists of a monotonous section of pillowed, phyllitic greenstone, and minor siliceous tuff, breccia, and pelite (Fig. 2, col. 6; Pinette and Osberg, 1989). Greenstones located near West Gouldsboro (Fig. 3) may be equivalent to the North Haven Formation. These rocks are considered to be pre-Silurian because they locally lie beneath a Silurian unconformity and, in general, their degree of deformation is greater than that in the overlying Silurian volcanic rocks (Brookins et al., 1973). They are undated and could be as old as Precambrian, perhaps related to the Coldbrook Group. On the other hand, Brookins et al. (1973) considered at least part of the North Haven Formation to be equivalent to the Ellsworth Formation, which they considered to be Ordovician. These possible relationships have not been tested by mapping. A third alternative that is consistent with their chemistry is that the North Haven Formation represents a remnant of Late Ordovician to earliest Silurian oceanic crust.

Red and gray sandstones and conglomerates (Fig. 2, col. 2) are exposed on the east side of the Ellsworth belt in New Brunswick, between the Belleisle and Wheaton Brook faults (Fig. 3). The trilobites *Strenuella* and *Harttella* in sandstones fix their age as Early Cambrian (Yoon, 1970; McCutcheon, 1981). McCutcheon (1981) and McCutcheon and Ruitenberg (1987) have correlated this Cambrian section with nearby interbedded red sandstones and felsic and mafic volcanic rocks, although Tanoli et al. (1985) have suggested that some of the latter may be Eocambrian. The Cambrian stratigraphic section in southern New Brunswick is incomplete and fragmented by faulting, and its base is not known. Locally, a Cambrian porphyry is interpreted to intrude foliated granite, suggesting that the granite is Precambrian and may possibly represent the basement upon which the Cambrian section lies.

The Ellsworth Formation (Smith et al., 1907), the schist of Columbia Falls (Terzaghi, 1946; Gilman, 1961), and the Queen Brook Formation (Ruitenberg, 1967) are interpreted here as equivalents (Fig. 2, cols. 2, 3, 4, 5). They consist of thinly interbedded grayish green phyllite and white to gray siliceous, feldspathic siltstone. All three units are locally cut by mafic dikes, and thin units of felsic and mafic volcanic rocks, tuffs, and pyroclastic rocks have been mapped within the Ellsworth Formation by McGregor (1965).

The age of the Ellsworth Formation, the schist of Columbia Falls, and the Queen Brook Formation is uncertain. The Ellsworth Formation is unconformably overlain by Silurian-Lower Devonian rocks at several localities. Bouley et al. (1981) have questioned the regional significance of these unconformities and proposed that they may represent a limited time span. Stewart and Wones (1974) presented evidence that the Ellsworth Formation was metamorphosed prior to deposition of the Silurian-Lower Devonian Castine volcanic rocks and so must be pre-Silurian. The Queen Brook Formation also lies below Silurian volcanic rocks suggesting that its age is Early Silurian or Ordovician (Ruitenberg and Ludman, 1978; McCutcheon, 1981; McCutcheon and Boucot, 1984). Layering in the Queen Brook Formation is parallel to that in the overlying volcanic rocks, but at one locality a conglomerate occurs along the contact (McCutcheon, 1981), raising the possibility that it may be disconformable. A lower age limit for the Ellsworth and Queen Brook Formations can be given only if the interbedded nature of the contact between the Ellsworth and Queen Brook Formations, and the Calais Formation and its equivalents observed by McGregor (1965) and McCutcheon and Ruitenberg (1987) are valid. These relationships would suggest a post-Tremadocian age. A Rb-Sr whole rock isochron based on 5 felsite and 7 greenschist samples from the Ellsworth Formation give a revised age using the new decay constant of 500±20 Ma (Brookins, 1976), consistent with an Early Ordovician or older age.

Silurian and Devonian strata of the Ellsworth belt are part of the regional coastal volcanic belt (Boucot, 1969; Gates, 1969) and in New Brunswick comprise the Mascarene-Nerepis zone of Ruitenberg et al. (1977). The volcanic and sedimentary units have different names in different places, but paleontologic and isotopic work has been quite successful in demonstrating the complicated facies relationships among them. Unconformities at the base of the section are exposed in several places and range in age from earliest Silurian to Late Silurian (Fig. 2).

The Quoddy Formation (Fig. 2, col. 3) near Eastport, Maine (Fig. 3) consists of rusty-weathering, sulfide-rich, black siltstone and siliceous argillite, interbedded with minor volcanic ash (Bastin and Williams, 1914; Gates, 1969, 1975, 1978). Graptolites indicate a late Llandoverian age (Gates, 1975). The Quoddy Formation is separated by faults from the rest of the Eastport sequence, but it is interpreted to be the oldest unit in the Silurian-Early Devonian section because of its fossil age. McLeod and Rast (1988) have interpreted an unconformity on Campobello Island (Fig. 3) between the Quoddy Formation and volcanic rocks that they equate on lithologic grounds to the Coldbrook Group.

Unnamed gray polymictic conglomerate near Sagwa, New Brunswick (Fig. 3), is interpreted by McCutcheon (1981) to unconformably overlie Cambrian sandstones (Fig. 2, col. 2). The conglomerate is clast-supported, with clasts of thinly bedded, grayish green, siliceous siltstone having been derived from the Coldbrook Group. This conglomerate is gradationally overlain by gray sandstone and then by gray to black siltstone.

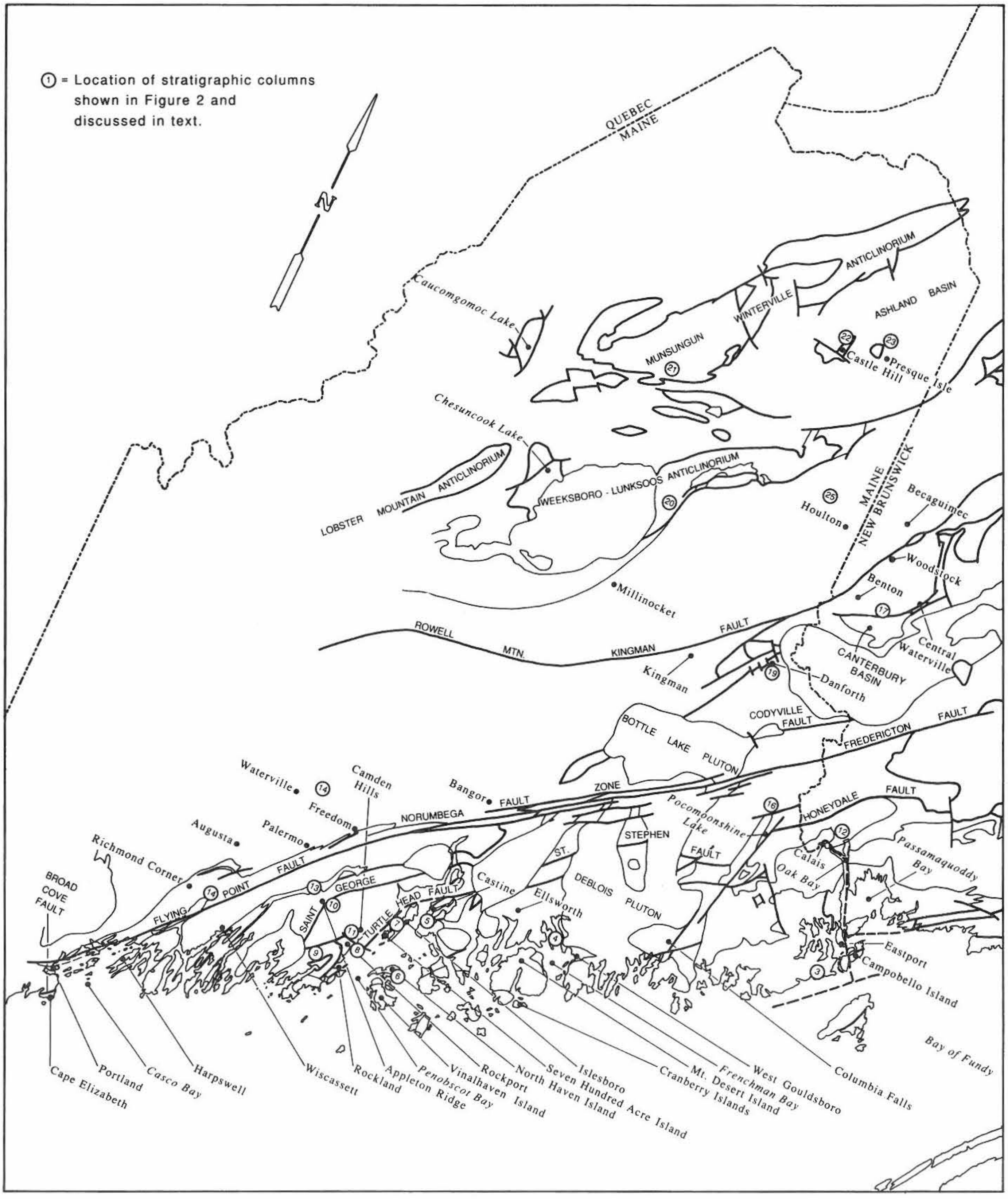

Figure 3. Geologic outline map, as in Figure 1, showing place names, geological features, and locations of stratigraphic sections (circled numbers) illustrated in Figure 2.

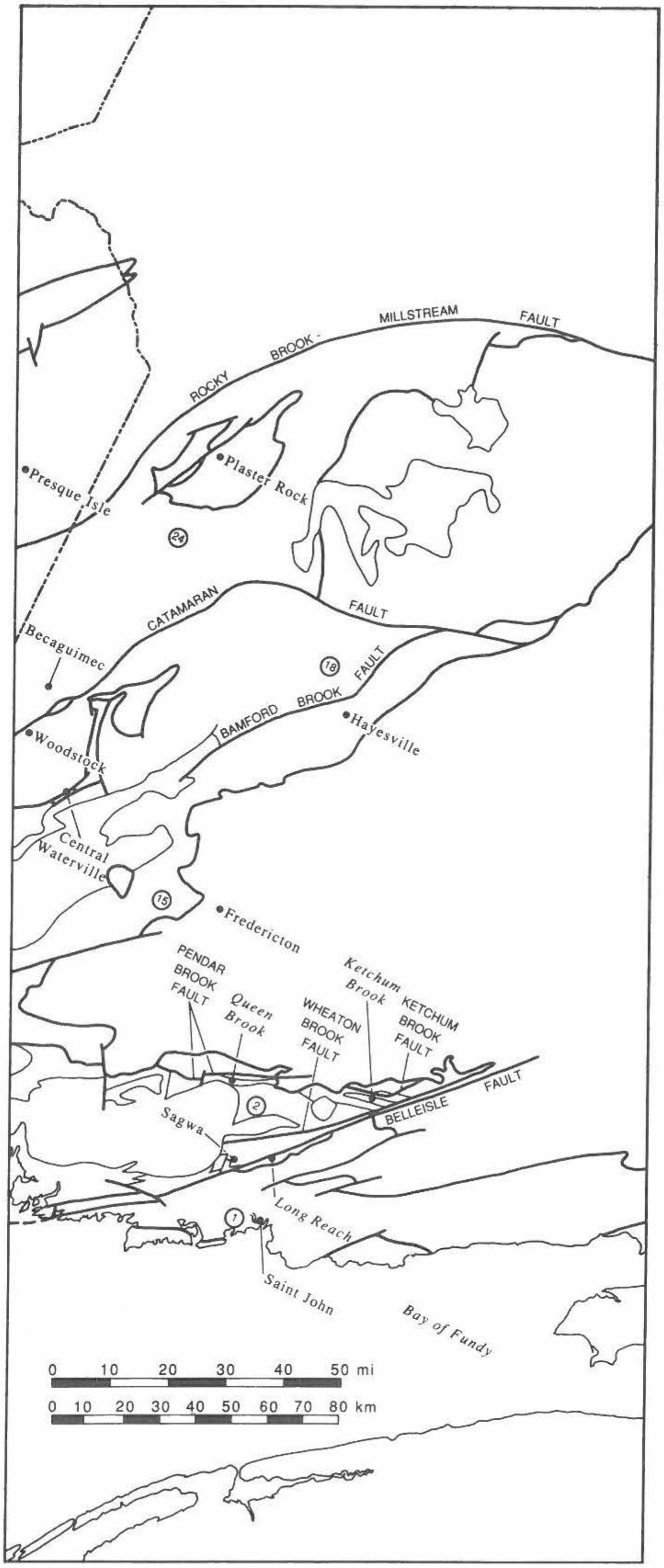

Figure 3. Continued.

Similar gray to greenish black sandstones and siltstones are present at Ketchum Brook and also near the eastern end of the Ellsworth belt. This eastern exposure contains Llandoverian (C5) brachiopods (McCutcheon and Ruitenberg, 1987). Because of their similarity in lithology and age, we equate the rocks in these three exposures with the Quoddy Formation. The western contact of the Queen Brook Formation is gradational with bedded black sandstones and siltstones of uncertain age along the Pendar Brook fault (Fig. 3). These might also correlate with the Quoddy Formation, but we have tentatively interpreted them as equivalent to the Calais Formation of the St. Croix belt (Fig. 1), after Ruitenberg and Ludman (1978).

The Ames Knob Formation (Fig. 2, col. 6), in southern Penobscot Bay (Fig. 3), is a thin unit consisting of a basal conglomerate overlain by calcareous sandstone, and then by gray, maroon, and green shales. The conglomerate contains randomly oriented, foliated clasts of the underlying North Haven Formation, suggesting that the greenstone was metamorphosed and deformed before deposition of the Ames Knob sediments. A diverse fauna in the Ames Knob Formation indicates late Llandoverian through Pridolian ages (Smith et al., 1907; Brookins et al., 1973).

The Bar Harbor Formation (Fig. 2, col. 4) contains an assemblage of siltstones, sandstones, and pebble conglomerates displaying even and regular bedding. In Frenchman Bay and on the mainland near West Gouldsboro (Fig. 3), basal conglomerates in the Bar Harbor Formation contain clasts of the underlying greenstone, suggesting an unconformable relationship (Metzger, 1959; Metzger and Bickford, 1972; Gilman, 1984; Gilman and Lash, 1988). Toward the top of the formation, layers of tuff and locally thick sequences of siliceous volcanic rocks are interbedded with the siltstones. No fossils have been reported from the Bar Harbor Formation, but a Rb-Sr whole rock age from the siliceous volcanic rocks suggests that the bulk of the formation predates Late Silurian to earliest Devonian (Metzger et al., 1982). It is tempting to interpret the unconformity at the base of the Bar Harbor Formation as basal Silurian, but this is not altogether clear. It could be entirely within the Silurian section, or it could transgress the lower part of the Silurian section and locally rest on pre-Silurian rocks.

The Long Reach and Jones Creek Formations make up most of the New Brunswick part of the Ellsworth belt (Fig. 2, col. 2). The Long Reach Formation (MacKenzie, 1964; McCutcheon, 1981), contains mainly mafic and lesser felsic volcanic rocks and tuffs. Parallel-laminated, fine-grained sandstone and minor limestone are intercalated with the volcanic rocks. The Long Reach Formation is interpreted to be gradational with the underlying rocks at Ketchum Brook and to be partly in facies relationship with them (McCutcheon and Ruitenberg, 1987). We take the position suggested by McCutcheon and Ruitenberg (1987) that the rocks at Ketchum Brook lie disconformably on the Queen Brook Formation. Where the Long Reach Formation lies directly on the Queen Brook Formation, the contact is interpreted here as a disconformity or a fault. A brachiopod fauna from limestone in the middle part of the Long Reach Formation indicates a Late Llandoverian (C6) to Wenlockian age (Boucot et al., 1966; McCutcheon and Boucot, 1984). The Jones Creek Formation conformably overlies the Long Reach Formation (McCutcheon, 1981). It is composed of thinly interbedded greenish gray sandstone and siltstone with minor amounts of felsic and basaltic flows and tuffs. Fossils in the upper part of the Jones Creek Formation are Pridolian in age (Berry and Boucot, 1970).

Near Eastport (Fig. 3), the volcanic section (Fig. 2, col. 3) includes the Dennys, Edmunds, Leighton, Hersey, and Eastport Formations (Bastin and Williams, 1914; Gates, 1969, 1975, 1978). The Dennys Formation is dominated by basalts, siliceous volcanic rocks, and volcaniclastic rocks with subordinate interbedded argillites containing a Wenlockian brachiopod fauna. The overlying Edmunds Formation is largely dacite and rhyolite, but contains significant amounts of volcanic breccia, tuff, and volcanic conglomerate. Basalts make up a small part of the formation. Black shale is minor but contains an abundance of Ludlovian fossils. The Leighton Formation is dominated by sedimentary rocks, mostly gray siltstone and shale, which are locally limey. Basaltic flows, rhyolitic debris flows, and tuffs make up the rest of the formation. A shallow-water fauna indicates the Leighton Formation is Pridolian. Above this, the Hersey Formation of maroon siltstone and shale contains shallow-water fossils which straddle the Silurian-Devonian boundary (Fig. 2, col. 3). At the top of the section is the Eastport Formation (Bastin and Williams, 1914), a volcanic unit dominated by basaltic and rhyolitic flows and tuff-breccias, with lesser tuffs, siltstones, shales, cherty limestone, and conglomerates. Some of the flows and tuffs are thought to be subaerial, and many of the sedimentary rocks are interpreted to have been deposited in shallow water. A brackish-water fossil assemblage indicates a Gedinnian age.

The Cranberry Island Formation (Fig. 2, col. 4), exposed on Little and Great Cranberry Islands near Mt. Desert Island (Fig. 3), consists of well layered tuffs, interbedded with felsic and intermediate volcanic flows and agglomerates. Mafic volcanic rocks make up a small part of the formation. An Early Devonian age is suggested by a Rb-Sr whole rock isochron of 379±9 Ma (Brookins et al., 1973).

The Castine Formation (Fig. 2, col. 5), along the eastern part of Penobscot Bay (Fig. 3) consists predominantly of siliceous volcanic rocks with abundant agglomerates, tuffs, and ash, and lesser mafic volcanic rocks, commonly pillowed (Smith et al., 1907). Sandstones, shales, and lenticular limestone form a decidedly minor part of the formation. Ostracodes from sedimentary layers and Rb-Sr whole rock ages from the volcanic rocks demonstrate Pridolian through Early Devonian ages (Brookins et al., 1973). Along Bagaduce Narrows, north of the town of Castine (Fig. 3), a thin sedimentary section at the base of the Castine Formation rests unconformably on Ellsworth Formation and contains clasts of that formation (Wingard, 1961; Stewart and Wones, 1974).

The Thorofare andesite and Vinalhaven rhyolite (Fig. 2, col. 6) overlie the Ames Knob Formation in Penobscot Bay. The Thorofare andesite contains andesite, rhyolite, tuff, and volcanogenic sedimentary rocks (Smith et al., 1907; Dow, 1965), and the Vinalhaven rhyolite consists of locally spherulitic rhyolitic flows, tuffs, and breccias (Smith et al., 1907). These rocks overlie greenish shales containing Pridolian fossils. Brookins et al. (1973) report a Rb/Sr whole rock isochron for the Vinalhaven rhyolite of 385±6 Ma (revised), consistent with an Early Devonian age.

Islesboro-Rockport Belt

Location. Rocks of the Islesboro-Rockport belt are exposed in three areas (Fig. 1). One is in West Penobscot Bay on Islesboro and several smaller islands to the south (Fig. 3). The second is on the mainland around Rockport, and the third is in a small thrust-bounded window just south of Rockport.

Stratigraphic Relationships. The area of the Islesboro-Rockport belt centered on Islesboro and Seven Hundred Acre Island is in a faulted anticlinal inlier nearly completely surrounded by Silurian-Devonian rocks of the Ellsworth belt. Rocks of this exposure belong to three sequences separated by unconformities (Fig. 2, col. 7). The oldest sequence, exposed on the eastern side of Seven Hundred Acre Island, on Spruce Island, and on other small islands, includes greenstone with prominent white quartzo-feldspathic layers, coarse muscovite-staurolite-andalusite-garnet schist, feldspathic biotite gneiss, and marble with lenticular amphibolite (Stewart, 1974; Stewart and Lux, 1988). These rocks were originally metamorphosed to amphibolite facies in the Precambrian, and have been retrograded. $^{40}Ar/^{39}Ar$ release spectra for hornblende separates yield apparent ages of about 670 and 650 Ma which are thought to date this metamorphic event (Stewart and Lux, 1988). A cross-cutting muscovite-bearing pegmatite in marble from Spruce Island yielded a Rb/Sr whole rock age of 620±20 Ma and a K/Ar muscovite age of 594±18 Ma (Brookins, 1976), establishing a Late Precambrian age for the pegmatite (Stewart, 1974).

The Islesboro sequence (Fig. 2, col. 7) is thought to rest unconformably on the retrograded Precambrian rocks because, in contrast to them, it was never metamorphosed above greenschist facies (Stewart, 1974; Stewart and Lux, 1988). The Islesboro sequence consists of an unnamed basal white quartzite which grades by interbedding into gray and green, thinly laminated feldspathic slate, containing two prominent, mappable, thin units of limestone within its lower part (Smith et al., 1907; Stewart, 1974). The Coombs Limestone (Smith et al., 1907), consisting of thinly bedded limestone, buff marble, and minor limestone conglomerate, overlies the slate. The Coombs Limestone is in turn overlain by white quartzite and quartzite conglomerate at the top of the Islesboro sequence. Since the Islesboro sequence is inferred to be younger than the Precambrian amphibolite-grade metamorphism and muscovite pegmatite, it might be latest Precambrian(?) or Cambrian(?).

We interpret a third sequence to be present in this same area on Islesboro and the western part of Seven Hundred Acre Island (Fig. 3). At its base, an unnamed polymictic pebble conglomerate and gray quartzite rest with profound unconformity across both the Islesboro sequence and the sequence of retrograded Precambrian rocks. The conglomerate and quartzite grade upward into unnamed thinly interbedded greenish gray chlorite-mica phyllite and sandstone. Because this third sequence is younger than the Islesboro sequence, it is presumably early Paleozoic, possibly Cambrian(?).

Rocks similar to those of the younger two sequences on Islesboro, but at higher metamorphic grade, are present on the mainland in the vicinity of Rockport (Fig. 3). Rocks of the Rockport sequence (Fig. 2, col. 8) were mapped and described by Osberg and Guidotti (1974), but the stratigraphic sequence is reinterpreted here. The oldest unit in the Rockport sequence is andalusite-mica schist with thin coticule beds, beds of sandstone, and lenses of polymictic conglomerate with clasts up to 15 cm across. A thin buff-colored marble lies near the bottom of the exposed section, and another thin white marble, barely exposed at low tide, is above this schist unit. Next is a sandy, gray schist that weathers to a woody texture. Discontinuous white quartzite beds and green calc-silicate pods occur near its base. The fourth unit consists of well bedded marble, thinly bedded calc-silicate marble, and limestone conglomerate like the Coombs Limestone of Islesboro (Smith et al., 1907; Bastin, 1908; Osberg and Guidotti, 1974). The limestones are overlain by white quartzite and quartzite conglomerate which correspond to similar rocks at the top of the Islesboro sequence.

Resting unconformably on these units in Rockport are metamorphosed polymictic pebble conglomerate and dark gray, biotitic sandstone which grade upward into andalusite-mica schist. It is proposed that the rocks above the unconformity correlate with the youngest sequence on Islesboro (Fig. 2, cols. 7 and 8).

The sections at Islesboro and Rockport are separated by the Turtle Head fault zone (Fig. 3) (Stewart, 1974; Stewart and Wones, 1974; Stewart, 1986). Stewart (1974) postulated very substantial motion along this fault zone because "the stratigraphic sections exposed on Islesboro have so few similarities with the stratigraphic columns in adjacent structural blocks." As indicated in the preceding paragraphs, however, we believe there are close similarities between the two sections. Even though the Turtle Head fault zone juxtaposes chlorite-grade rocks on the east against coarse-grained andalusite schist on the west (Stewart, 1974; Berry, 1986; Guidotti, 1989), in our view it does not separate different stratigraphic sequences, but two parts of the same sequence metamorphosed to different grade.

Southwest of Rockport (Fig. 3), the third exposure of the Islesboro-Rockport belt is surrounded by rocks of the Benner Hill sequence (Fig. 1). This small exposure is interpreted to be in a thrust-bounded window (Osberg and Guidotti, 1974). It contains rocks similar to those exposed at Rockport (Bastin, 1908; Osberg and Guidotti, 1974).

Benner Hill Belt

Location. The Benner Hill belt occupies part of the mainland west of Penobscot Bay and south of Rockport (Fig. 1). A high-angle fault separates it from the St. Croix belt to the west, and it is juxtaposed against the Islesboro-Rockport belt along thrust faults.

Statigraphic Relationships. The Benner Hill sequence (Osberg and Guidotti, 1974; Guidotti, 1979) can be divided into three units (Fig. 2, col. 9). The lower unit, informally called "ribby" schist, consists of gray quartz-mica schist with ribs of thinly interbedded feldspathic quartzite. The middle unit, biotitic quartzite and quartz-feldspar-mica grit, has a siliceous coquinite at the top containing abundant deformed Caradocian brachiopods (Boucot et al., 1972; Neuman, 1973; Boucot, 1973). The upper unit, including the schist at Benner Hill, is variably light to dark gray quartz-mica schist with distinctive thin intercalated garnetiferous quartzite beds. A rusty part of this unit (Fig. 2, col. 9) has been mapped separately and informally called the Prison Farm lithology (Osberg and Guidotti, 1974; Guidotti, 1979; Hussey, 1985).

St. Croix Belt

Location. The St. Croix b elt (Fig.1) is in fault contact with the Benner Hill and Islesboro-Rockport belts, but its contact with the Ellsworth belt, although partly faulted, has been reported to be a sedimentary contact north of Ellsworth (McGregor, 1965) and in southwestern New Brunswick (McCutcheon and Ruitenberg, 1987; Fig. 2). Its western contact is with Silurian to Lower Devonian rocks of the Casco Bay - south-central Maine and Fredericton belts along the Saint George-St. Stephen fault in Maine (Bickel, 1976; Osberg, 1978; Loiselle and Wones, 1980) and along the Honeydale (Ruitenberg, 1967) and Pendar Brook faults (McCutcheon, 1981) in New Brunswick (Fig. 3). A small exposure faulted against the east side of the Islesboro-Rockport belt near Rockland (Fig. 3) is also included in this belt (Osberg and Guidotti, 1974).

Stratigraphic Relationships. The Jam Brook Formation (Fig. 2. col. 10), exposed along the margin of the belt west of the Camden Hills (Fig. 3), contains thin units of light gray marble, quartz-rich mica schist, white quartzite, polymictic conglomerate and mica schist, and andalusite-quartz-mica schist (Bickel, 1971, 1976). There is no direct age control for the Jam Brook Formation.

The Megunticook Formation (Fig. 2, col. 11), well exposed in the Camden Hills (Fig. 3), consists predominantly of massive, light gray, andalusite-quartz-mica schist, with scattered beds and thin units of quartzite, coticule, and rusty weathering, dark gray, mica schist (Bickel, 1971; Berry, 1986). The base of the formation is faulted against the Islesboro-Rockport belt. The Megunticook Formation contains within it a tightly bound stratigraphy that has the Battie Quartzite member, formerly the Battie Quartzite of Bastin (1908), as its oldest unit. The Battie Quartzite consists of massive white quartzite and massive pebble to cobble conglomerate with subangular quartzite clasts supported by a biotitic quartzite matrix. Higher in the Megunticook section is a unit of brown-weathering, graphitic mica schist, overlain directly by discontinuous, thin lenses of marble and then tremolite- and actinolite-bearing quartzites. The quartzite unit is overlain in turn by a unit of black, graphitic, chiastolite-biotite schist with graded beds near its upper contact which set the local stratigraphic younging sense. Finally, a discontinuous, white marble unit locally occupies the contact between the Megunticook Formation and the base of the Penobscot Formation. The Megunticook Formation is of unknown age, but is thought to be Cambrian(?) because it underlies the Penobscot Formation, which can be traced into the Tremadocian Calais Formation (see below).

The lithologic similarity between andalusite schists in the Megunticook Formation and the andalusite schist in the upper part of the Jam Brook Formation makes it tempting to equate them. If such a correlation is valid, then the polymictic conglomerate and andalusite schist at the top of the Jam Brook Formation could represent the basal part of the Megunticook Formation, which is faulted out of the Camden Hills section (Fig. 2, col. 11). Likewise, the upper part of the Megunticook Formation, containing the several units described above, is missing from the Jam Brook section and is presumably cut out of that section by a fault (Fig. 2, col. 10; Bickel, 1976). This interpretation suggests the lower part of the Jam Brook Formation underlies the Megunticook Formation and may be Precambrian(?).

In the Big Lake quadrangle, Maine (Fig. 2, col. 12), Ludman (1985b, 1987) has assigned the rocks previously mapped as the Cookson Formation (Ruitenberg, 1967; Ruitenberg and Ludman, 1978) to the Cookson Group which contains two formations that could be correlative with the Megunticook Formation.[1] The Kendall Mountain Formation is dominated by massive quartzose, quartzo-feldspathic, lithic arenites, and conglomerate, similar to the Battie Quartzite, but it also contains in minor amount black shale, rhyolitic lava, and tuff. The overlying Woodland Formation[1] consists of quartzo-feldspathic wacke and gray shale in graded sets, somewhat reminiscent of the bulk of the Megunticook Formation.

Near Calais (Fig. 3), Ludman (1978, 1987) has shown that the Woodland Formation is overlain by the Calais Formation[1], which contains black, sulfidic shales and, in its upper part, layers of basalt and coticule. At Oak Bay, New Brunswick (Fig. 2, col. 12), equivalents of the Calais Formation (Cookson of Ruitenberg, 1967) contain graptolites first thought to be Arenigian (Cumming, 1967), but which are more likely Tremadocian (Ruitenberg and Ludman, 1978). Similar rusty-weathering, bedded, black sandstone, siltstone, and shale are present near Queen Brook in New Brunswick, between the Pendar Brook fault and the Queen Brook Formation (Fig. 3). Ruitenberg and Ludman (1978) correlated these rocks with the Calais Formation, but McCutcheon and Ruitenberg (1987) were uncertain, allowing that these rocks might alternatively rest above the Queen Brook Formation. The Calais Formation was mapped for 20 miles southwestward into Maine by Ruitenberg and Ludman (1978), and except for intervening plutons, the same lithofacies is traceable for nearly a hundred miles southwest into the Penobscot Formation, west of Penobscot Bay (Fig. 3).

The Penobscot Formation (Fig. 2, col. 11) conformably overlies the Megunticook Formation (Smith et al., 1907; Bickel, 1976; Berry, 1986). Its characteristic lithology is very rusty weathered, black, graphitic, and sulfidic mica schist. A thin unit of white marble is locally present at its base (Osberg and Guidotti, 1974), and subordinate quartzite is irregularly interbedded with the schist in the lower part of the formation. Higher in the formation is an amphibolite unit (Fig. 2, col. 10), the Gushee Member (Bickel, 1976), interpreted as metamorphosed basaltic volcanic rocks. Because of its lithologic correlation with the fossiliferous Calais Formation on Cookson Island, the Penobscot Formation is inferred to be Tremadocian.

At Oak Bay (Fig. 3), the Calais Formation (Fig. 2, col. 12) is unconformably overlain by the Silurian Oak Bay and Waweig Formations (Cumming, 1967; Ruitenberg, 1967; Ruitenberg and Ludman, 1978; Ludman, 1981). The Oak Bay Formation consists of polymictic pebble to boulder conglomerate which grades upward into graywacke. A limestone clast in the conglomerate yielded a late Llandoverian pentamerid brachiopod (Cumming, 1967) giving a maximum age of deposition for the conglomerate. The Oak Bay Formation grades into overlying fine-grained calcareous and non-calcareous sandstones and siltstones of the Waweig Formation. A Ludlovian to Pridolian fauna from the Waweig Formation gives a minimum age for the unconformity (Pickerill, 1976).

Ludman (1987) defined the Pocomoonshine Lake Formation, mapped along the western edge of the St. Croix belt near Pocomoonshine Lake (Fig. 3), and assigned it to the Cookson Group, stratigraphically below the Kendall Mountain Formation. However, because of its lithic similarity to the Seboomook Formation of northern Maine, we have placed the Pocomoonshine Lake Formation in a Silurian-Early Devonian sequence and will describe it with the rocks of the Fredericton belt (Fig. 2, col. 16).

Casco Bay - South-Central Maine Belt

Location. The Casco Bay - south-central Maine belt has three major components. An area of pre-Ashgillian(?) rocks extending from south of Portland northeastward to Bangor is interpreted to be in a faulted anticlinal inlier (Fig. 1). Ashgillian(?) to Silurian cover extends westward from this inlier into

[1] See Editors' Note regarding stratigraphic order in the Cookson Group.

south-central Maine. Fault-bounded rocks east of the inlier are lithologically similar to the Ashgillian(?) to Silurian cover sequence. To the northeast, the boundary of the Casco Bay - south-central Maine belt is arbitrarily placed at the Bottle Lake and Deblois plutons (Fig. 3).

Stratigraphic Relationships. The oldest rocks of the Casco Bay - south-central Maine belt are lumped into the Cushing Formation (Katz, 1917; Hussey, 1971, 1985, 1988). The many members of the Cushing Formation are separated by the Flying Point fault (Fig. 3) into the Falmouth-Brunswick sequence northwest of the fault, and the Saco-Harpswell sequence southeast of the fault. The relative age of the two sequences is not known.

The Peaks Island and Yarmouth Island Members of the Saco-Harpswell sequence (Fig. 2, col. 13) are characterized by light gray quartz-feldspar gneisses and amphibolites. Pyroclastic textures and blue quartz phenocrysts in the Peaks Island Member indicate a volcanic protolith. These two units also contain thin beds of marble, calc-silicate skarn, and schist. The Merepoint Member consists of sulfidic schist, and the Bethel Point Member has thinly interbedded quartzite and sulfidic schist. At the top of the Saco-Harpswell sequence (Hussey, 1988), the Wilson Cove Member is a heterogeneous assemblage of feldspathic garnet-biotite schist, garnetiferous quartzite, cummingtonite-garnet-hornblende-biotite-quartz-plagioclase gneiss, amphibolite, and very rusty biotite-muscovite schist.

The Sebascodegan Formation (Fig. 2, col. 13) in the Saco-Harpswell sequence may be a lateral facies of the Cushing Formation (Hussey, 1988). It is characterized by thinly bedded, light gray, quartz-plagioclase-biotite granulite with lesser amounts of interbedded calc-silicate granulite and amphibolite, some of which contains abundant anthophyllite and cummingtonite.

The Cushing Formation in the Falmouth-Brunswick sequence (Fig. 2, col. 14) has been mapped by Newberg (1984, 1985) and Hussey (1985, 1988) as the Richmond Corner, Torrey Hill, Mount Ararat, and Nehumkeag Pond Members. The Richmond Corner Member consists of garnet-magnetite-mica schist with spectacular beds of coticule and interlayered quartz-feldspar gneisses and amphibolite. The Torrey Hill Member consists of sulfidic mica schist containing scattered beds of rusty quartzite and having a rusty, buff colored marble at its upper contact. The Mount Ararat Member is dominated by light gray, quartz-feldspar gneisses with lesser interlayered amphibolite. The Nehumkeag Pond Member contains kyanite-staurolite-mica schist, rusty sulfidic mica schist, lineated, white quartzite and minor amphibolite, calc-silicate marbles and flaggy quartz-felspar gneisses, and minor amphibolite.

The ages and stratigraphic relationships of the members in the Cushing Formation are not well constrained. A Rb/Sr whole rock age from feldspathic gneisses near Portland (Fig. 3) has yielded 481±40 Ma (Brookins and Hussey, 1978), and a Rb/Sr whole rock age from the Mount Ararat Member to the north has given a somewhat similar age (W. J. Olszewski, Jr., pers. commun., 1982). Taken at face value, these Rb/Sr whole rock ages would indicate that the Cushing Formation is Ordovician or older, but it is likely that the isotopic systems in the protoliths were not homogeneous and, therefore, the required conditions for the method may never have been met. Tentatively, the Cushing Formation is regarded to be Ordovician or older, and it may be as old as Precambrian (Hussey, 1988).

The Cape Elizabeth Formation (Fig. 2, col. 13) overlies the Cushing Formation (Katz, 1917; Hussey, 1971, 1988). A local granule conglomerate at its base and map-scale truncation of underlying units argue for a low-angle unconformity beneath it (Hussey, 1971). The Cape Elizabeth Formation consists of 2 to 6 cm thick, interbedded light to dark gray, quartz-mica schist and metamorphosed quartz-rich and variably feldspathic sandstone. Some beds are graded, giving valuable younging sense near the unconformity. The Cape Elizabeth Formation is unfossiliferous, and its age is uncertain. It is older than the Ashgillian(?) to Silurian cover rocks that overlie the Casco Bay Group, so its age could be Ordovician or older.

The rocks exposed along the shore at Cape Elizabeth (Fig. 3) are, unfortunately, atypical of the Cape Elizabeth Formation as mapped for more than 50 miles to the northeast. At the type locality, the rocks are interbedded, buff-weathering, calcareous siltstones and dark gray phyllite in beds up to a meter thick (Hussey, 1985, 1986), which we believe are more like rocks of the nearby Merrimack Group to the southwest, or the Hutchins Corner Formation to the northwest, than to the rest of the Cape Elizabeth Formation. Such speculation about the regional correlation of the type Cape Elizabeth is rendered insignificant, however, by the post-metamorphic Broad Cove fault (Fig. 3) which isolates it from the rest of the mainland.

The upper part of the Casco Bay Group (Fig. 2, col. 13) consists of a conformable sequence of thin, distinctive units (Katz, 1917) which have only been identified southeast of the Flying Point fault (Fig. 3). Locally exposed contacts and primary facing indicators establish the relative age sequence of these units, as described in detail by Hussey (1971, 1985, 1988) and Swanson et al. (1986). The Spring Point Formation is dominated by metamorphosed mafic tuff, but near Harpswell (Fig. 3) it is divided into mafic lower and felsic upper parts, separated by a 15-m-thick coticule horizon (Swanson et al., 1986). The Spring Point Formation is overlain by a thin unit of massive, black, sulfidic phyllite, the Diamond Island Formation. The Scarboro and Jewell Formations both contain light gray quartz-mica schist and phyllite, and dark gray, rusty weathering, quartz-mica schist and phyllite. They are lithologically indistinguishable, but stratigraphically separated by the intervening Spurwink Limestone, containing fine-grained, ribbony-bedded gray limestone with interbedded phyllite. The age of the upper part of the Casco Bay Group is likewise fixed only by the Ashgillian(?) to Silurian age of the cover sequence, and, consequently, permissible ages are Ordovician or older.

In the vicinity of Bangor (Fig. 3), at the northeast end of the anticlinal inlier of older rocks, is a sequence of rocks called the

Passagassawakeag Gneiss (Bickel, 1976), including feldspar augen gneiss, migmatitic gneiss, amphibolite, calc-silicate rocks, and interlayered schist and feldspathic granulite, all at sillimanite+K-feldspar metamorphic grade. Bickel (1976) and Stewart and Wones (1974) thought these high-grade rocks comprised a sequence unrelated to surrounding rocks. Alternatively, we believe these rocks are deformed, high-grade equivalents of those along strike to the south in the Casco Bay Group. Although stratigraphic relationships at the northeast end of the inlier are known only in reconnaissance, the Copeland Formation (J. M. Trefethen, pers. commun., 1950) strongly resembles the Cape Elizabeth Formation and it is gradational into the sillimanite+K-feldspar gneisses of the Passagassawakeag. Calc-silicate marbles may equate with the Spurwink Formation, and amphibolites can be traced nearly continuously from the Spring Point Formation at Wiscasset into the Passagassawakeag Gneiss (Osberg et al., 1985).

The Ashgillian(?) through Silurian south-central Maine cover sequence (Fig. 1) unconformably overlies rocks of the Cushing Formation (Osberg, 1988) in the vicinity of Palermo and Freedom (Fig. 3). This conformable sequence consists of the Hutchins Corner through Madrid Formations (Fig. 2, col. 14). The Hutchins Corner Formation (Osberg, 1988), formerly part of the Vassalboro Formation, consists of thickly bedded quartz wackes and graywackes interbedded with less abundant phyllite. The basal part of the formation is slabby bedded feldspathic sandstone with thin partings of phyllite. At the top of the basal unit is a prominent gray limestone member. No fossils have been identified from the Hutchins Corner Formation, but stratigraphic interpretation suggests it may be Ashgillian(?) to Llandoverian.

The Waterville Formation (Fig. 2, col. 14) overlies the Hutchins Corner Formation, and consists of thinly interbedded greenish gray phyllite and light gray quartzite. A prominent gray limestone member is present in the middle of the formation. Dendroidal graptolites and trace fossils suggest a Wenlockian age (Osberg, 1968).

A thin, black, sulfidic phyllite overlies the Waterville Formation and forms the base of the Sangerville Formation (Fig. 2, col. 14). The bulk of the Sangerville Formation consists of thickly bedded graywacke, lesser phyllite, and minor conglomerate (Ludman, 1976). Gray, thinly bedded limestone forms an eminently traceable member within the Sangerville Formation. Graptolites at scattered localities suggest a late Llandoverian (Osberg, 1968) to Wenlockian age (Ludman, 1976).

The Perry Mountain Formation (Fig. 2, col. 14) is a thin unit which overlies the Sangerville Formation conformably and consists of cross-bedded quartzite in beds 8 to 15 cm thick, thinly interbedded sandstone and greenish gray phyllite, and contains minor pods of limestone near its upper boundary (Osberg et al., 1968; Osberg, 1988).

The Smalls Falls Formation (Fig. 2, col. 14) is dominated by rusty weathering, black, highly sulfidic phyllite (Osberg et al., 1968). Its lower part contains thick beds of white quartzite. Graptolites from several localities suggest a Ludlovian age (Pankiwskyj et al., 1976).

The Madrid Formation (Fig. 2, col. 14), conformably overlying the Smalls Falls Formation, consists of slightly calcareous, light gray quartz wacke and lesser pelite (Osberg et al., 1968). Beds of wacke range from 8 cm to 4 m thick. No fossils have been found in the Madrid Formation, but its stratigraphic position implies a probable Pridolian to possibly lowermost Gedinnian age.

The Macworth Formation (Katz, 1917) is thought to be equivalent to the lower part of the south-central Maine cover sequence (Hussey, 1988), although it originally was assigned to the Casco Bay Group. The Macworth Formation (Fig. 2, col. 13) is a thinly bedded to laminated, calcareous and quartzose phyllite and siltstone containing quartz and feldspar granules. It is exposed in a fault slice near Portland and in the core of a syncline in eastern Casco Bay (Fig. 3), where it overlies the Jewell Formation. Its upper contact is not preserved. The age of the Macworth Formation is based on its lithic similarity to the lower part of the south-central Maine sequence and to its position above the Jewell Formation.

The area southeast of the pre-Ashgillian(?) inlier and northwest of the St. Croix belt is underlain by the Bucksport and Appleton Ridge Formations (Wing, 1957; Stewart and Wones, 1974; Bickel, 1976; Hussey, 1985). Although separated by faults from the pre-Ashgillian rocks and the Ashgillian(?) to Early Devonian cover sequences, they are interpreted on lithologic grounds to be equivalent to the upper part of the cover sequence. The Bucksport Formation (Fig. 2, col. 13) is characterized by purplish gray, feldspathic wacke interbedded with green, calcareous wacke and minor gray biotite schist. It might correlate with the Madrid Formation to the west. The Appleton Ridge Formation (Fig. 2, col. 13) consists of well bedded, gray staurolite-mica schist with interbedded gray quartzite. On the basis of sparse graded beds, Bickel (1971) suggested the Appleton Ridge Formation is older than the Bucksport Formation, but this relationship was uncertain. We believe instead that the Appleton Ridge Formation may be younger than the Bucksport Formation and possibly lies above the Silurian rocks exposed in south-central Maine.

Fredericton Belt

Location. The Fredericton belt (Fig. 1) is located northwest of the St. Croix belt with which it is in fault contact. Its eastern contact is the Carboniferous unconformity, and its western contact is along the Bamford Brook and Codyville faults. The southwestern boundary of the Fredericton belt is placed arbitrarily at the Bottle Lake and Deblois plutons (Fig. 3).

Stratigraphic Relationships. As much as 3000 meters of turbiditic graywacke and slate (Poole et al., 1970) are exposed northwest of Fredericton, New Brunswick (Fig. 3). These rocks (Fig. 2, col. 15), originally mapped as the Kingsclear series

(Freeze, 1936) and informally named the Kingsclear Group (Williams et al., 1985), are chiefly made up of thick bedded graywacke and thinly interbedded gray, green, or red slate, locally displaying graded couplets. Sparse beds of feldspathic, pebbly sandstone are found at a few localities (Gordon, 1973). Graptolites from the Kingsclear Group collected by W. H. Poole are early Llandoverian (A. C. Lenz, written commun. to M. J. Copeland, 1982), and graptolites collected by A. Gordon include early Ludlovian forms (L. M. Cummings, written commun. to A. Gordon, 1966).

The rocks in the southern part of the Fredericton belt have been assigned to the Digdeguash and Flume Ridge Formations (Fig. 2, col. 16). The lower unit, the Digdeguash Formation (Ruitenberg, 1967), consists of thickly bedded, non-calcareous graywacke with interbedded gray phyllite and slate. No fossils have been recovered from this unit, but because of the lithic similarity to some of the graywackes and slates to the north, it is considered to be Silurian. The Flume Ridge Formation (Ruitenberg, 1972; Flume Formation of Ruitenberg, 1967) overlies the Digdeguash Formation. It consists of greenish gray, calcareous, micaceous wacke, siltstone, phyllite, and slate. Vascular plant fossils recovered from the Flume Ridge Formation (Ludman, 1986) are probably Silurian or Devonian, but do not indicate a precise age (A. Ludman, pers. commun., 1987). We equate these rocks with those exposed in the vicinity of Fredericton (Fig. 3).

The Pocomoonshine Lake Formation (Fig. 2, col. 16) is located along the southern margin of the Fredericton belt in eastern Maine (Fig. 3). It consists of buff to gray weathered, fine-grained sandstone and dark gray to black, generally carbonaceous shale and slate in beds 5 mm to 10 cm thick (Ludman, 1985b, 1987). Graded beds are common. Although Ludman (1985b) included the Pocomoonshine Lake Formation in a Cambrian(?) sequence, it is fault-bounded and, therefore, its stratigraphic position is uncertain. On the basis of its lithologic similarity to rocks in the Seboomook Formation of northern Maine, we tentatively interpret it to lie stratigraphically above the rest of the Fredericton belt and to have an age possibly as young as Siegenian.

Miramichi Belt

Location. The Miramichi belt (Fig. 1), exposed in a broad area of west-central New Brunswick and eastern Maine, is separated by the Bamford Brook-Codyville fault (Fig. 3) from the Ashgillian(?)-Early Devonian(?) rocks of the Fredericton belt. To the northwest, its boundary with the northern Maine belt follows the Mill Stream fault, an unnamed fault, and the Catamaran fault (Fig. 3). At its southern extremity and along its western margin, the Miramichi belt is juxtaposed against the Ashgillian(?)-Silurian cover sequence of the Casco Bay - south-central Maine belt by an unnamed thrust (Ludman, 1985a, pers. commun., 1987); along its southwestern margin, its contact with the cover sequence of the Casco Bay - south-central Maine belt may be sedimentary (Olson, 1972).

Stratigraphic Relations. We are principally concerned with that part of the Miramichi belt that is south of the Catamaran fault (Fig. 3). This part of the Miramichi belt is dominated by Cambrian(?) and Ordovician strata at low metamorphic grade. Although fossil age control has demonstrated complex facies distributions of the Ordovician strata, large packages within the established stratigraphic sections have similar lithologies and age constraints. In addition to the Cambrian(?)-Ordovician strata, unconformable Silurian-Lower Devonian cover rocks are locally present.

In the Benton, New Brunswick area (Fig. 3), Venugopal (1978, 1979) has mapped a sequence dominated by red, green, and gray chlorite-rich, feldspathic wacke and less abundant slate in beds 10 cm to 5 m thick, locally displaying some grades (Fig. 2, col. 17). This unit has not yielded fossils, but its position beneath rocks regarded as Tremadocian suggests a Cambrian(?) age. Similar rocks (Fig. 2, col. 18) exposed near Hayesville, New Brunswick (Fig. 3) are thought to be equivalent (Irrinki, 1980), and rocks near Danforth, Maine (Fig. 3), assigned to the Baskahegan Lake Formation (Fig. 2, col. 19) by Ludman (1985a), have similar lithology and are therefore thought to belong to the same Cambrian(?) package.

The Cambrian(?) sandstones and slates at Benton are overlain by a unit of rusty weathering, thinly interbedded dark gray shale and subordinate quartz-rich wacke (Fig. 2, col. 17). This lithology grades upward into thinly bedded, carbonaceous, sulfidic slate and gray-white weathering siltstone (Venugopal, 1978). Fyffe et al. (1983) report late Tremadocian graptolites from these slates. Ludman (1985a) describes similar carbonaceous, sulfidic shale to conformably overlie the Baskahegan Lake Formation near Danforth (Fig. 2, col. 19). The upper part of this unit is rusty weathering, sulfidic sandstone and slate in beds ranging from 2 to 30 cm thick. An unnamed unit of basaltic flows, tuffs, and diabase separates the lower black shale from the upper pyritiferous sandy unit. A thin unit of heavily pitted lenses of actinolite-plagioclase rock and calcite-epidote-actinolite rock is interstratified with the basalt. No comparable unit is present in the Hayesville section (Fig. 2, col. 18).

Near Hayesville, Irrinki (1980) has mapped calcareous slate with lenticular tuff beds (Fig. 2, col. 18), from which Neuman (1968) reported Arenigian brachiopods. The relationships to the underlying Cambrian(?) sandstones and slates have been interpreted to be conformable (Irrinki, 1980), but because of the apparent absence of Tremadocian black shale, a disconformity could be present.

Above the Arenigian slates at Hayesville (Fig. 2, col. 18) are red grits which at their base locally contain clasts of calcareous siltstone similar to the lithology of the underlying Arenigian unit, suggesting the possibility that the contact is disconformable (Irrinki, 1980). The grits grade upward through a meter or two into thick-bedded red and green slate, followed by gray, green, and red cherty siltstone and slate, and finally by black sulfidic slate with graptolites of the *Nemagraptus gracilis* zone (Irrinki, 1980). Locally, felsic and mafic volcanic rocks are interbedded with red, green, and black slates of this sequence. The Pocomoonshine Volcanics (Fig. 2, col. 17) described by Venugopal (1979) near Benton are equated to this sequence (Fyffe et al., 1983). The Pocomoonshine Volcanics contain felsic tuffs, thin mafic flows, red manganiferous siltstone, chert, and argillite. In the upper part of the unit are massive flows with minor tuff and breccia.

Ludman (1985a) and Sayres and Ludman (1985) have described somewhat similar rocks (Fig. 2, col. 19) from the Danforth area (Fig. 3). Gray slate with interbedded gray siltstone and wacke are overlain by local conglomerate. These slates are followed by a heterogeneous unit dominated by felsic volcanic rocks including aphanitic and porphyritic felsite, volcanic agglomerate, conglomerate, sandstone, slate, manganiferous siltstone, and iron formation. Graptolites indicating possible Caradocian age have been recovered from slates interbedded with the volcanic rocks (Larrabee et al., 1965), and a single brachiopod of "probable Ordovician age" (R. B. Neuman, pers. commun. to Glen Lutes, 1983) came from the same unit.

The Belle Lake Formation (Fig. 2, col. 17), overlying the Middle Ordovician volcanic rocks, consists of alternating dark gray to black slate and graywacke with beds typically 2 to 25 cm thick and up to 75 cm thick (Venugopal, 1978). Many of the wacke beds contain granules of quartz, feldspar, and volcanic rocks, and locally they are graded. Some beds are rusty-weathered and sooty, whereas others are clean and gray. The Belle Lake Formation near Benton (Fig. 3) has yielded upper Llandeilo-lower Caradoc (upper *gracilis* zone) graptolites (Fyffe et al., 1983), and near Central Waterville (Fig. 3), whitish gray, crystalline limestone occurring as lenticular beds in the Belle Lake Formation (Venugopal, 1979) has yielded conodonts indicating a late Llanvirnian age (Nowlan, 1979), somewhat older than the graptolites near Benton. At Hayesville (Fig. 3), a graywacke and slate unit, which overlies the black slates containing Llandeilian to lower Caradocian fossils, has been correlated with the Belle Lake Formation by Fyffe et al. (1983).

Unfossiliferous conglomerate, grit, and slate (Fig. 2, col. 19) overlying the volcanic section near Danforth (Fig. 3) are interpreted by A. Ludman (pers. commun., 1988) to be Ashgillian(?) on the basis that similar conglomerates are interbedded with the lower part of the Carys Mills Formation in the immediately adjacent Casco Bay - south-central Maine belt. The Carys Mills Formation contains fossils which indicate that its age ranges from Ashgillian to Llandoverian.

Silurian and Early Devonian rocks in the Canterbury basin and a smaller basin west of Danforth (Figs. 1, 3) lie unconformably on the Caradocian(?) and older sections described above. In the Canterbury basin, the basal unit is the Pocowogamis Conglomerate (Fig. 2, col. 17), containing pebbles derived from the underlying Cambrian-Ordovician quartzite and slate (Venugopal, 1979). The conglomerate is followed by gray slates with interbedded mafic volcanic rocks of the Scott Siding Slate, mafic to felsic volcanic rocks with minor sandstone of the Dorrington Hill Formation, and sandstone, calcareous slate, minor conglomerate, limestone, and volcanic rocks of the Hartin Formation (Venugopal, 1979). Fossils indicate the Scott Siding Slate may be upper Llandovery to Lower Devonian and the Hartin Formation is Gedinnian (Venugopal, 1979; A. J. Boucot, pers. commun., 1987). Equivalent rocks in the basin west of Danforth include the Daggett Ridge Formation (Fig. 2, col. 19) of sandstone and polymictic conglomerate with clasts containing Ludlovian fossils (Larrabee et al., 1965; Ludman, 1985a), and a suite of unnamed gray slates with minor conglomerate, sandstones, limestones, and calcareous slate that are tentatively equated with the Hartin Formation.

The Northern Maine Belt

Location. The northern Maine lithotectonic belt (Fig. 1) occupies northern and northeastern Maine and northwestern New Brunswick (Fig. 3). It is bounded to the east against the Miramichi belt by the Mill Stream fault, an unnamed fault, and by the Catamaran fault. It is in contact with the Casco Bay - south-central Maine belt along the Kingman and Rowell Mountain faults. The northwestern boundary of the northern Maine belt is beyond the area of this study.

Stratigraphic Relationships. In the northern Maine belt, pre-Ashgillian rocks are exposed in the Lobster Mountain, Weeksboro-Lunksoos, and Munsungun-Winterville anticlinoria and near Caucomgomoc Lake (Osberg et al., 1985). Smaller inliers of pre-Ashgillian rocks occur at Becaguimec, New Brunswick and at Castle Hill, Maine (Fig. 3). Because of our interest in the eastern part of this belt, the stratigraphic relationships for the pre-Ashgillian rocks will be taken principally from the Weeksboro-Lunksoos and Munsungun-Winterville anticlinoria and the small inliers, with only selected reference to the other areas. These sections of older rocks are unconformably overlain by diverse Ashgillian to Emsian cover sequences, which can be divided conveniently into thin and thick Ashgillian to Ludlovian cover sequences, Pridolian to Gedinnian volcanic sequences, and Siegenian to Emsian flysch.

Pre-Ashgillian Rocks. The Grand Pitch Formation (Neuman, 1967) is exposed in the core of the Weeksboro-Lunksoos anticlinorium (Fig. 2, col. 20). Most of the formation consists of variegated gray, green, and red slate and phyllite containing thin beds of quartzite and units of red and green wacke and slate. The trace fossil *Oldhamia smithi* Reudeman, present in red slate at several places, is the basis for assigning a Cambrian age to at least the lower part of the Grand Pitch Formation (Neuman, 1967; Neuman and Rankin, 1980). At Chesuncook Lake (Fig. 3), the variegated phyllite and quartzite lithology is overlain by grayish green siliceous wacke and minor slate.

A distinctive black slate and mélange unit (Fig. 2, col. 20) overlies the wacke and slate unit. This black slate and mélange unit is similar to the Hurricane Mountain Formation in the Lobster Mountain anticlinorium (Boone, 1983, this volume) and to the Chase Brook Formation (Fig. 2, col. 21) of the Munsungun-Winterville anticlinorium (Hall, 1970). The Chase Brook has yielded *Caryocaris* from black slate and an assortment of conodonts from a thinly bedded limestone, possibly a lithostrome (B. A. Hall, written commun., 1986). The *Caryocaris* may indicate a Tremadocian age (W. D. I. Rolfe, pers. commun. to R. B. Neuman, 1983), whereas the conodonts suggest an Arenigian age (J. E. Repetski, pers. commun. to B. A. Hall, 1983). If the correlation is correct, these ages from the Chase Brook Formation also apply to the uppermost unit of the Grand Pitch Formation.

The Shin Brook Formation (Fig. 2, col. 20) unconformably rests on the Grand Pitch Formation (Neuman, 1967). At the base of the Shin Brook Formation are thin, gray slate and phyllite of tuffaceous origin, overlain by approximately 50 m of conglomerate, tuffaceous sandstone, and volcanic conglomerate. The coarse-grained rocks are in turn overlain by a thick sequence of gray and green tuff, crystal tuff, and slightly calcareous, tuffaceous sandstone. Sparse volcanic flows of intermediate composition also occur in the formation, and the whole sequence is overlain by a thick pile of possibly intrusive diabase. The upper contact of the formation is unknown (Neuman, 1984). Sandstones of the Shin Brook Formation contain abundant brachiopods and fewer trilobites that suggest a late Arenigian age (Neuman, 1967, 1968, 1984; Neuman and Rankin, 1980). Rocks of the Shin Brook Formation are less deformed than those of the Grand Pitch Formation, and clasts and fragments of the Grand Pitch Formation are present in the lower part of the Shin Brook Formation (Neuman and Rankin, 1980). This unconformity records the type Penobscottian disturbance of Neuman (1967).

The Chase Lake Formation (Fig. 2, col. 21) in the Munsungun-Winterville anticlinorium overlies mélange and black slate as does the Shin Brook Formation. The basal part of the Chase Lake Formation is graywacke and conglomerate with clasts of dark slate, basaltic and felsic volcanic rocks, chert, and dolerite. The upper part of the formation is interbedded medium-dark gray slate and graywacke on a scale of 1 to 25 cm (Hall, 1970). This upper part of the section is thought to interfinger with the Bluffer Pond Formation, consisting dominantly of pillowed basalt (Hall, 1970). Abundant graptolites, a few gastropods, and a brachiopod indicate that the Chase Lake and Bluffer Pond Formations have an early Caradocian age (Hall, 1970). The Chase Lake Formation lies unconformably on the black slate and mélange of the Chase Brook Formation, so that the sequence and lithologies are nearly identical to those exposed on Shin Brook in the Weeksboro-Lunksoos anticlinorium. But the Chase Lake Formation is Caradocian, whereas the Shin Brook Formation is Arenigian. Therefore, we interpret these sequences as distinctly different in age, representing onlap sequences on an ancient uplifted area. Presumably the time spanned by the Chase Lake unconformity decreases to the west and the volcanic sections merge.

Unnamed andesites and basalts (Fig. 2, col. 20) occur in fault-bounded blocks along the east side of the Weeksboro-Lunksoos anticlinorium (Neuman and Rankin, 1980). These volcanic rocks are largely massive, dark greenish gray, locally pillowed lava and flow breccia. Many are highly porphyritic, containing both plagioclase and pyroxene phenocrysts in a fluidal groundmass.

Also exposed in a fault-bounded block on the east side of the Weeksboro-Lunksoos anticlinorium is the Wassataquoik Chert (Fig. 2, col. 20). It consists of thinly bedded, medium to dark gray, greenish gray, and red chert, with interbedded felsic and mafic tuffs and possible dark grayish green flows (Neuman, 1967). Based on the presence of interbedded volcanic rocks, the Wassataquoik Chert is interpreted to conformably overlie and to partly interfinger with the nearby unnamed volcanic section just described. Interbedded siliceous shales contain graptolites of the *Climacograptus bicornus* and *Orthograptus truncatus* var. *intermedius* zones, as well as conodonts and brachiopods (Neuman, 1967). These fossils indicate the Wassataquoik Chert is Caradocian.

Mafic flows and tuffs, felsic flows and tuffs, volcanic breccia, and minor chert and slate have been assigned to the Munsungun Lake Formation (Fig. 2, col. 21) in the Munsungun-Winterville anticlinorium (Hall, 1970). This section is similar to the combined volcanic-Wassataquoik Chert section described above for the Weeksboro-Lunksoos anticlinorium, and it is of the same age.

In the Munsungun-Winterville anticlinorium, the Blind Brook Formation (Fig. 2, col. 21) overlies the Munsungun Lake Formation (Hall, 1970). The Blind Brook Formation consists of pyritiferous black slate overlain by lighter gray silty slate. It has been assigned to the *Orthograptus truncatus* var. *intermedius* zone and, therefore, may correlate with black slates in the Wassataquoik Chert.

The small inlier at Castle Hill, Maine (Fig. 3) exposes mafic and felsic volcanic rocks and minor black sulfidic slate of the Winterville Formation (Fig. 2, col. 22). Fossils found at Castle Hill suggest a Caradocian age (Boucot et al., 1964; Roy and Mencher, 1976), so that they correlate with the volcanic rocks in the Munsungun-Winterville anticlinorium. At the Becaguimec inlier, however, St. Peter (1982) has described the Craig Brook Limestone containing Llanvirnian and Llandeilian conodonts (Fig. 2, col. 24). Fyffe et al. (1983) note that these limestones have equivalent age and lithology to limestone beds in the Belle Lake Formation near Waterville in the Miramichi belt (Fig. 3).

Thick Ashgillian to Ludlovian Sequences. Thick Ashgillian to Ludlovian cover sequences (Fig. 2, col. 25) lie west of the Miramichi and Casco Bay - south-central Maine belts, east of the Weeksboro-Lunksoos and the Munsungun-Winterville anticlinoria, and surround the Becaguimec and Castle Hill inliers (Fig. 3). North of Houlton, Maine, (Fig. 3) the cover sequence contains abundant carbonates. These carbonates decrease in importance southward, so that south of the vicinity of Houlton, the section is mainly a thick series of turbiditic graywackes, slates, and minor ribbon limestones and conglomerate.

The composition of the carbonate-rich sequence to the north varies asymmetrically across strike, with detrital rocks prominent along the eastern margin of the Munsungun-Winterville anticlinorium (Fig. 3) and sections with more abundant carbonates to the east (Roy and Mencher, 1976). Adjacent to the east flank of the Munsungun-Winterville anticlinorium, the Aroostook River Formation (Fig. 2, col. 22), consisting of thinly bedded greenish gray slate, calcareous wacke, and siltstone, rests conformably or with disconformity on the older rocks of the anticlinorium (Roy and Mencher, 1976). These slates and sandstones grade laterally toward the east through the Pyle Mountain Argillite (Boucot et al., 1964) into at least part of the Carys Mills Formation (Pavlides, 1968). The Carys Mills Formation (Fig. 2, col. 25) consists of 7 to 15 cm thick, interbedded, buff weathering, light gray limestone and gray calcareous shale. The limestones are locally cross-bedded, and at many places there is gradation between the limestone and the shale. This formation contains both Ashgillian and Llandoverian fossils, so the systemic boundary lies within it (Rickards and Riva, 1981).

The Frenchville Formation (Fig. 2, col. 22) oversteps the Aroostook River Formation, and to the west it rests directly on older rocks (Roy and Mencher, 1976). It consists of polymictic conglomerate with clasts of mafic and felsic volcanic rocks, chert, graywacke, red slate, and manganiferous ironstone, some of which were clearly derived from rocks of the Munsungun-Winterville anticlinorium (Boucot et al., 1964; Roy and Mencher, 1976). The Frenchville Formation becomes more distal toward the east and grades laterally into the New Sweden (Roy and Mencher, 1976) and Spragueville Formations (Pavlides, 1965) (Fig. 2, col. 23). The New Sweden Formation is principally laminated calcareous slate containing discontinuous beds of dense limestone. The Spragueville Formation contains thinly interbedded gray shale, light gray silty limestone, and calcareous siltstone. Scattered fossils from these rocks indicate a Llandoverian to possibly Wenlockian age (Boucot et al., 1964; Roy and Mencher, 1976).

The Jemtland Formation (Fig. 2, cols. 22, 23) overlies the Frenchville Formation and its equivalents, and it may be equivalent in part to the top of the Spragueville Formation (Roy and Mencher, 1976). The Jemtland Formation consists of calcareous slate, silty shale, fine grained graywacke, and minor aquagene tuff and micritic limestone. Scattered graptolites suggest a late Wenlockian to Ludlovian age (Roy and Mencher, 1976).

The Fogelin Hill Formation (Fig. 2, col. 22) conformably overlies the Jemtland Formation (Roy and Mencher, 1976). It consists of red and green silty slate and interbedded calcareous graywacke and siltstone. Rather wide-ranging fossils indicate a Ludlovian to Siegenian age (Roy and Mencher, 1976).

To the east in New Brunswick (Fig. 3), the Carys Mills Formation (Fig. 2, col. 24) is conformably overlain by rocks equated with the Smyrna Mills Formation by Rast et al. (1980). At the type section of the Smyrna Mills Formation, located near Houlton, Maine (Pavlides, 1971), it consists of slightly calcareous, gray sandstone in beds 10 cm to 4 m thick, prominent sequences of thinly bedded greenish phyllite and sandstone, minor black slate, red and maroon siltstone, local manganiferous slate and ironstone, and local conglomerate. Graptolites from these rocks indicate a Llandoverian to Wenlockian age. North of Woodstock (Fig. 3) amygdaloidal basaltic rocks are interbedded with the Smyrna Mills Formation.

South of Houlton, Maine (Fig. 3) where carbonates are a minor part of the section, Ekren and Frischknecht (1967) described feldspathic graywacke, gray and green siltstone and slate to make up a unit that they called the Mattawamkeag Formation. The sandstones, siltstones, and slates of this unit occur in 2 to 60 cm beds that are commonly graded. A coarse conglomerate containing clasts of felsite, greenstone, quartzite, chert, and quartz porphyry occurs at the top of this unit. No fossils have been found in the Mattawamkeag Formation, but fragmental brachiopods in the conglomerate suggest a Silurian age (Ekren and Frischknecht, 1967). We interpret the Mattawamkeag unit to be equivalent to part of the Smyrna Mills Formation (Fig. 2, col. 25) and the conglomerate to be a tongue that is proximal toward the west, possibly equivalent to the Frenchville Formation as suggested by Neuman (1967).

The Allsbury Formation (Ekren and Frischknecht, 1967; Neuman, 1967) probably partly overlies and partly has a facies relationship with the Smyrna Mills Formation (Fig. 2, col. 25). It consists of thinly bedded greenish gray to maroon phyllite and light gray quartzite and sandstone. Locally thick beds of graywacke are interbedded with the dominant phyllitic rocks. Thinly bedded gray limestone forms a minor unit within the formation. Graptolites from scattered localities suggest a late Llandoverian through Wenlockian age (Neuman, 1967).

A thin horizon of rusty, black, sulfidic slate (Fig. 2, col. 25) which overlies the Allsbury Formation has yielded late Llandoverian graptolites (Roy and Forbes, 1970). It is followed by a thick section of massive, thickly bedded, slightly calcareous quartz wacke, graywacke, and gray slate of the Lawler Ridge Formation (Roy, 1981; Roy et al., 1983).

Thin Sequences of Ashgillian to Ludlovian Rocks. Thin sequences of Ashgillian to Ludlovian rocks are exposed in the Munsungun-Winterville, the Weeksboro-Lunksoos, and the Lobster Mountain anticlinoria and near Caucomgomoc Lake (Fig. 3). A discontinuous Ashgillian section (Fig. 2, col. 20) is exposed in the Weeksboro-Lunksoos anticlinorium, where it includes polymictic conglomerate with clasts of quartz, quartzite, volcanic rocks, and slate, interbedded with lithic wacke, green siltstone, and minor pillowed basalts (Neuman, 1967; Neuman and Rankin, 1980). Ashgillian brachiopods, trilobites, and corals have been identified from these rocks (Neuman and Rankin, 1980).

A thin sequence of Silurian strata (Fig. 2, col. 20) is also exposed on the west side of the Weeksboro-Lunksoos anticlinorium (Neuman, 1967), where it unconformably overlies Ashgillian and older rocks. The basal unit consists of polymictic conglomerate interbedded with dark gray sandstone and siltstone containing an upper Llandoverian assemblage of brachiopods and corals. An unnamed unit of light gray calcareous siltstone and fine grained sandstone overlies the conglomerate (Neuman, 1967). These rocks are laminated and commonly well bedded, with limestone nodules and thin discontinuous beds of limestone locally present. Scattered fossils within this unit indicate that it is late Llandoverian to Wenlockian in age. Limestones, composed largely of reefal material and debris with minor calcarenite, overlie the calcareous siltstone and sandstone unit, and in turn they are overlain by an upper unit of calcareous siltstone. A few scattered brachiopods suggest a Wenlockian to Ludlovian age (Neuman, 1967).

Pridolian to Gedinnian Volcanic Sections. Pridolian to Gedinnian volcanic rocks are exposed in the Weeksboro-Lunksoos and the Musungun-Winterville anticlinoria, and near Presque Isle (Fig. 3). In the west limb of the Weeksboro-Lunksoos anticlinorium, these volcanic rocks (Fig. 2, col. 20) consist of intermediate to mafic flows, tuff, and volcanic breccia (Neuman, 1967). These unfossiliferous volcanic rocks are between Ludlovian siltstones and the Siegenian Seboomook Formation, suggesting that their age is latest Ludlovian to Gedinnian.

A somewhat similar volcanic section (Fig. 2, col. 21), as much as 3250 m thick, has been referred to as the Spider Lake Formation (Hall, 1970) in the southern part of the Munsungun-Winterville anticlinorium (Fig. 3). The Spider Lake Formation has a basal polymictic conglomerate, overlain by andesitic flows, lesser dacite, and gray calcareous siltstone and medium to dark gray limestone. The volcanic rocks are interpreted to interfinger with a dominantly sedimentary section composed of calcareous siltstone and shaly limestone. Abundant fossil finds within these sections indicate a Pridolian to Gedinnian age (Hall, 1970).

In the vicinity of Presque Isle (Fig. 3), the Jemtland Formation is overlain unconformably by the Dockendorf Group (Boucot et al., 1964). This group (Fig. 2, col. 23) consists of andesite, trachite, rhyolite, and tuff, the upper part of which interfingers with the Chapman Sandstone (Williams, 1899), consisting of greenish brown argillaceous sandstone with minor beds and lenses of brown mudstone. The Chapman Sandstone, in turn, interfingers with dark green, slightly calcareous shale with sparse thin beds of quartzite that has been assigned to the Swanback Formation (Boucot et al., 1964). Many of these rocks are richly fossiliferous and the fossils indicate a Gedinnian age (Boucot et al., 1964).

Similar rocks (Fig. 2, col. 24) have been recognized in New Brunswick, to the northeast, south, and west of Plaster Rock (Fig. 3). There, a Silurian(?) basaltic and felsitic section (Fyffe, 1982) underlies the Tobique Group (St. Peter, 1978). A lower unit of the Tobique Group consists of dark gray to dark green siltstone and sandstone in beds from a few centimeters to several meters thick. Minor beds of graywacke, conglomerate, and andesite are intercalated with the siltstone-sandstone sequence. This dominantly sedimentary sequence is followed by a dominantly volcanic section (St. Peter, 1979). The volcanic rocks include amygdaloidal andesite, andesitic tuff, hyaloclastic tuff, and minor volcanic conglomerate and siltstones. Toward the top of the group, aphanitic rhyolite, rhyolitic ash-flow tuff, and minor quartz porphyry are present. Brachiopods from the Tobique Group indicate a Gedinnian age.

Siegenian to Emsian Flysch. The Carrabassett Formation (Boone, 1973) is reported to lie conformably on equivalents of the Lawler Ridge Formation (Osberg et al., 1985) south of Millinocket (Fig. 3), although the presence of sparse lenses of polymictic conglomerate near its base suggest the possibility that locally it is disconformable. The Carrabassett Formation (Fig. 2, col. 25) consists of massive gray pelite with lesser beds of siltstone and sandstone and rhythmically interbedded siltstone and gray pelite in distinctive graded sets. Poorly preserved brachiopods suggest that the Carrabassett Formation is Early Devonian (Espenshade and Boudette, 1967).

Rocks of Siegenian to Emsian age, mostly belonging to the Seboomook Formation (Fig. 2, cols. 20 and 21), form an extensive blanket that covers the older sequences of northern Maine and westernmost New Brunswick. These rocks are typically cyclically layered dark sandstone and slate, but the proportions of sandstone and slate vary considerably over the outcrop (Boucot et al., 1964). In western Maine these flyschoid rocks may be as old as Gedinnian and are reported to have conformable relationships with the underlying Madrid Formation, but in northern Maine they are locally unconformable on older sections, resting on rocks as old as Cambrian (Roy, 1980). The youngest strata of the Seboomook Formation are thought to be Emsian because they are above Siegenian fossils, and because Emsian fossils are known from the correlative Temiscouata Formation near the Maine-New Brunswick-Quebec border (St. Peter and Boucot, 1981). The Seboomook Formation in north-central Maine is interpreted as representing an eastward-derived, westward-prograding prodelta sequence related to the onset of the Acadian orogeny (Hall et al., 1976; Roy, 1980). Roy and Mencher (1976) describe somewhat similar rocks (Fig. 2, col. 22) south of the Ashland basin (Fig. 3).

Post-Early Devonian Rocks. Middle Devonian red beds overlie the Dockendorf Group near Presque Isle (Fig. 3), and Mississippian red beds and basalts overlie the units of the Tobique Group in the vicinity of Plaster Rock (Fig. 3). The post-Early Devonian rocks are less deformed than the underlying rocks, and the unconformity at their base is interpreted to mark the end of the Acadian orogeny (Boucot et al., 1964).

CORRELATIONS

General Statement

We now consider correlations between lithotectonic belts with reference to Figure 4. The diagrams are coded as to age into eight groups. In addition, formations for which age constraints are good are distinguished from those for which constraints are poor or lacking. In formations with poor age control,

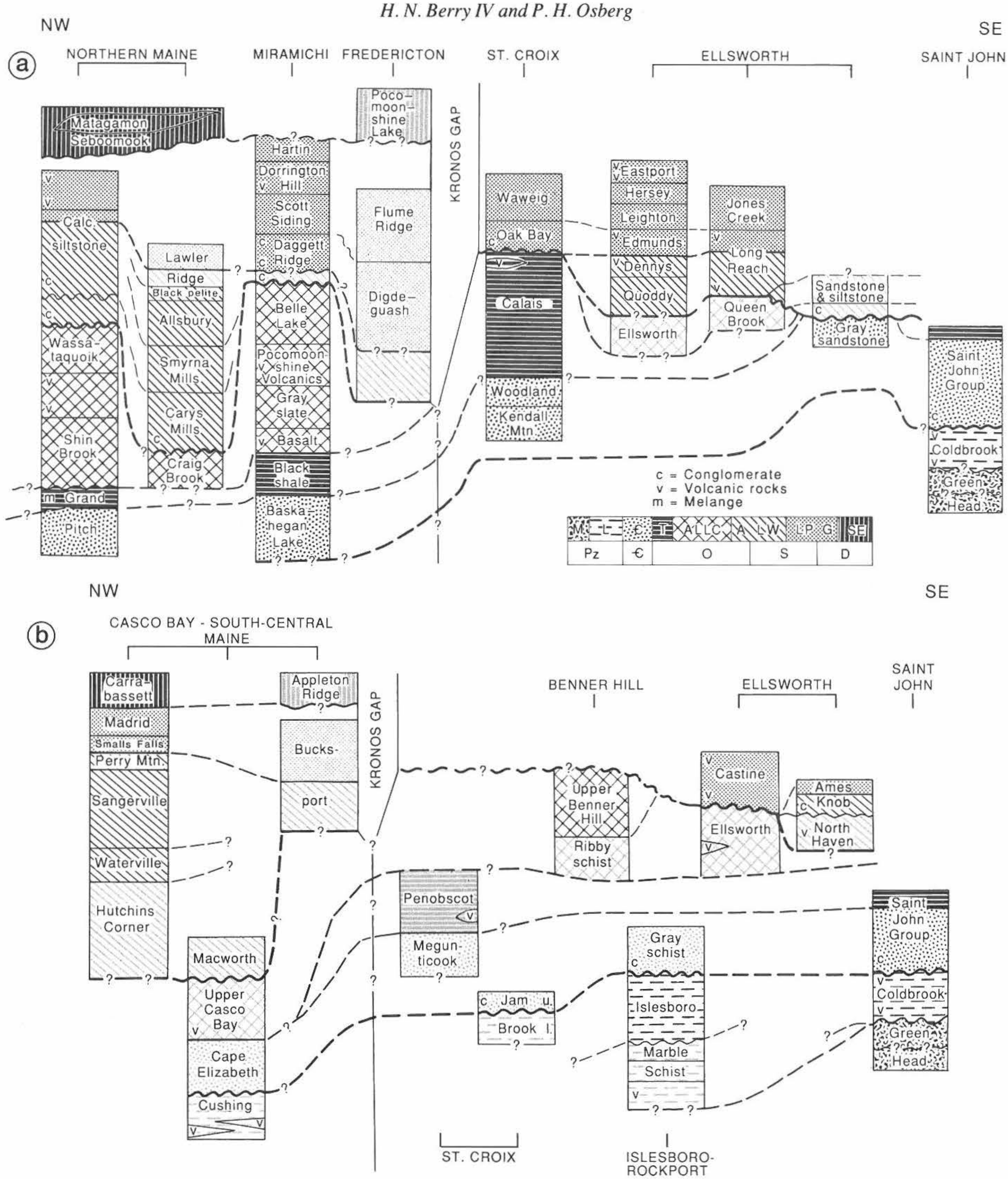

Figure 4. Schematic correlation among lithotectonic belts. Time-lines divide units into age groups as indicated by patterns; units with light patterns have poor age control. Columns are composite sections based on Figure 2. Silurian-Devonian correlation across the Kronos gap is precluded, but pre-Ashgillian correlations are considered. Shape of Ashgillian unconformity (emphasized) reflects distribution of thin and thick Silurian sequences (not to scale). (a) Northern Maine southeast to Saint John, New Brunswick. (b) South-central Maine east to Penobscot Bay, with Saint John section for comparison. See Editors' Note regarding the stratigraphic order in the St. Croix belt (4a).

assigned ages are based on comparisons with fossiliferous sequences.

Figure 4 displays two stratigraphic diagrams, one from northern Maine southeast to Saint John (Fig. 4a), and the other from south-central Maine eastward to Penobscot Bay, with the Saint John section added for comparison (Fig. 4b). The Saint John belt was included in the Avalon platform of Rast et al. (1976) and the Avalon zone of Williams (1979) because its sequence of Late Precambrian through Early Ordovician rocks and Cambrian Old World fauna closely matches that of the Avalon Peninsula, Newfoundland. The Saint John belt, due to its Avalon connection, provides an important reference section to which other belts in the eastern Maine-western New Brunswick region may be compared (Fig. 2, col. 1). We will discuss the correlations system by system.

Precambrian

Rocks with established Precambrian ages are exposed in the Saint John and the Islesboro-Rockport belts. In addition, rocks that, because of their stratigraphic position or lithic similarity to established Precambrian rocks are interpreted to be Precambrian, are found in the Ellsworth, the St. Croix, and the Casco Bay - south-central Maine belts. The recognition of volcanic rocks near Sagwa and on Campobello Island lithologically similar to those of the Coldbrook Group suggests that the Ellsworth and Saint John belts were associated with one another in Late Precambrian time.

The Precambrian sequence on Seven Hundred Acre Island in the Islesboro-Rockport belt has been equated to the Green Head Group on the basis of general lithologic similarity (Stewart, 1974), but the abundance of volcanic rocks in the Seven Hundred Acre Island section makes it quite different. Moreover, its radiometric age is more like that of the Coldbrook Group than the Green Head Group. The overlying Islesboro sequence (Fig. 4b), interpreted here to be Precambrian, contains sedimentary facies that are similar to those in the Green Head Group, but its position above the Seven Hundred Acre Island section suggests that such a correlation is not valid. We believe these observations argue against a close connection between the Saint John and the Islesboro-Rockport belts in the late Precambrian, although they could have been distant parts of the same terrane.

In the St. Croix belt, the lower part of the Jam Brook Formation (Fig. 4b), consisting of marble, quartzitic schist, and quartzite, is similar in sequence and lithology to part of the Islesboro sequence in the Islesboro-Rockport belt. This similarity suggests that the two belts were associated in the Late Precambrian.

Rocks of the Cushing Formation in the Casco Bay - south-central Maine belt (Fig. 4b) contain too much marble to be a good match with the Coldbrook Group and too much amphibolite to be equivalent to the Green Head Group in the Saint John belt. However, the Cushing Formation contains the same spectrum of lithologies as does the Seven Hundred Acre Island sequence in the Islesboro-Rockport belt, and on this basis, they might be correlative.

Cambrian

Fossiliferous Cambrian rocks occur in the Saint John, Ellsworth, and northern Maine belts (Fig. 4a). Rocks assigned to the Cambrian because of stratigraphic position and/or lithic similarity to known Cambrian rocks, are exposed in the Islesboro-Rockport, St. Croix, Casco Bay - south-central Maine, and Miramichi belts. Sparsely exposed Cambrian sandstones in the Ellsworth belt in southwestern New Brunswick have lithic similarity and contain identical fossils to the Ratcliffe Brook Formation in the Saint John belt, indicating contiguity of the two belts in the Cambrian. Farther afield, in the Islesboro-Rockport belt (Fig. 4b), the upper sequence of conglomerate, sandstone, and gray schist exposed on Islesboro and near Rockport contains lithologies that are like those in the Ratcliffe Brook section. In the neighboring St. Croix belt, the upper part of the Jam Brook Formation is essentially identical to the upper sequence at Islesboro, and if the Megunticook Formation truly overlies the Jam Brook Formation, the Jam Brook-Megunticook sequence would be quite similar to the Ratcliffe Brook Formation through Silver Falls section in the Saint John belt. The Kendall Mountain and Woodland Formations might represent correlative units in the eastern part of the St. Croix belt (Fig. 4a).

Rocks assigned to the Cambrian in the Casco Bay - south-central Maine, Miramichi, and northern Maine belts are sufficiently removed from the Saint John belt that correlations are more tenuous. However, the fossiliferous Cambrian rocks in the northern Maine belt consist of sandstones and shales that are much like the lower part of the Saint John section, and, therefore, it is conceivable that they might be equated. Rocks regarded as Cambrian in the Miramichi belt (Fig. 4a) are lithologically identical to those in the northern Maine belt. Rocks tentatively assigned to the Cambrian in the Casco Bay - south-central Maine belt, consisting of interbedded sandstones and shales, might also be equated to the northern Maine Cambrian rocks, but with less assurance.

Ordovician

Fossiliferous Ordovician rocks are known in the Saint John, Benner Hill, St. Croix, Miramichi, and northern Maine belts (Fig. 4a). In addition, rocks interpreted to be Ordovician occur in the Ellsworth and Casco Bay - south-central Maine belts (Fig. 4b). Black, sulfidic shales, principally carrying Tremadocian graptolites, but some with earliest Arenigian fossils as well, are prominent in the Saint John, St. Croix, and Miramichi belts (Fig. 4a). The black sulfidic shale and mélange deposits in the northern Maine belt are, within uncertainty, Early Ordovician as well, and they may be contiguous with the black shales to the southeast. Black shales are absent in some sections in the

Miramichi belt, and we interpret their absence to be due to local erosion. The apparent absence of black shales at this stratigraphic position in other belts may be due to erosion also. Alternatively, black shales in the Prison Farm unit of the Benner Hill belt, the Diamond Island, upper Wilson Cove, and other units of the Casco Bay - south-central Maine belt are all of uncertain age, and any of them could be unrecognized Lower Ordovician strata. The occurrence of lithologically similar Tremadocian shales in so many of the lithotectonic belts is the strongest argument that all of the belts in eastern Maine and western New Brunswick were near each other during the earliest Ordovician.

The Saint John belt contains no rocks younger than the Arenigian, so we make use of sections in the northern Maine belt as reference for rocks younger than Tremadocian. The Shin Brook Formation, its overlying volcanic sections, the Wassataquoik Chert, and the Chase Lake, Bluffer Pond, and Munsungun Lake Formations in the northern Maine belt (Fig. 4a) are nearly identical to the volcanic section near Danforth in the Miramichi belt, and are sufficiently similar so as to invite correlation with the Pocomoonshine Volcanics to the north near Benton and Hayesville. The Spring Point greenstone (upper Casco Bay Group) in the Casco Bay - south-central Maine belt, although not dated, occupies a somewhat similar position and so it may be equivalent. These stratigraphic similarities suggest that the northern Maine, Miramichi, and Casco Bay - south-central Maine belts were juxtaposed during the Middle and Late Ordovician.

Extending these correlations to more easterly belts is far less certain. Correlations to the east depend, in part, on the nature of the contact between the Ellsworth and St. Croix belts. If this contact is sedimentary as McGregor (1965) and McCutcheon and Ruitenberg (1987) suggest, there is little reason to separate the St. Croix and the Ellsworth into distinct lithotectonic belts. The distinction is significant only if a fault everywhere separates the two, as suggested by Stewart and Wones (1974). In particular, if the contact between the Ellsworth-Queen Brook Formations and the Tremadocian Calais Formation is stratigraphic, the Ellsworth Formation would have a stratigraphic position similar to that of the Pocomoonshine Volcanics in the Miramichi belt. In fact, the Ellsworth Formation does contain relatively thin felsic and mafic volcanic rocks, but the Queen Brook Formation has none. If the Ellsworth-Queen Brook rocks are the same age as the Pocomoonshine Volcanics and equivalents to the northwest, they must be related through a facies change in which the volcanic rocks thin toward the southeast into a dominantly shaly sequence. The "ribby" schist in the Benner Hill belt is lithologically similar to the Ellsworth Formation except for its lack of volcanic rocks. On the other hand, if a fault everywhere separates the Ellsworth and St. Croix belts, stratigraphic correlations with more western belts are uncertain at best.

The Blind Brook Formation in the northern Maine belt has the same age and contains some of the same lithologies as does the Belle Lake Formation in the Miramichi belt (Fig. 4a), suggesting stratigraphic contiguity between the two belts. This possibility is strengthened by the occurrence at Becaguimec in the northern Maine belt (Fig. 3) of limestones that are similar in age and lithology to limestone interbedded with the Belle Lake Formation near Waterville, New Brunswick. To the south in the Casco Bay - south-central Maine belt (Fig. 4b), the Scarboro, Spurwink, and Jewell Formations (upper Casco Bay Group) lie above a volcanic unit as does the Belle Lake Formation in the Miramichi belt, and they have a similar range of lithologies, consistent with the possibility that the northern Maine, Miramichi, and Casco Bay - south-central Maine belts were adjacent to one another in the Caradocian. Correlation with more easterly belts is uncertain. The Benner Hill sequence in the Benner Hill belt (Fig. 4b) contains units that are at least in part the same age as the Belle Lake Formation. However, the lithologies and sequence do not make an especially good match with the section included in the Belle Lake Formation, and if they are to be regarded as correlatives, a considerable facies change connects them.

Ashgillian rocks are reported in the northern Maine and Miramichi belts. These rocks are sufficiently similar to suggest that they were deposited in close proximity to one another. In addition, the Carys Mills Formation in the northern Maine belt, which contains the Ordovician-Silurian systemic boundary, contains a basal conglomerate at Becaguimec (Fig. 3). A lithic counterpart to basal Carys Mills Formation is exposed in the northeastern part of the Casco Bay - south-central Maine belt, where Ludman (pers. commun., 1979) has mapped 5 to 25 cm bedded limestones intercalated with slates in the cover sequence south of Danforth (Fig. 3). These basal beds of the Carys Mills Formation are also like an unnamed sequence of conglomerates, grits and slate near Danforth in the Miramichi belt (Ludman, pers. commun., 1988). Farther south in the Casco Bay - south-central Maine belt, the Carys Mills Formation must thin to a feather edge, but even near Palermo (Fig. 3) it could be represented by the limestone member of the Hutchins Corner Formation.

Silurian and Early Devonian

Above the Carys Mills Formation, the sequence consisting of Smyrna Mills, Allsbury, unnamed sulfidic slate, and Lawler Ridge Formations of the northern Maine belt (Fig. 4a) is nearly identical in lithology and age to the section consisting of the upper part of the Hutchins Corner, Waterville, unnamed sulfidic phyllite, and Sangerville Formations in the Casco Bay - south-central Maine belt, and as noted earlier, the Silurian sequence of the Casco Bay - south-central Maine belt (Fig. 4b) would extend into the turbidites of the Fredericton belt if the intervening plutons were not present. Although some of the rocks in the Miramichi belt have the same range in age, comparable sequences are not found there. Even so, conglomerates containing clasts derived from Miramichi rocks within the Silurian sequence of the Fredericton belt serve to link the Miramichi and Fredericton belts in the Silurian (Fyffe and Fricker, 1987).

Pridolian and Gedinnian volcanic rocks with minor sedimentary rocks are exposed in a broad zone across the northern Maine and Miramichi belts. The Spider Lake Formation and equivalent rocks in the northern Maine belt are similar in lithology and age to the Dorrington Hill and Hartin Formations of the Miramichi belt (Fig. 4a). Comparable volcanic sections are not found farther to the south in the northern Maine belt, or in the Casco Bay - south-central Maine or the Fredericton belts.

These volcanic and associated sedimentary rocks are lithologically and temporally similar to rocks in the Ellsworth belt, but the rocks in the Ellsworth belt cannot have been contiguous because they contain fauna that have Old World affinities, whereas the northern Maine and Miramichi volcanic rocks contain fauna that are North American. This disctinction between faunal provinces is indicated in Figure 4 by the Kronos gap. The presence of sandstone clasts of the Coldbrook Group in basal Silurian conglomerates in the Ellsworth belt points to the closeness of the Ellsworth and Saint John belts in the Silurian.

The Carrabassett Formation and at least part of the Seboomook Formation in the northern Maine belt are lithologically similar to the Pocomoonshine Lake Formation in the Fredericton belt and to the Appleton Ridge Formation in the eastern part of the Casco Bay - south-central Maine belt. These formations are, therefore, tentatively correlated (Fig. 4). If these correlations are valid, Early Devonian flysch must have spread across large parts of the Fredericton, Casco Bay - south-central Maine, and Miramichi belts.

TECTONIC EVOLUTION

Neuman (1984) has claimed that Arenigian rocks from the northern Maine and Miramichi belts include Celtic fauna, which he argues are distinct from those that inhabited North America. He interprets the Celtic assemblage as an island community, isolated by its environment, but with sporadic incursions of both North American and Old World forms. We interpret the area over which the Celtic assemblage occurred to include at least the northern Maine, the Miramichi, and possibly the Casco Bay - south-central Maine belts, and we name this paleogeographic region the Celtic archipelago.

If this interpretation is correct, the Celtic fauna in the northern Maine and Miramichi belts indicate that a faunal barrier (Iapetus Ocean) separated these belts from North America during at least earliest Ordovician time. The longevity of the Celtic archipelago has been discussed by Neuman (1984). The Celtic biogeographic province was established in Arenigian time and lasted into the Llanvirnian. Neuman (1984) states that during the Llandeilian and Caradocian, the Celtic habitat ceased to exist and the fauna of northern Maine and Miramichi adopted North American characteristics, suggesting that these lithotectonic belts were in close proximity to North America at that time. The Caradocian volcanic rocks in the northern Maine belt, which may have an arc-type chemical signature (Hynes, 1976), probably were erupted during the closure of Iapetus.

The eastern limit of the Celtic archipelago is questionable. The stratigraphic linkages of the northern Maine, Miramichi, and Casco Bay - south-central Maine belts to the more easterly belts, although permissive of continuity, do not require it, and the identification of faunal realms that might serve to delineate the Celtic archipelago is beset by controversy. Boucot et al. (1972) and Boucot (1973) argue that Caradocian fossils in the Benner Hill belt are Old World, and if this is the case, a faunal boundary must have existed in Caradocian time and possibly earlier between at least the Benner Hill and the Miramichi and northern Maine belts, which both contain North American Caradocian fauna. Thus, if Boucot's interpretation of the Benner Hill fossils is correct, the Celtic province encompassed only the northern Maine, the Miramichi, and possibly the Casco Bay - south-central Maine belts. The more easterly belts (Avalon?) were separated by a pre-Silurian ocean from the Celtic province and belonged to an Old World province. On the other hand, if Boucot's interpretation is incorrect as suggested by Neuman (1973), there is little reason to postulate a pre-Silurian ocean within the lithotectonic belts defined in this paper, and the Celtic province would extend eastward into the Saint John belt. The Celtic archipelago and Avalon would be one and the same.

If an eastern pre-Silurian ocean existed at all in Cambrian through Ordovician time, it must have been located between the Celtic archipelago and the Ellsworth - Islesboro-Rockport - Saint John belts, since these eastern belts have stratigraphic linkages to one another. But the position of the St. Croix belt relative to this ocean is uncertain. If the contact between the Ellsworth and the St. Croix belts is sedimentary, the St. Croix belt would have been linked to the Ellsworth belt, and the pre-Silurian ocean would lie west of the St. Croix. But if a major fault exists along the St. Croix-Ellsworth contact, there are no constraints as to the position of the St. Croix belt; it could lie either east or west of the pre-Silurian ocean.

Although the existence of an eastern pre-Silurian ocean is uncertain, it is clear that an ocean did exist to the west of the St. Croix belt from Silurian into Early Devonian time. This is indicated by Silurian - Devonian stratigraphic linkages between the Casco Bay - south-central Maine and Fredericton belts and between the St. Croix, Ellsworth, and Saint John belts. These linkages and the occurrence of North American fauna in the northern Maine and Miramichi belts in contrast to Old World fauna in the Ellsworth belt indicate that a barrier to the migration of fauna existed across what is now the western boundary of the St. Croix belt. Because the Silurian to Early Devonian sedimentary rocks that blanket the lithotectonic belts are marine, we conclude that the barrier was an ocean, which we call the Kronos Ocean. In the Late Silurian and Early Devonian, the Kronos Ocean was of sufficient width to prevent the migration of faunal elements. This ocean could have existed earlier, during the entire Silurian, without having its consequence to faunal migra-

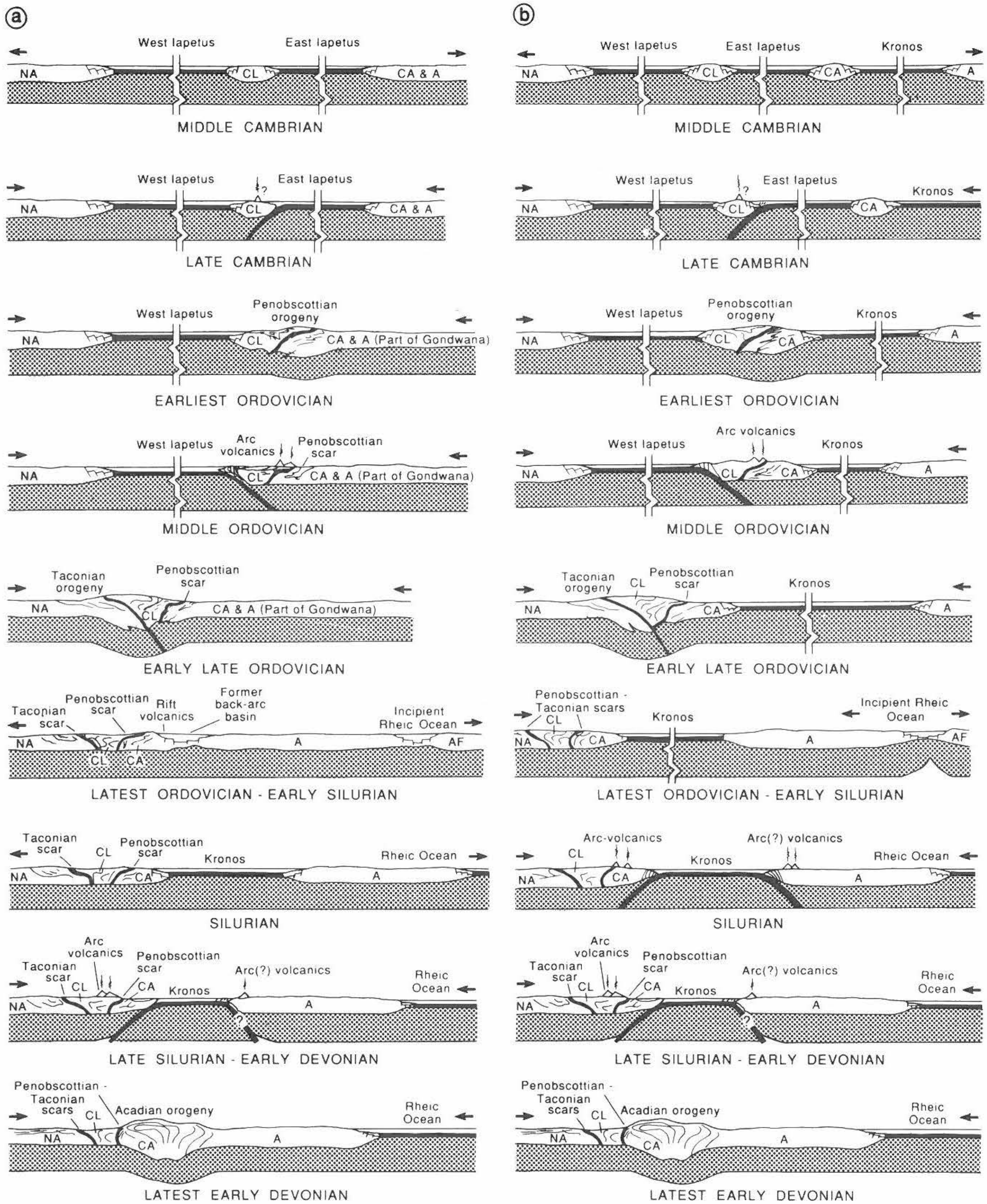

Figure 5. Graphic representation of two tectonic models which explain the lithotectonic belts, their fauna, and correlations from Middle Cambrian time through the Early Devonian. See text for discussion. NA, North America; CA, Celtic archipelago; A, Avalon; CL, Chain Lakes terrane. Arrows indicate relative plate motion.

tion register because of the cosmopolitan character of fossils during most of the Silurian (Cocks and McKerrow, 1977).

Figure 5 shows schematically our view of the tectonic events surrounding the closings of Iapetus and the Kronos as they affect the lithotectonic belts under discussion. All of the belts in eastern Maine and western New Brunswick are interpreted to have continental basements, based on the abundance of felsic volcanic rocks in their stratigraphic records and the presence of S-type plutons that intrude them.

Two schemes are shown because of the uncertainty about the existence of the Kronos in the Cambrian and Ordovician. In the first scheme (Fig. 5a) the Kronos Ocean is assumed to have existed only from Early Silurian through Early Devonian time. In this model the Celtic archipelago included all of the lithotectonic belts described in this paper, and its eastern limit is beyond the area of this study. Regionally, we interpret the archipelago to have been at the western edge of Gondwana, as suggested by Neuman (1984). It was bordered to the west by the eastern Iapetus Ocean that separated it from the Chain Lakes terrane. The Chain Lakes terrane was in turn separated from North America by the western Iapetus Ocean.

The eastern Iapetus Ocean closed from Late Cambrian into earliest Arenigian time along a west-dipping subduction zone located on the east side of the Chain Lakes terrane. The polarity of this subduction zone is suggested by the contiguity of Tremadocian black shales in the Miramichi belt with the black shale-mélange deposits of the northern Maine belt. If we consider the Tremadocian black shales to be trench deposits, from east to west we encounter trench deposits, mélange containing ophiolitic fragments, and in the Lobster Lake anticlinorium (Fig. 3), accretionary wedge deposits (Boudette, 1982; Boone, this volume), implying a west-dipping subduction zone. The lack of arc-type volcanic rocks of Late Cambrian to earliest Ordovician age on the Chain Lakes terrane may suggest that collision was oblique, or their absence may be due to their partial erosion, or they may be hidden under the thick and extensive cover of Devonian rocks. In any event, the final closing of the eastern Iapetus resulted in the Penobscottian orogeny, which is not very strongly expressed in the area of this study. This closure brought together the Chain Lakes terrane and the Celtic archipelago.

From Llanvirnian time into the Caradocian, the western Iapetus closed along an east-dipping subduction zone developed on the west side of the Chain Lakes terrane. A consequence of this subduction was the construction of an arc-type volcanic pile on the remnants of the Chain Lakes terrane and the western part of the northern Maine lithotectonic belt. This volcanic edifice rests in part unconformably across the Tremadocian black shale and mélange deposits of trench origin in the northern Maine belt. To the east in the Miramichi belt, the compositions of the volcanic rocks give way to rocks having within-plate characteristics, suggesting a back-arc basin (van Staal, 1987). The western Iapetus Ocean closed by subduction beneath the combined Chain Lakes terrane and Celtic archipelago terrane, and this closure finally led to the collision of this terrane with North America in the Taconian orogeny, the effects of which were either relatively mild in northern New England and western New Brunswick or more probably are partly confused with younger tectonic features.

In this model (Fig. 5a), the rifting that opened the back-arc basin to the east of the Caradocian volcanic arc continued into the Silurian. The thin and thick facies of the Silurian sections represent deposits in a rifting environment, and the bimodal volcanic rocks with possible within-plate characteristics in the Ellsworth belt represent rift-type volcanic rocks erupted onto extending crust during the opening of the Kronos. Greenstones of the North Haven Formation could be a remnant of the crust of the Kronos. What started as a back-arc basin developed into an ocean wide enough that, by the Late Silurian, the migration of fauna across it was barred. The Kronos Ocean had its maximum width in early Late Silurian, and after that time the ocean started to close by subduction westward beneath medial New England and possibly intermittently eastward along the west side of Avalon. A substantial volcanic arc was formed across the northern Maine and the Miramichi belts during the latest Silurian and into the Early Devonian, and local accumulations of arc-type volcanic rocks may be superimposed on the rift-type volcanic rocks in the Ellsworth belt. By Siegenian time the Kronos Ocean had been reduced in width so flysch spread across it in front of advancing Avalon. This flysch deposit blanketed much of the Casco Bay - south-central Maine and the northern Maine belts, nearly smothered the remaining volcanic arc, and spread westward through gaps in the arc into the foreland beyond. In a short span of time, between the Emsian and the Givetian, Avalon collided with North America in the Acadian orogeny, which imposed a complex array of structural features on the rocks of the region. This scheme of tectonic events is compatible with that proposed by Robinson and Hall (1980) for southern New England.

Figure 5b depicts an alternative tectonic scheme, permitted by the uncertainties of the stratigraphic relationships, in which the Kronos Ocean is assumed to have persisted from Cambrian time until the Early Devonian. The Celtic archipelago, underlain by the northern Maine, Miramichi, and Casco Bay - south-central Maine belts, as in the previous scheme was separated to the west by the Iapetus Ocean from North America, but to the east it was separated by the Kronos Ocean from Avalon. The Celtic archipelago existed from at least Early Cambrian time into the Llanvirnian. Avalon, which included the Benner Hill, Islesboro-Rockport, St. Croix, Ellsworth, and Saint John belts, may have been a part of Gondwana, but after the opening of the Rheic Ocean in the Early Silurian it was attached to northern Europe outboard of Gondwana into Carboniferous time (Cocks and Fortey, 1982).

As in the scheme shown in Figure 5a, the eastern Iapetus was closed along a west-dipping subduction zone beneath the Chain Lakes terrane, culminating in a collision between the Celtic archipelago and the Chain Lakes terrane in latest Cambrian to earliest Ordovician time, in the Penobscottian orogeny. The western Iapetus Ocean closed along an east-dipping subduction zone beneath the western margin of the Chain Lakes terrane, affecting the northern Maine belt by the construction of a volcanic arc across it and affecting the Miramichi belt by the venting of within-plate volcanic rocks suggesting an active back-arc basin. The amalgamated Chain Lakes and Celtic archipelago terrane collided with North America in Caradocian time producing the Taconian orogeny.

Special to this model (Fig. 5b), the Kronos Ocean of unknown width separated the Celtic archipelago and Avalon from Cambrian time to the end of the Early Devonian. The Kronos Ocean began to close in the Silurian on subduction zones along both of its margins as suggested by McKerrow and Zeigler (1971) and Bradley (1983). Subduction along the east side of the Kronos is thought to have been initiated in the Llandoverian and continued into the Gedinnian, because of the occurrence of possible arc-type volcanic rocks of that age range in the Ellsworth belt (Thirlwall, 1988; for opposing view see Gates and Moench, 1981; Pinette and Osberg, 1989). Subduction along the west margin of the Kronos Ocean was initiated in Late Silurian time and continued intermittently until the Emsian.

Subduction beneath the amalgamated Celtic archipelago - Chain Lakes - North American craton produced a large volcanic arc in the northern Maine and Miramichi belts. The situation on the western margin of the Kronos in the Late Silurian was perhaps analogous to modern subduction of the Mediterranean plate beneath the Aegean plate near Crete, where the Mediterranean oceanic plate is under compression, and the Aegean continental plate is under extension (Le Pichon, 1982). This has produced, in the Aegean plate, a series of graben-like, sediment-filled oceanic basins that extend back into the volcanic arc. We imagine that horst and graben features in the northern Maine, Miramichi, and Fredericton belts suggested by their thin and thick Silurian sedimentary facies could have had a similar origin.

As the Kronos Ocean closed and Avalon approached the eastern margin of North America, flysch spread westward from Avalon onto the submerged portions of North America and nearly covered the vestiges of the volcanic arc. Toward the end of Emsian time, Avalon and North America collided in the powerful Acadian orogeny, which produced great crustal thickening and intense polydeformation of the Paleozoic cover rocks. Acadian orogenic activity in this region must have been short-lived, approximately 10 Ma, because relatively undeformed Givetian red beds unconformably overlie the older Devonian and pre-Devonian sections.

Final Comment

The stratigraphic analysis which is the main basis for this study obviously must be combined with petrologic and structural information in order to make a satisfactory tectonic synthesis, but even if such information is added, the uncertainties in current knowledge in all fields make isolation of a singular model impossible. We have presented two models that are in ageement with the geology as we know it. Other interpretations of the stratigraphy and petrologic character of the volcanic rocks are possible, but we feel that most differences could be acommodated by the models presented above. For example, the Casco Bay Group and the Ellsworth Formation could conceivably be Precambrian and may not correlate with other units in the region. Even so, the range of stratigraphic permutations that could be entertained would alter the tectonic schemes presented above only in detail.

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We thank the many people who have freely exchanged information and ideas with us through the years, most of whom will challenge some of our interpretations of their data. This paper specifically has benefited from conversations and field trips with S. R. McCutcheon, Allan Ludman, D. B. Stewart, A. J. Boucot, A. M. Hussey II, R. B. Neuman, L. R. Fyffe, Peter Robinson, and members of the New England geology class at the University of Massachusetts. We appreciate the suggestions and additional information offered by reviewers A. A. Ruitenberg, R. G. Marvinney, and especially L. R. Fyffe, which led to reconsideration of many points and a nearly total revision of the text and figures. We thank the editors and the Maine Geological Survey for allowing us the space necessary to present the data upon which our interpretations are based. During manuscript preparation, Berry was supported by an N. S. F. research grant to Peter Robinson.

EDITORS' NOTE

Graptolites recently discovered in the Kendall Mountain Formation by L. S. Fyffe have been identified by John Riva as Middle Caradocian (A. Ludman, written commun., 1989). This new information arrived too late to be incorporated in the text. The Kendall Mountain Formation, shown incorrectly here (Fig. 2, col. 12, and Fig. 4a), belongs at the top of the Cookson Group, above the Calais Formation rather than below it. In addition, Ludman now favors assigning the Pocomoonshine Lake Formation to the Fredericton belt (Fig. 2, col. 16), rather than the Cookson Group (Ludman, 1987), but its age is still uncertain. While the graptolite age invalidates correlation between the Megunticook and Kendall Mountain Formations, it suggests that correlation between the Benner Hill sequence and Kendall Mountain Formation might now be considered. In addition, the volcanic component of the Kendall Mountain Formation suggests ties with the Caradocian volcanic sections to the northwest. These considerations do not detract from the conclusions of the paper, but reemphasize the regional importance of the Benner Hill faunal provinciality.

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