Polyphase Deformation in the Penobscot Bay Area, Coastal Maine

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ABSTRACT

Metasediments from portions of two terranes in the Penobscot Bay area of coastal Maine shared a common history of four phases of ductile deformation (D₁ to D₄). The lithologies involved are the Late Precambrian(?) to Ordovician(?) aged Copeland Formation, including the Rider Bluff Member, at the eastern margin of the Passagassawakeag terrane, and the Silurian(?) to Devonian(?) aged Bucksport Formation of the adjacent Bucksport terrane. D₁ (early Acadian?) produced tight to isoclinal folds with axial plane foliations that deform all three sedimentary units and the contacts between them. These units and, therefore, the Passagassawakeag and Bucksport terranes, must have been together by the time of D₁ at the latest. No observed structural or metamorphic gradient coincides with this terrane boundary in the Penobscot Bay area. A static thermal event (M₁₂) of granitic plutonism (412±14 Ma, Stricklen Ridge granite) and porphyroblast growth occurred between D₁ and D₂. D₂ (Acadian) produced open, upright folds and a strong, hinge-parallel elongation lineation under upper greenschist to lower amphibolite conditions. D₃ ductile strike-slip shear zones formed under greenschist facies to lower amphibolite conditions, and D₄ open, steeply-reclined folds formed under greenschist facies conditions. D₃ and D₄ appear to be related to the same stress system as the dextral strike-slip Norumbega fault zone (Alleghenian) and probably represent a continuum of deformation. Late brittle faults exploit D₃ shear zones and may also be related to the same stress system.

INTRODUCTION

The geology of the Maine coastal region is composed of six terranes (Stewart and Wones, 1974) (Fig. 1). The boundaries between these terranes (Fig. 1b) have been variously interpreted as strike-slip, normal, thrust, or reverse faults (Stewart and Wones, 1974; Wones, 1976, 1978; Wones and Stewart, 1976; Loiselle and Wones, 1983; Gates et al., 1984; Osberg et al., 1985). Predominantly Lower and Middle Paleozoic movements have been interpreted for the faults because they bound Lower and Middle Paleozoic rocks (Stewart and Wones, 1974) and are intruded by various Devonian and younger plutons (Fig. 1b)
J. P. Kaszuba and C. Simpson

Figure 1. Geology of coastal eastern Maine. (a) Index map of Maine (after Osberg et al., 1985). BM, Boundary Mountains anticlinorium; KCM, Kearsarge-central Maine synclinorium. Outlined area containing diagonal lines is enlarged in b. (b) Regional geology of coastal Maine (after Osberg et al., 1985; Stewart et al., 1986). C, Canada; FT, Fredericton trough; NFZ, Norumbega fault zone; TFZ, Turtle Head fault zone; L, Lucerne pluton; W, Mount Waldo pluton; PB, Penobscot Bay; SB, Sheepscot Bay area. Area outlined is enlarged in Figure 2.

(Wones, 1974). Carboniferous through Triassic and Jurassic fault reactivations have been postulated for the Norumbega fault zone because it truncates some Devonian granitic plutons (Fig. 1b) and because Triassic and Jurassic sedimentary basins have been interpreted as forming during fault movements (Wones, 1978). Although this framework is generally accepted by New England geologists, specific details are unknown. Intraterrane correlations are often tenuous, the nature of terrane boundaries is still uncertain, and specific ages of sediment deposition and metamorphism are largely unknown.

The boundary between the Passagassawakeag and Bucksport terranes (Fig. 1b) has been described both as tectonic (Stewart and Wones, 1974; Kaszuba and Wones, 1985; Osberg et al., 1985) and lithologic (Wones, 1976). This paper presents the results of a detailed structural analysis at the boundary between these terranes in a portion of the Penobscot Bay area to the west of the Lucerne pluton (Figs. 1, 2). The implications of these data for the nature of this terrane boundary and for the structural and tectonic evolution of the study area and the coastal Maine region as a whole are then discussed.
Figure 2. Simplified geologic map of study area. See Figure 1 for location. Cross section highly schematic, vertical dimension not to scale. Data for Lucerne pluton from Wones (1980).
GEOLOGY OF PENOBSCOT BAY AREA

The Penobscot Bay area of coastal Maine contains five terranes, the Vassalboro, Passagassawakeag, Bucksport, Penobscot, and Ellsworth terranes (Stewart and Wones, 1974; Zen, 1983; Gates et al., 1984; Osberg et al., 1985) (Fig. 1b). “Terrane” as used in this paper is defined by Coney et al. (1980) as a rock package characterized by internal homogeneity and continuity of stratigraphy, tectonic style, and history. Boundaries between terranes are discontinuities not explained by facies changes or unconformities.

Vassalboro Terrane

The Silurian Vassalboro Formation (Perkins and Smith, 1925; Osberg, 1968, 1979; Pankiwskyj et al., 1976) constitutes the Vassalboro terrane in the Penobscot Bay area (Fig. 1b). The Vassalboro Formation is a slightly calcareous wacke containing thin, interbedded quartz-mica sulfide-bearing phyllite/schist (Osberg, 1968; Stewart and Wones, 1974).

Passagassawakeag Terrane

Southeast of the Vassalboro terrane lie the complexly folded and metamorphosed rocks of the Passagassawakeag terrane (Fig. 1b). In the Penobscot Bay area, this terrane is composed of the Passagassawakeag Formation, including the Mixer Pond Member, the Copeland Formation, including the Rider Bluff Member, and the Stricklen Ridge granite, but only the latter three units crop out in the study area (Fig. 2). The Passagassawakeag Formation (Bickel, 1976) is a quartz-feldspar-biotite gneiss containing sillimanite-bearing metamorphic assemblages intercalated with layers and boudins of biotite schist (Stewart and Wones, 1974; Bickel, 1976; Guidotti, 1985). The Mixer Pond Member of the Passagassawakeag Formation (Bickel, 1976) is a light-colored, feldspathic quartz-oligoclase-microcline gneiss interlayered with the Passagassawakeag Formation. It also crops out within fault-bounded slices within the Norumbega fault zone.

The Copeland Formation (Trefethen, 1950; Wing, 1957) (Fig. 2) is composed of interlayered, bluish-gray pelitic schist and quartzite (Fig. 3a). It crops out in the study area (Fig. 2) near the eastern margin of the Passagassawakeag terrane and extends ca. 10 km to the southwest within fault-bounded slices of the Norumbega fault zone. Compositional grading of layers in this unit from quartzite to pelite on individual outcrops (Fig. 3a), and from east to west across the study area, is interpreted to represent bedding. The Copeland Formation has been correlated on the basis of similar lithologies (Wones, 1976; Loiselle and Wones, 1983) with the Hogback Schist (Perkins and Smith, 1925; Bickel, 1976) and the Ordovician Appleton Ridge Formation (Bickel, 1976; Brookens, 1976) (both located 10-12 km west of the Mount Waldo pluton, Fig. 1b), and the Late Precambrian to Ordovician Cape Elizabeth Formation of the Sheepscot Bay area (Hussey, 1968, 1985; Brookens and Hussey, 1978) (Fig. 1b). The contact between the Copeland and Passagassawakeag Formations is not exposed in the study area. It is not known whether this contact is lithologic or tectonic, and the relative structural position between these two units is also unknown.

The Rider Bluff Member (informal usage by Stewart and Wones, 1974; Wones, 1976) of the Copeland Formation is a green, finely-laminated pelitic siltstone. It occurs only in the study area, where it crops out as a <2 km-wide, discontinuous belt along the eastern margin of the Copeland Formation (Fig. 2). Local magnetite- or garnet-rich layers may represent bedding. It has been assigned to the Copeland Formation on the basis of lithological similarities (Wones, unpublished data). Kaszuba and Wones (1985) interpreted the Rider Bluff Member as ductilely deformed Copeland Formation. Subsequent work revealed that the two units are lithologically distinct where undeformed, but that ductilely deformed Rider Bluff Member is almost indistinguishable from deformed Copeland Formation. No fossils have been found in the Rider Bluff Member, and no radiometric dating has been attempted. The contact between the Rider Bluff Member and the Copeland Formation is not exposed.

The area on Figure 2 showing the extent of Stricklen Ridge granite (informal usage by Stewart and Wones, 1974; Wones, 1974, 1976) outlines the occurrence of a two-mica, garnet-bearing leucogranite that intruded the Copeland Formation and Rider Bluff Member as dikes, sills, and very small stocks. The regional stratigraphy is traceable through the area intruded by granite. Overall, the granite comprises less than one-eighth of the total map area outlined as Stricklen Ridge granite except where two stocks of granite crop out (Fig. 2). Within these two stocks, the granite comprises from 50% to 90% of the total map area. Dikes of Stricklen Ridge granite have been assigned a 412±14 Ma age based on concordant U-Pb data from zircons (Zartman and Gallego, 1979).

The Vassalboro and Passagassawakeag terranes are separated by the Norumbega fault zone in the northern Penobscot Bay area (Wones and Stewart, 1976) and by a thrust fault farther to the southwest (Fig. 1b). The Norumbega fault zone is a steeply-dipping, northeast-trending zone of ductile and brittle deformation (Devonian or younger) which consists of five distinct fault traces (Johnson and Wones, 1984). These fault traces bound slices of the units through which the fault extends. At least 13 km of right-lateral strike-slip movement is suggested by displacement of a Devonian syenite body north of the Lucerne pluton (Osberg et al., 1985). Johnson and Wones (1984) suggest a 35 km of right-lateral offset along the fault zone and a component of south-side-up dip-slip movement based on best-fit considerations of units across it. Loiselle and Wones (1983) first suggested that the Norumbega fault zone is a major crustal fracture at depth, separating two different source regions for plutonic rocks exposed on either side of the fault zone.
Bucksport Terrane

The Bucksport terrane (Fig. 1b) is composed of the Bucksport Formation (Trefethen, 1950; Wing, 1957), a greenschist facies calcareous siltstone that is finely-laminated and sometimes interlayered with sulfide-bearing pelite beds. This unit is recognized from southwest of Sheepscot Bay to the Lucerne pluton (Fig. 1b). In the study area, it crops out between the Passagassawakeag terrane and the Lucerne pluton (Fig. 2). The Bucksport Formation is lithologically similar to the Silurian Vassalboro Formation and has been correlated with it (Stewart and Wones, 1974; Wones, 1976, 1980; Osberg, 1980; Loiselle and Wones, 1983; Gates et al., 1984; Hussey, 1984). However, no fossils are found within the Bucksport Formation, and it cannot be traced directly into the Vassalboro Formation. Correlations are based strictly on lithologic similarities between units that occur on either side of the Norumbega fault zone. The Late Silurian to Early Devonian Flume Ridge Formation (Ruitenbergh, 1967; Ruitenbergh and Ludman, 1978), part of the Fredericton trough (Fig. 1b), has also been correlated with the Bucksport Formation based on lithologic similarities (Wones, 1980; Ludman, 1981; Gates et al., 1984).

The boundary between the Passagassawakeag and Bucksport terranes in the Penobscot Bay area has previously been interpreted as a normal or reverse fault (Stewart and Wones, 1974), an unconformity (Wones, 1976), and a thrust fault (Kasuba and Wones, 1985; Osberg et al., 1985). To the southwest in the Sheepscot Bay area (Fig. 1b), the Passagassawakeag terrane has been interpreted to structurally overlie the Bucksport terrane along a premetamorphic thrust fault (Hussey, 1985). The boundary is not exposed in the study area and no observed strain gradient or structural discontinuity coincides with it. The boundary between the Vassalboro and Bucksport terranes is the Norumbega fault zone (Fig. 1b) (Stewart and Wones, 1974).
TABLE I. CHARACTERISTICS OF DEFORMATIONAL AND METAMORPHIC EVENTS. NFZ, NORUMBEGA FAULT ZONE.

<table>
<thead>
<tr>
<th>Feature</th>
<th>Event</th>
<th>D1</th>
<th>M1,2</th>
<th>D2</th>
<th>D3</th>
<th>D4</th>
<th>Brittle</th>
</tr>
</thead>
<tbody>
<tr>
<td>Structures</td>
<td>- F1 folds</td>
<td>- porphyroblasts overgrow S1</td>
<td>- L2 mineral lineation and</td>
<td>- ductile shear zones</td>
<td>- F4 folds</td>
<td>- brittle joints and faults</td>
<td></td>
</tr>
<tr>
<td></td>
<td>- S1 axial plane foliation</td>
<td></td>
<td>crenulation - weak S2 axial plane foliation</td>
<td>- weak L4 mineral lineation and crenulation - S3 mylonitic foliation and compositional layering</td>
<td>- weak S4 axial plane foliation</td>
<td>- slickensides - veins filled with quartz ± hematite</td>
<td></td>
</tr>
<tr>
<td>Metamorphic conditions</td>
<td>greenschist (?)</td>
<td>amphibolite</td>
<td>upper greenschist to lower amphibolite</td>
<td>upper greenschist to lower amphibolite</td>
<td>greenschist</td>
<td>sub-greenschist</td>
<td></td>
</tr>
<tr>
<td>Relative timing</td>
<td>pre-Stricklen Ridge granite intrusion</td>
<td>between D1 &amp; D2, syn-granite (?) intrusion</td>
<td>post-granite intrusion</td>
<td>post-L2</td>
<td>post-mylonite</td>
<td>after all ductile deformation</td>
<td></td>
</tr>
<tr>
<td>Orogeny</td>
<td>pre-412 ± 14 Ma, early Acadian (?), possibly Taconic (?)</td>
<td>Acadian (?), -412 ± 14 Ma</td>
<td>Acadian (380 ± 4 Ma), syn-NFZ, Alleghanian</td>
<td>post-D3 shear zones, syn-NFZ, Alleghenian</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

**Penobscot Terrane**

To the south of the Bucksport terrane, the Penobscot terrane (Fig. 1b) contains the Ordovician- to Silurian-aged Penobscot Formation (Boucot et al., 1972; Smith et al., 1907), a graphitic, sulfidic greenschist facies schist composed of interlayered siltstone and pelite (Stewart and Wones, 1974). The contact between the Bucksport and Penobscot terranes has been interpreted as both a thrust fault (Loiselle and Wones, 1983; Gates et al., 1984; Osberg et al., 1985) and an unconformity (Wones, 1976).

**Ellsworth Terrane**

Southeast of the Penobscot terrane lies the Ellsworth terrane (Fig. 1b). This terrane contains the polydeformed, Precambrian- to Ordovician-aged Ellsworth Formation (Smith et al., 1907), which is feldspathic and chlorite-rich (Stewart and Wones, 1974).

**STRUCTURAL GEOLOGY OF THE PASSAGASSAWAKEAG AND BUCKSPORT TERRANE BOUNDARY ZONE**

The boundary between the Passagassawakeag and Bucksport terranes (Fig. 1b) separates Precambrian (?) to Lower Paleozoic rocks to the west from Middle Paleozoic rocks to the east. The nature of the contact between these two terranes is therefore important to regional tectonic models for coastal Maine. For this reason, the mesoscopic and microscopic structures in the rocks that comprise the terrane boundary zone in the Penobscot Bay area have been studied in detail (Fig. 2). Temporal relationships among different small-scale structures were established in outcrop using overprinting relationships. Sets of structures were correlated on the basis of orientation, style, and position within temporal sequences. Microstructural characteristics of the various fabrics and the relationship of these fabrics to metamorphic mineral growth were determined in thin section. Using these methods, four major groups of structures (D1, D2, D3, and D4) (Table 1) have been distinguished in the study area (Fig. 2). A separate event of granitic plutonism and thermal metamorphism (M1-2) is recognized and interpreted to have occurred between D1 and D2 (Kaszuba, 1986) (Table 1).

**D1 Deformation**

The earliest recognizable deformation phase (D1) (Table 1) produced F1 folds and a strong axial plane foliation (S1) (Figs. 2, 3b) defined by a uniform compositional layering. F1 folds occur in the Copeland Formation, the Rider Bluff Member, and the Bucksport Formation, but are prevalent in the Copeland Formation in the southern part of the study area. They deform bedding (S0) (Fig. 3b) and are cut by dikes of Stricklen Ridge granite in the Copeland Formation (Fig. 3c). Style and geometry of folds vary among the three metasedimentary units in the area (Table 2). F1 folds in the Copeland Formation are steeply-reclined to vertical (classification of Fleuty, 1964) and tight to isoclinal (Figs. 3c, 3d). Their axial planes strike predominantly
Polyphase deformation, coastal Maine

TABLE 2. CHARACTERISTICS OF FOLD GENERATIONS. FOLD CLASSIFICATION OF FLEUTY (1964).

<table>
<thead>
<tr>
<th></th>
<th>F1</th>
<th></th>
<th>F2</th>
<th></th>
<th>F3</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Style</td>
<td>steeply reclin. to vertical; tight to isoclinal</td>
<td>recumbent to gently reclin. isoclines</td>
<td>incl. tight</td>
<td>upright to incl. open to tight</td>
<td>steeply reclin. to vertical; open to tight</td>
<td></td>
</tr>
<tr>
<td>Wavelength</td>
<td>rarely &gt;0.5 m</td>
<td>rarely &gt;10 cm</td>
<td>---</td>
<td>6 cm to 6 m</td>
<td>6 cm to 1.5 m</td>
<td></td>
</tr>
<tr>
<td>Amplitude: wavelength</td>
<td>4:1 to 8:1</td>
<td>4:1 to 8:1</td>
<td>---</td>
<td>1:1 to 2:1</td>
<td>1:9 to 1:2</td>
<td></td>
</tr>
<tr>
<td>Relative abundance</td>
<td>abundant</td>
<td>rare</td>
<td>rare</td>
<td>common</td>
<td>abundant</td>
<td>rare</td>
</tr>
<tr>
<td>Structures it deforms</td>
<td>S0</td>
<td>quartz veins</td>
<td>S0</td>
<td>S1</td>
<td>S1, L1</td>
<td>S0, S1</td>
</tr>
</tbody>
</table>

east-west (Fig. 4a), and fold hinges plunge steeply to the east (Fig. 4b). F1 folds in the Rider Bluff Member are recumbent to gently-reclined isoclines that deform quartz veins. In the Bucksport Formation, F1 folds are tight and inclined (Fig. 3b) with hinges that plunge moderately to the northeast. Preliminary work by Kaszuba and Wones (1985) did not recognize F1 folds in the Bucksport Formation.

Foliation measured in outcrops where fold closures are not observed strike predominantly north-northeast and dip steeply west (Fig. 4c) in contrast to the F1 axial plane data measured within fold noses (Fig. 4a). These foliations may either be a bedding-parallel foliation, which is common to the area and relatively easy to recognize, or the S1 axial plane foliation. The S1 is easily confused with the bedding-parallel foliation. Therefore the data plotted in Fig. 4c probably represent the bedding-parallel foliation with a small component of S1. A point maximum such as that displayed in Fig. 4c represents an isoclinal sequence, one limb of a more open fold, or a homoclinal sequence. The data plotted in Figs. 4a and 4c preclude the presence of an isoclinal sequence, and the presence of an open fold is unlikely because there is no evidence for the existence of the other limb. So/S1(?) measurements (Fig. 4c) therefore represent a homoclinal sequence, and this subvertical section of stratigraphy within the study area is folded in places by F1.

S1 compositional layering in the Copeland Formation is defined by alternating quartz- and mica-rich layers up to 5 mm thick. Plagioclase (An21-28) exhibits undulatory extinction and larger grains contain subgrains along their margins. Quartz contains undulatory extinction, elongate 20-40 µm diameter subgrains and recrystallized new grains, deformation bands, and core-and-mantle structures (White, 1973, 1976). In mica-rich layers, plagioclase and quartz grain boundaries are pinned by adjacent micas and the resultant elongate grains are oriented subparallel to S1. S1 is enclosed in porphyroblasts interpreted to have grown between D1 and D2 (Kaszuba, 1986) (event labeled M1-2, see Table 1). In quartz-rich layers, quartz grain-shape preferred orientations (elongate subgrains and recrystallized new grains) and deformation bands are parallel to axial planes of F2 (Fig. 5a) and F4 microfolds, a texture also developed in the Rider Bluff Member and calcareous Bucksport Formation. Quartz veins in the Copeland Formation and Rider Bluff Member are sometimes flattened within S1 and extended parallel to quartz rods that define an L1 mineral lineation.

S1 compositional layering in the Rider Bluff Member is defined by millimeter-scale quartz- (90% quartz) and chlorite-rich (50% quartz, 50% chlorite) laminae. Magnetite-rich laminae, 0.1-0.3 mm diameter elongate quartz and plagioclase grains, and fibers of chlorite and white mica are preferentially oriented parallel to S1. Plagioclase and quartz grains display undulatory extinction and larger quartz grains (up to 1 mm in diameter) also contain elongate subgrains and recrystallized new grains, deformation bands, and core-and-mantle structures.

The character of S1 in the Bucksport Formation varies with the different lithologies and grain sizes of that unit. In siltstone containing little to no carbonate, S1 is a continuous to discontinuous cleavage (classification of Powell, 1979) (Fig. 5b). A continuous, slaty cleavage occurs in very fine-grained siltstone, whereas coarser-grained siltstone displays a 0.1 mm-scale discontinuous cleavage that ranges from smooth to rough to anastomosing. A mm-scale compositional layering is defined by
Figure 4. Lower hemisphere, equal area stereographic projections of D1 field data. For all projections, original orientation data is plotted on left and contoured on right. For all contoured projections, highest density contour interval is shaded, second highest is stippled. (a) Poles to axial surfaces of F1 folds. Data measured where fold closures occur. (b) F1 fold hinges. (c) Poles to So and S1 foliations. Data measured where fold closures not observed.
Figure 5. Microstructural characteristics of deformational features. (a) Elongate garnet (Gt) in Copeland Formation. Garnet is restricted to pelitic compositional layer. Elongate quartz subgrains and recrystallized new grains parallel axial surface of F2. Crossed nicols. Scale bar = 200 µm. (b) Non-calcareous siltstone of Bucksport Formation. Elongate quartz grains (Q) and fibrous beards within pressure shadows (arrowhead) define S1 and are deflected into and truncated by S2. Bt, biotite. Plane polarized light. Scale bar = 100 µm. (c) Calcareous Bucksport Formation. S1 compositional layering contains quartz veins (Q) and M1-2 amphibole porphyroblasts (arrowhead) and is deformed by F2. F1 microfold preserved by quartz vein. Plane polarized light. Scale bar = 500 µm. (d) and (e) Mylonitic Stricklen Ridge granite. Quartz ribbon (Q), mica fish (muscovite, Mu), and S3 mylonitic foliation are folded by F4. Elongate quartz subgrains and recrystallized new grains in quartz ribbons parallel F4 axial surface. Crossed nicols. Scale bar = 100 µm.
alternating quartz- and phyllosilicate-rich layers in all three types of siltstone. Quartz grains display undulatory extinction, and larger grains (up to 0.5 mm in diameter) also contain ca. 20 μm diameter equant subgrains and recrystallized new grains. Adjacent quartz grains may also be sutured together and/or indent each other. Beards in quartz pressure shadows contain fibrous quartz, white mica, and chlorite that are oriented parallel to S1 (Fig. 5b). Cleavage domains contain chlorite and sometimes pyrite, and may truncate quartz grains (Fig. 5b). S1 in carbonate-rich Bucksport Formation is a compositional layering up to 1 cm thick (Fig. 5c) that is defined by a range in modal composition of phyllosilicates, calcite, quartz, and plagioclase. Quartz in quartz-rich compositional layers occurs as elongate subgrains and recrystallized new grains.

**D2 Deformation**

The second deformation phase (D2) (Table 1) produced F2 folds that deform S1 in all three metasedimentary units and fold dikes of Stricklen Ridge granite in the Copeland Formation. F2 folds (Table 2) are open to tight, upright to inclined folds. Axial plane and hinge line orientations are grouped into two geographic domains. In the northern half of the study area, axial planes are subvertical and strike northeast (Fig. 6a). Hinge lines plunge gently northeast (Fig. 6b). In the southern half of the study area, axial planes strike north-south and range in dip (Fig. 6c). Hinge lines plunge at moderate angles to the north-northeast and south-southwest (Fig. 6d).

A strong mineral lineation (L2) occurs parallel to F2 fold hinges throughout the study area (compare Figs. 6b and 6e), but is best developed in the north. The L2 mineral lineation is defined by a strong grain-shape preferred orientation of elongate quartz grains in quartz veins and quartz-rich layers, and by elongate micas, feldspars, and garnet aggregates in pelitic layers. In Stricklen Ridge granite, L2 is defined by quartz rods and by stretched and reoriented feldspars and micas. In outcrop, F2 folds often appear as a crenulation of S1; their fold hinges were measured as L2 and are included in the data plotted in Figure 6e.

In the Copeland Formation, asymmetric F2 microfolds deform S1. Quartz displays undulatory extinction and core-and-mantle structures. Deformation bands, elongate subgrains, and recrystallized new grains in quartz-rich layers display grain-shape preferred orientations parallel to F2 axial planes (Fig. 5a) and quartz rods parallel L2. These quartz textures are also developed in the Rider Bluff Member and Bucksport Formation.

Biotite and muscovite are recrystallized into an S2 foliation which is axial planar to F2 microfolds and into polygonal grains in F2 microfold noses. Micas are also locally folded and kinked with kink band boundaries oriented subparallel to F2 axial planes, and elongate micas are also preferentially oriented parallel to L2. S2 is not as pronounced as S1 because the mica foliation is not as pervasive and no new compositional layer is developed.

In the Rider Bluff Member, S2 is a variable, mm-scale zonal crenulation cleavage (Powell, 1979) that deforms S1. White mica and chlorite in cleavage zones (<1 mm thick) are recrystallized parallel to S2. M1-2 albite porphyroblasts display undulatory extinction, pericline twins, and <1 mm diameter recrystallized grains.

**D3 Shear Zones**

The Norumbega fault zone is a major right-lateral strike-slip fault which strikes ca. 060° and crops out just northwest of the study area (Fig. 1b). It has a long history of Paleozoic ductile and Late Paleozoic and Mesozoic brittle deformation (Wones and Stewart, 1976; Wones, 1978; Loiselle and Wones, 1983; Johnson and Wones, 1984). Ductile shear zones are ubiquitous in the study area. They occur at both outcrop- and map-scales, and offset contacts and juxtapose units within their boundaries. The shear zones deform Stricklen Ridge granites and L2 and are deformed by later folds (F4, see below), therefore they are labeled D3 (Table 1). The majority of shear zones are superimposed on the Norumbega fault zone and display right-lateral movements as determined by shear bands in metasediments (Platt and Vissers, 1980) and S (schistosity) and C (cisailement, shear) foliations in granites (Berthé et al., 1979; Simpson and Schmid, 1983; Lister and Snake, 1984; Simpson, 1984). One major right-lateral shear zone is nearly parallel to the Norumbega fault zone trend, while another major shear zone is orthogonal to the Norumbega fault zone and displays left-lateral movements (Fig. 2). Conjugate sets of vertical northeast-trending dextral and north- to northwest-trending sinistral ductile shear zones and brittle faults and fractures occur along the entire extent of the Norumbega fault zone (Wones, 1978; Loiselle and Wones, 1983; Johnson and Wones, 1984).

In ductile shear zones developed in Stricklen Ridge granites, quartz shows ribbon structure (type 2B of Boullier and Bouchez, 1978) and 20-50 μm diameter elongate subgrains and dynami-
Polyphase deformation, coastal Maine

(a) F2 AXIAL SURFACES, NORTH

N

n = 15

- Copeland Formation
- Rider Bluff Member
- Bucksport Formation

(b) F2 HINGES, NORTH

N

n = 15

(c) F2 AXIAL SURFACES, SOUTH

N

n = 25

contours at 1, 3, and 5 points per 4.0% of circle’s area

(d) F2 HINGES, SOUTH

N

n = 25

contours at 1, 3, and 5 points per 4.0% of circle’s area

(e) L2 LINEATIONS

N

n = 32

- L2 mineral elongation lineations
- Crenulation axes

Figure 6. Lower hemisphere, equal area stereographic projections of D2 field data. For all paired projections, original orientation data is plotted on left and contoured on right. For all contoured projections, highest density contour interval is shaded, second highest is stippled. (a) Poles to axial surfaces of F2 folds in all metasedimentary units in northern half of study area. (b) F2 fold hinges in all metasedimentary units in northern half of study area. (c) Poles to axial surfaces of F2 folds in all units in southern half of study area. (d) F2 fold hinges in all units in southern half of study area. (e) L2 mineral elongation lineations and crenulation axes in all metasedimentary units in northern half of study area.
cally recrystallized new grains. Grain-shape preferred orientations of these grains are either oblique to the macroscopic foliation (e.g. Simpson and Schmid, 1983, p. 1285) or parallel to axial planes of F4 microfolds when these are present (Figs. 5d, 5e). Mylonitic Copeland Formation displays similar quartz textures, fine-grained, recrystallized tails of mica fish (asymmetric mica porphyroclasts, e.g. Lister and Snake, 1984), and rigid porphyroclasts wrapped with mica. Large porphyroclasts (ca. 1-2 mm in diameter, although ca. 1 cm diameter and larger porphyroclasts occur in deformed pegmatites) of plagioclase (An10-15), microcline, and orthoclase display brittle fractures, undulatory extinction, and dynamically recrystallized tails. Elongate, 10-20 µm diameter subgrains and recrystallized new grains occur within the tails and along intragranular fractures and grain margins of these porphyroclasts. Smaller feldspar porphyroclasts (<0.5 mm) also contain core and mantle structures; the subgrains and new grains occur within the 50-70 µm thick mantles. Crystal plastic microstructures are more extensively developed in K-feldspar than in plagioclase porphyroclasts.

Ductile shear zones in the Copeland Formation display a mm- to µm-scale compositional layering defined by alternating quartz- and mica-rich (usually muscovite) layers and quartz ribbons (types 2A and 2B of Boulleir and Bouchez, 1978). In the Rider Bluff Member and Bucksport Formation, ductile shear zones contain abundant M1-2 porphyroclasts (now asymmetric porphyroclasts) distributed in a foliated matrix of ≤50 µm diameter phyllosilicates and quartz.

Shear bands are developed in the metasedimentary units outside the main shear zone boundaries. They range in thickness (ca. 40 µm to 0.5 mm) and spacing (<1 mm to 5 mm and larger), and contain recrystallized phyllosilicates and quartz and concentrations of magnetite, ilmenite, and pyrite. Preexisting foliations are asymptomatically deflected into the shear bands, and rigid mineral grains (e.g. albite porphyroclasts) are locally offset across them.

Microstructural criteria for movement direction in the shear zones are internally consistent and support field observations. These criteria include asymmetric feldspar, tourmaline, and garnet porphyroclasts; mica fish (Lister and Snake, 1984); antithetic and synthetic offset along microfractures in tourmaline and garnet and along microfractures and cleavages in feldspars (Simpson and Schmid, 1983; Simpson, 1986); asymmetric microfolds; elongate, dynamically recrystallized quartz grains (Simpson and Schmid, 1983); S and C foliations in granites (Berthel et al., 1979; Simpson and Schmid, 1983; Lister and Snake, 1984; Simpson, 1984); and shear bands in metasediments (Platt and Vissers, 1980).

**D4 Deformation**

The youngest observed deformation phase (D4) (Table 1) produced F4 folds that occur in all three metasedimentary units, but are prevalent in the Copeland Formation in the southern portion of the study area. They deform S0, D1 (Fig. 3d), D2, and D3 structures. F4 folds (Table 2) are open to tight, steeply-reclined to vertical folds (Fig. 3d). Axial planes strike predominantly north-south (Fig. 7a) and hinge lines are steeply-plunging (Fig. 7b). F4 microfold hinges were measured in outcrop as a lineation (L4). An L4 mineral lineation defined by elongate micas and quartz rods occurs parallel to these fold hinges, but it is not as pervasive as the L2 mineral lineation.

F1 and F4 fold hinges are subparallel (Figs. 3d, 4b, 7b), suggesting that the range of F1 axial plane orientations (Fig. 4a) may be partly the result of refolding about the F4 fold axis. In addition, the dominant distribution of F2 fold orientations may be controlled either by the superposition of F2 on F1, or by refolding of F2 by F4. However, a change in strike of F2 axial planes from north-south (Fig. 6c) to northeast-southwest (Fig. 6a) occurs with closer proximity to the Norumbega fault zone. F2 fold hinge orientations also change from moderate north-northeast and south-southwest plunges (Fig. 6d) to gentle northeast plunges (Fig. 6b) as the Norumbega fault zone is approached. These observations lead us to suggest that right-lateral strike-slip movement on the Norumbega fault zone reoriented F2 structures in the northern portion of the study area.

In thin section, F4 microfolds in mylonitic and non-mylonitic Copeland Formation occur as crenulations of S1 compositional layers and F2 fold limbs, especially in pelitic layers. Micas and mica fish are recrystallized along grain boundaries and into polygonal grains in F4 microfold noses, and are folded by F4 microfolds (Figs. 5d, 5e). An S4 foliation, defined by a preferred orientation of recrystallized micas and kink band boundaries, occurs parallel to F4 axial planes, but it is not well-developed and is not as penetrative as S1 or S2. Quartz displays undulatory extinction and core-and-mante structure, and in quartz-rich layers elongate ≤0.12 mm diameter subgrains and recrystallized new grains and deformation bands display grain-shape preferred orientations parallel to F4 axial planes (Figs. 5d, 5e). These quartz textures are also developed in the Rider Bluff Member and Bucksport Formation.

In Bucksport Formation siltstones, symmetrical microfolding of S1 produces a zonal crenulation cleavage (S4a). Cleavage domains and microfoliations are evenly-spaced (ca. 0.5 mm) and have diffuse boundaries. Phyllosilicates, originally parallel to S1 in pressure shadows of quartz grains, now parallel S4.

**Brittle Deformation**

Brittle faults and joints are ubiquitous in the study area; they crosscut all ductile fabrics and are often filled with quartz (Fig. 3b) and/or hematite. All joint and minor brittle fault orientations are parallel to the major brittle faults in the region. Relative movement along these fractures, as determined by fibrous mineral growth (stick-sides) on, and offset across, fault planes, is the same as the D3 shear zones which they exploit. In thin section, quartz which fills the fractures occurs as polygonal grains that display no crystallographic or grain-shape preferred orientation.
RELATIONSHIP BETWEEN STRUCTURE AND METAMORPHISM

In the absence of suitable index minerals, the metamorphic conditions for each ductile event were based on the mineral assemblages present for each fabric, the microstructures displayed by these minerals, and the textural relationships among minerals of different tectonic fabrics.

The mineral assemblages interpreted to be stable during D1 are listed in Table 3. The presence of quartz pressure shadows containing fibrous minerals, the truncation of quartz grains along S1 cleavage zones, and the concentration of phyllosilicates and sulfide minerals within these zones in Bucksport Formation siltstones suggest that pressure solution (Rutter, 1976) was an important deformation mechanism in this lithology during D1. However, evaluation of deformation mechanisms operative at a specific temperature and grain-size is imprecise because these mechanisms are also functions of mineralogy, strain rate, fluid pressure, fluid composition, and confining pressure.

The porphyroblasts listed in Table 3 occur in all three metasedimentary units in the study area and share the following textural relationships: they are euhedral to subhedral and large (up to several millimeters in diameter) relative to other minerals in thin section; they overgrow and often enclose S1; porphyroblasts containing S1 lack synmetamorphic rotational textures (e.g. Spry, 1969, p. 253); and grain-shape preferred orientations of porphyroblasts parallel D2, D3, or D4 structures. These textural relationships suggest that the porphyroblasts, labeled M1-2, crystallized between D1 and D2, perhaps as part of a continuum of events at the end of D1 or the beginning of D2.
TABLE 3. INFERRED STABLE MINERAL ASSEMBLAGES IN METASEDIGMENTS.

M1-2 assemblages are listed in order of decreasing metamorphic grade. All M1-2 assemblages in the Copeland Formation include muscovite + quartz. Abbreviations: musc, muscovite; bt, biotite; qtz, quartz, olig, oligoclase (Ab12-38); mag, magnetite; ilm, ilmenite; chl, chlorite; plag, plagioclase; pyr, pyrite; Cc, calcite; sill, sillimanite; gt, garnet; tour, tourmaline; and, andalusite; stau, staurolite; ab, albite; Ca-amph, calcic amphibole; ti, titanite; ep, epidote.

<table>
<thead>
<tr>
<th>Unit</th>
<th>Event</th>
<th>D1</th>
<th>M1-2</th>
<th>D2</th>
<th>D4</th>
</tr>
</thead>
<tbody>
<tr>
<td>Copeland Formation</td>
<td>- musc + bt + qtz + olig + mag + ilm</td>
<td>- sill + bt + gt + tour and + bt + gt staur - bt + gt + tour</td>
<td>- bt + musc + qtz</td>
<td>- bt + musc + qtz + chl + ilm + hem</td>
<td></td>
</tr>
<tr>
<td>Rider Bluff Member</td>
<td>- qtz + chl + white mica + plag + mag + pyr</td>
<td>- bt + gt + ab</td>
<td>- qtz + white mica + chl + bt</td>
<td>- chl + qtz</td>
<td></td>
</tr>
<tr>
<td>Bucksport Formation</td>
<td>- qtz + white mica + chl + Cc + plag + mag + ilm + pyr</td>
<td>- Ca-amph + ab + ti bt + chl + ti + ep - bt + chl + tour</td>
<td>- bt + Cc + qtz - qtz + white mica + chl + bt + Cc + ilm + mag + pyr</td>
<td>- chl + white mica + qtz bt + Cc + qtz</td>
<td></td>
</tr>
</tbody>
</table>

Stricklen Ridge granite intrusion also occurred between D1 and D2. Sillimanite-bearing M1-2 mineral assemblages occur in Copeland Formation pelitic schist next to and within Stricklen Ridge granite stocks, and progressively lower-grade M1-2 assemblages occur with increasing distance from these stocks. These relationships suggest that M1-2 metamorphism may have occurred in a contact aureole in association with the intrusion of Stricklen Ridge granite.

The mineral assemblages interpreted to be stable in S2 and S2 are listed in Table 3. These assemblages commonly occur as retrograde products of M1-2 porphyroblasts and it is suggested that D2 deformation occurred under greenschist facies metamorphic conditions.

Although amphibole generally does not deform plastically until lower amphibolite facies (Brodie and Rutter, 1985), the presence of free water may have lowered its strength and enhanced its ductility, in a manner similar to that in feldspars (e.g. Tullis and Yund, 1980) or olivines (e.g. Mackwell et al., 1985). The observed microstructures in albite and calcic amphibole are therefore interpreted as having formed under upper greenschist to lower amphibolite facies metamorphic conditions. The presence of quartz pressure shadows containing fibrous minerals and the truncation of quartz grains along the concentration of sulfide minerals within S2 cleavage zones in siltstones of the Bucksport Formation suggest that pressure solution (Rutter, 1976) was an important S2-forming mechanism in that lithology.

Phyllosilicates are stable in the mylonitic foliation (S3), suggesting the presence of water in D3 shear zones. Temperature/pressure estimates for D3 are imprecise, but the presence of microstructures in feldspars that indicate recrystallization-accommodated dislocation creep (Tullis and Yund, 1980), and the lack of complete recovery in quartz grains, suggest deformation conditions of approximately uppermost greenschist facies. Microstructures in feldspar, quartz, and micas reported for shear zones in the Norumbega fault zone (Johnson and Wones, 1985) are similar to those in D3 shear zones in the study area, suggesting that all of these shear zones formed under similar deformation conditions, probably during the same deformation event.

The mineral assemblages interpreted to be stable in S4 (Table 3) also suggest that D4 deformation occurred under greenschist facies metamorphic conditions. The prevalence of mineral assemblages consistent with greenschist facies metamorphism in all fabrics seems a bit fortuitous. Greenschist facies assemblages clearly define the presently observable L2, S2, and S4 and are probably representative of the conditions prevalent during D2 and D4. However, it is possible that D1 contained different metamorphic assemblages which have not survived M1-2 metamorphism or subsequent deformation. It is also possible that the metamorphic conditions prevalent during D4 completely overprinted all previous metamorphic signatures in the rocks.

TECTORIC SIGNIFICANCE

There has been some debate concerning the nature of the boundary between the Passagassawakeag and Bucksport terranes (Stewart and Wones, 1974; Wones, 1976; Osberg, 1980; Gates et al., 1984; Kaszuba and Wones, 1985; Osberg et al., 1985; Hussey, 1968, 1985). Although the present study does not conclusively resolve the conflict, several observations are relevant. D1 is characterized by tight to isoclinal folding and the formation of a strong S1 axial plane cleavage. The Stricklen Ridge granite cuts D1 structures and has been assigned a 412±14 Ma age (Zartman and Gallego, 1979), therefore D1 must be pre-Late Silurian to Early Devonian in age. Since F1 folds occur within, and deform contacts between, the Copeland Formation, Rider Bluff Member, and Bucksport Formation, these units and therefore the Passagassawakeag and Bucksport terranes must have been in contact and have shared a common structural and metamorphic history by the time of D1 at the latest. Emplacement of the Passagassawakeag terrane as a thrust sheet over the Bucksport terrane could not have taken place between 390 and
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410 Ma as suggested by the preliminary work of Kaszuba and Wones (1985). Our interpretations are consistent with Ludman's (1981, 1985) suggestion that the terranes of coastal Maine were not exotic to each other by Early Devonian at the latest. Ludman (1981, 1985) based his interpretation on the lack of stratigraphic and metamorphic discontinuities between, and the structural and radiometric age continuity across, the terranes of northeast Maine (i.e. Fredericton trough area, Fig. 1b).

Several authors have grouped the Bucksport, Vassalboro, and Flume Ridge Formations into one terrane (the Silurian-Early Devonian central Maine turbidite belt) based on lithologic and age similarities (Hussey, 1978; Osberg, 1980; Gates et al., 1984). If this correlation of the Bucksport Formation with the Silurian-Devonian-aged Vassalboro and Flume Ridge Formations is valid, then D1 must be no older than Silurian. Precambrian (?) to Early Ordovician (?) units of the Passagassawakeag terrane are interpreted to structurally overlie units of the central Maine turbidite belt along a premetamorphic thrust fault because of the apparent age disparity between these units (Osberg, 1980; Gates et al., 1984; Hussey, 1985). The occurrence of a premetamorphic thrust fault between the Passagassawakeag and Buckport terranes (Copeland and Buckport Formations) is neither proven nor disproven by this study. However, any premetamorphic thrust fault separating the Passagassawakeag and Bucksport terranes could be no older than Silurian. D1 and premetamorphic thrusting would therefore be Acadian events. If the correlation of Vassalboro and Buckport Formations is not correct, however, and the Bucksport Formation is pre-Silurian in age, then D1, and premetamorphic thrusting if present, could be Taconic events. Pre-Acadian (Taconic ?) isoclinal folding has been recognized in northeastern Maine both north and east of the Fredericton trough (Ludman, 1981, 1985) (Fig. 1b). Most other first-recognized folds in coastal Maine are recumbent isoclines assigned to the Acadian orogeny (Bickel, 1976; Ludman, 1981; Hussey, 1985).

Correlation of the Bucksport Formation with the Vassalboro and Flume Ridge Formations and separation of the Passagassawakeag and Bucksport terranes by an Acadian premetamorphic thrust fault are based on extensive mapping and stratigraphic analysis (Hussey, 1978, 1985; Osberg, 1980; Gates et al., 1984; Osberg et al., 1985). No existing evidence contradicts these interpretations, and D1 is therefore interpreted as an Acadian event. Quartz veins are abundant in the Copeland Formation and Rider Bluff Member near the Passagassawakeag/Bucksport terrane boundary and are deformed by the entire structural sequence. Extensive quartz veining may reflect the presence of large amounts of fluid associated with faulting (e.g. Etheridge et al., 1983) and may mark the trace of the premetamorphic thrust fault.

D2 is characterized by open, upright folding and the formation of a strong, hinge-parallel mineral lineation under upper greenschist to lower amphibolite facies metamorphic conditions. D2 structures deform the 412±15 Ma Stricklen Ridge granite, but reportedly do not effect the 380±4 Ma Lucerne pluton (Kaszuba and Wones, 1985). D2 is therefore assumed to have occurred during the Acadian orogeny when upright folding occurred throughout New England (Osberg, 1978; Ludman, 1981; Hussey, 1985).

The orientation, sense of motion, and inferred deformation conditions (upper greenschist to lower amphibolite) for D3 shear zones suggest that these shear zones are related to the same deformation event which produced the Norumbega fault zone, a major structure near the study area. D3 is characterized by open, reclined folding under greenschist facies metamorphic conditions. The relationship between the orientation of F4 folds and the Norumbega fault zone (Fig. 7) suggests that the stress field which produced right-lateral strike-slip movement on the Norumbega fault zone could also have caused F4 folding. Eusden et al. (1986) suggest a similar relationship between folding and strike-slip faulting in southwestern Maine along a continuation of the Norumbega fault zone. D4 may therefore represent a later, lower temperature episode of the same stress event which produced the Norumbega fault zone. Alternatively, D4 could represent a discrete ductile event unrelated to the one that produced the Norumbega fault zone, but no such event is recognized elsewhere in coastal Maine (Hussey and Newberg, 1978; Hussey, 1985). D3 shear zones and D4 folds do not represent deformation associated with the intrusion of the Stricklen Ridge granite or the Lucerne pluton. D3 deforms the Stricklen Ridge granite, and shear zones similar to those of D3 occur within and deform the Lucerne pluton (Wones, 1980). Furthermore, D3 shear zones are probably related to the same deformational event as the Norumbega fault zone, and the Norumbega fault zone truncates the Lucerne pluton (Fig. 1b). The brittle joints and faults that cut all ductile structures observed in the study area may be related to the same stress field which initially produced the Norumbega fault zone, or to a later brittle event which exploited the Norumbega fault zone as a preexisting zone of weakness.

CONCLUSIONS

In the Penobsot Bay area of coastal Maine, metasediments of the Copeland Formation and Rider Bluff Member in the eastern margin of the Passagassawakeag terrane and of the Bucksport Formation in the adjacent area of the Bucksport terrane experienced four phases of ductile deformation and an interdeformational event (M1 -2) of metamorphism and granitic plutonism. The earliest recognizable deformation phase (D1) produced F1 folds and a strong axial plane foliation (S1) defined by a uniform compositional layering. Since D1 structures occur within and deform contacts between all metasedimentary units, these units and therefore the Passagassawakeag and Bucksport terranes must have been in contact and shared a common history by the time of D1 (early Acadian?) at the latest. To the southwest of the study area, these terranes are bound by a premetamorphic thrust fault. No structural or metamorphic discontinuity coincides with the boundary between these two terranes in the study area.
area, and the occurrence of a premetamorphic thrust fault is neither proven nor disproven by this study. A static thermal event ($M_{1.2}$) of granitic plutonism ($412\pm14$ Ma Stricklen Ridge granite) and porphyroblast growth occurred between $D_1$ and $D_2$ (Kaszuba, 1986). It may have been a discrete event of contact metamorphism or a part of a continuum at the end of $D_1$ or the beginning of $D_2$. The second ductile deformation event ($D_2$, Acadian) produced open, upright folds and a strong, hinge-parallel mineral elongation lineation under upper greenschist to lower amphibolite facies metamorphic conditions. The orientation, sense of motion, and inferred deformation conditions (upper greenschist to lower amphibolite) for $D_3$ shear zones suggest they are related to the same stress system that produced right-lateral strike-slip movement on the Norumbega fault zone (Alleghenian). The orientation of open, reclined $F_4$ folds that formed under greenschist facies conditions suggests $D_4$ may represent a later, lower-temperature episode related to the same stress system. Brittle structures truncate all ductile structures, and their orientation and sense of movement suggests they are either related to the same stress field which initially produced $D_3$ and $D_4$ ductile structures, or to a later brittle event which exploited $D_3$ shear zones and the Norumbega fault zone as preexisting zones of weakness.

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