The Mings Bight Ophiolite Complex, Newfoundland: Appalachian oceanic crust and mantle

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The Mings Bight Ophiolite Complex, of probable early Ordovician age, is disposed in four major thrust sheets with an eastward vergence at the northern end of the Baie Verte Lineament. This narrow ophiolitic belt, and (to the south) an adjacent zone of early Devonian volcanic rocks and sediments, were affected by a strong Acadian (Middle Devonian?) deformation between more resistant blocks consisting mainly of rocks that were deformed and regionally metamorphosed, prior to the development of the ophiolites and overlying mafic sediments and volcanic rocks, probably in Late Cambrian to earliest Ordovician. The ophiolite sequence and conformably overlying sedimentary and volcanic sequence define an overturned synclinal structure with an eastward vergence; the three western thrust sheets contain an inverted sequence, the eastern sheet is upright. The thick mafic volcaniclastic and pillow lava sequence overlying the ophiolite complex suggests that the ophiolite complex was generated as the the floor of a small rear-arc or intra-arc basin. The ophiolite complex, although dissected by faults, consists of an ordered sequence from non-cumulate tectonite harzburgite through cumulate ultramafic rocks, gabbro and sheeted dike complex to pillow lavas. The continuous, coastal exposures show the relationships between the lithologies of the ophiolite complex unusually clearly, and these are described in some detail. In particular, the relationships between the sheeted dikes and both the homogeneous upper gabbro and the pillow lavas, and the intrusive complexities and the high-temperature deformation in the layered gabbros and ultramafics, are very clearly displayed. An ocean floor fault containing diapiric serpentinite is preserved in one thrust sheet. Two new formations are proposed, for the mafic volcaniclastic sediments (Big Head Formation) and for the overlying pillow lavas (Barry-Cunningham Formation) above the ophiolite complex.

Le complexe ophiolitique de Mings Bight, datant probablement du début de l'Ordovicien, est disposé en quatre écailles de chevauchement majeures avec regard vers l'est dans la partie nord du linéament de Baie Verte. Cette étroite bande ophiolitique et (au sud) une zone adjacente de roches volcaniques et de sédiments du début du Dévonien ont été affectées par une forte déformation Acadienne (Ordovicien moyen?) entre les blocs plus résistants comprenant surtout des roches qui se sont déformées et métamorphisées régionalement avant le développement des ophiolites et des sédiments et roches volcaniques mafiques qui les recouvrent, probablement à la fin du Cambrien ou au tout début de l'Ordovicien. La séquence ophiolitique et la séquence sédimentaire et volcanique qui la recouvre en concordance définissent une structure synclinale renversée avec regard à l'est; les trois écailles de chevauchement à l'ouest contiennent une séquence inversée alors que l'écaille à l'est est verticale. La séquence épaisse de roches volcan iclastiques mafiques et de laves en coussins qui recouvre le complexe ophiolitique suggère que le complexe ophiolitique a été à l'origine le fond d'un petit bassin en arrière ou à l'intérieur de l'arc. Le complexe ophiolitique, bien que découpé par des failles, consiste en une série ordonnée allant de tectonites non cumulées de harzburgite à des roches ultramafiques cumulées, un complexe de gabbro et de dykes minces jusqu'à des laves en coussins. Les affleurements continus en bordure de la côte laissent ordinairement voir clairement les relations entre les lithologies du complexe ophiolitique et on les décrit avec assez de détail. En particulier, en observe clairement les relations entre les dykes et le gabbro supérieur homogène et les laves en coussins, les complexités de l'intrusif et les déformations à haute température dans les gabbros et les roche ultramafiques litées. Une faille de fond océanique contenant de la serpentinite diapirique a été préservée dans une des écailles. On propose deux nouvelles formations pour les sédiments volcaniclastiques mafiques (formation de Big Head) et pour les laves en coussins sus-jacentes (formation de Barry-Cunningham) au-dessus du complexe ophiolitique.

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Introduction

Fully developed ophiolite complexes (Penrose Field Conference 1972) are now regarded by many workers (e.g. Coleman 1971; Dewey and Bird 1971; Moores and Vine 1971; Thayer 1969) as slices of oceanic crust and mantle generated by plate accretion at oceanic ridges, or in marginal basins (Karig 1971) and, subsequently, tectonically emplaced during arc and (or) continental collision. The western Newfoundland Appalachians are well-endowed with excellently preserved and exposed ophiolite complexes (Stevens 1970; Church and Stevens 1971; Dewey and Bird 1971; Bird *et al.* 1971; Upadhyay *et al.* 1971; Williams 1971; Williams *et al.* 1972). These authors have described the various ophiolite complexes shown in Fig. 1 and argued for an oceanic crust-mantle origin. There has been, however, a great diversity of views concerning the relationships and significance of Newfoundland ophiolite complexes, both on a local and regional scale.

This paper has two purposes: (1) to document the local geology and its regional implications and (2) to describe the geological relationships within the ophiolite complex, information that has nowhere previously been documented and that is essential for the understanding of plate accretion processes (Dewey and Kidd 1977). We describe and interpret firstly, the overall structure and lithologic distribution, secondly, the evidence for the large thrusts (tectonic slides) that bound the major slices, thirdly, the overall lithologies and stratigraphy of the complex are documented, and fourthly, we emphasize some aspects of the ophiolite lithologies and their relationships. We intend this paper to serve as both a useful guide for those visiting these exposures, as well as a description of old oceanic crust and mantle.

Previous Work and Interpretations

Watson (1947) was the first to recognize and map a distinct association of greenstones, ultramafics and gabbro in the Baie Verte region. He noted the concordant relationships of the gabbro to the greenstones (mafic volcanics and mafic volcaniclastics) although he considered the gabbros to intrude the greenstones. He also described cumulate textures, graded layers, 'gneissic' gabbro and pegmatite within the gabbro. Neale (1958) produced a more detailed map of the Baie Verte region and noted that the close spatial relationship of gabbro and ultramafic bodies suggests a cogenetic origin. Church and Stevens (1971), Bird *et al.* (1971) and Dewey and Bird (1971) argued that the greenstones, gabbros and ultramafics of the Baie Verte area form

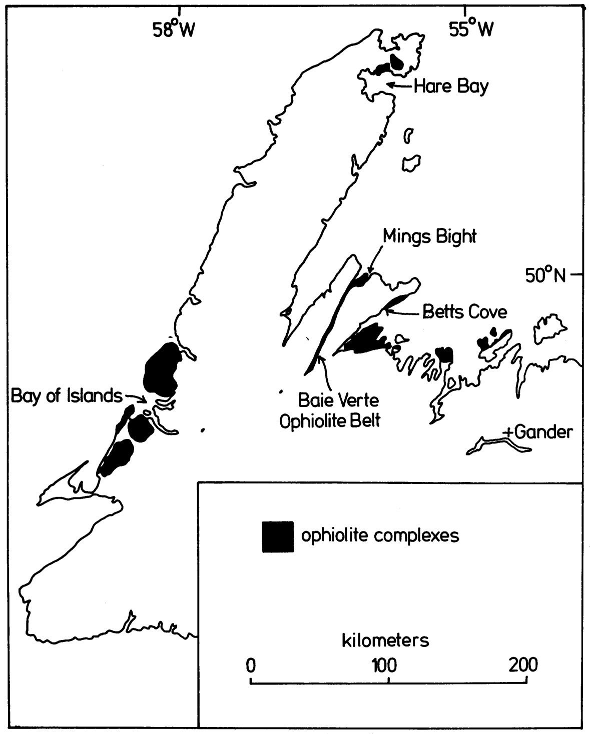

FIG. 1. Sketch map of western and north-central Newfoundland showing the distribution of late Precambrian - Lower Paleozoic ophiolite complexes.

a cogenetic ophiolite sequence representing more or less disrupted slices of Ordovician oceanic crust and mantle, generated after the polyphase deformation and high T-P metamorphism of the adjacent Fleur de Lys Supergroup. This view has been contested by Kennedy and Phillips (1971), who argue that all the ultramafic rocks in the Mings Bight area were deformed and metamorphosed with rocks of the Fleur de Lys Supergroup, and therefore belong to a basement complex. They cite alleged common schistosities in the Fleur de Lys schists and the envelopes of the ultramafic bodies, and two unconformities between the ultramafics and 'serpentine sediments' in Hammer Cove and Devils Cove (Fig. 2). They regard the 'serpentine sediments' as the basal member of the Baie Verte Group, laid down across a Fleur de Lys metamorphic basement. Our observations in this particular area lead us to disagree with these conclusions. Norman and Strong (1975) gave a generalized description and map of the Mings Bight Ophiolite Complex in a paper mainly devoted to an account of its geochemistry, which seems, to us, to be partly in error.

General Distribution of Rock Types and Structure

The two main assemblages (Fig. 2) are the Fleur de Lys Supergroup metamorphics (Church 1969)

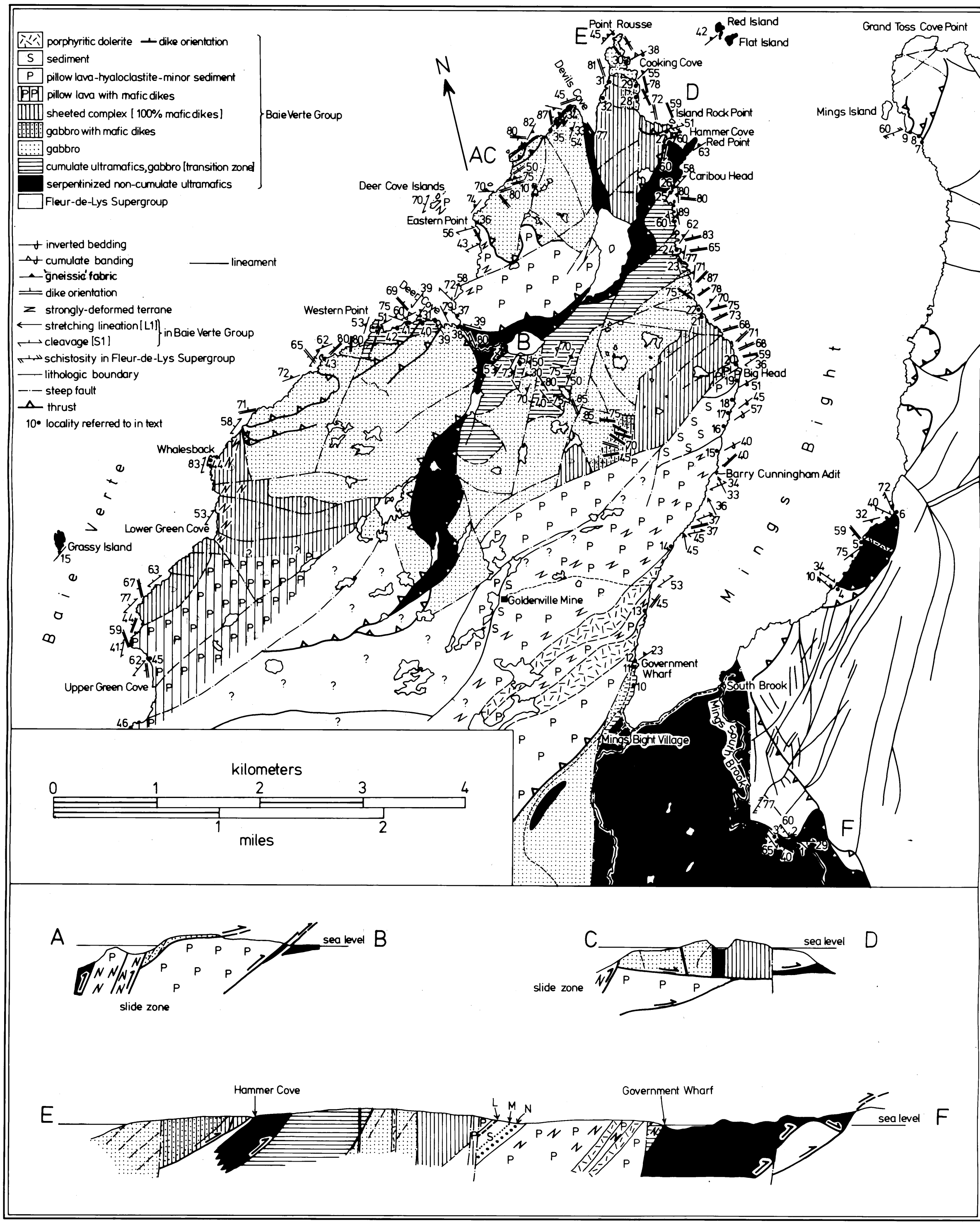

FIG. 2. Geologic map and sections of the Mings Bight Ophiolite Complex, western north-central Newfoundland. Location of Mings Bight Peninsula shown in Fig. 1. S = Big Head Formation. Pillow lavas where differentiated above Big Head Formation belong to Barry-Cunningham Formation.

and the ophiolitic Baie Verte Group (Watson 1943). In this area, the Fleur de Lys schists consist of three main lithologic units: psammitic schists with a few metagabbro sills, mafic metavolcanic schists and black graphitic schists. All are affected by polyphase deformation with an accompanying upper greenschist to epidote-amphibolite facies metamorphic recrystallization. Granitic rocks, in two small areas, were intruded late in the kinematic sequence. In contrast, the pillow lavas and sediments of the Baie Verte Group are affected by a single, moderately northwest-dipping penetrative cleavage of variable intensity, accompanied by a steeply pitching fabric lineation in the more strongly cleaved rocks. The plutonic rocks of this complex (ultramafics, gabbro, diabase) are internally undeformed, except in and near narrow syncleavage thrust zones. The fine-grained diabase and mafic volcanic rocks are recrystallized to fine-grained, low greenschist facies assemblages containing colorless actinolite - chlorite - clinozoisite/epidote - albite - sphene - calcite - quartz. Minerals in the undeformed gabbros are, with the exception of some clinopyroxene, usually similarly altered. The undeformed ultramafic rocks are always partly and usually serpentinized to a large extent and, in some areas, are partly or wholly replaced by carbonate minerals.

The main refinements that we make to the earlier geologic maps are in distinguishing 100% sheeted diabase dike complex from gabbro and pillow lava with diabase dikes, cumulate from non-cumulate ultramafics, and mafic volcaniclastics from pillow lava. With the general upright sequence of an undisrupted ophiolite complex in mind (ultramafics - gabbro - sheeted dikes - pillow lava - sediment), together with the overall medium-angle dip to the north and northwest, it can be seen (Figs. 2, 3A) that the rocks are broadly disposed in four thrust sheets, although in detail there are apparent anomalies in this pattern. The upper three sheets are inverted, whereas the fourth, lower, southeastern sheet only preserves the basal half of the sequence and appears to be upright, thus outlining a telescoped synclinal structure. The contacts between these sheets are thrust faults and the base of the lowest sheet is marked by a major thrust of ultramafic rocks over Fleur de Lys Supergroup schists.

The other main structural feature in the younger rocks is a late, pervasive, steep fault-block pattern. Where the faults cut contrasting rock units or thrusts, they always downthrow to the west and (or) north, in the opposite direction to that of the thrusts.

Throughout the area, post-cleavage, rusty-weathering, dike-like zones, usually about 1-10m wide, containing metasomatic ferroan carbonate and usually associated with quartz veins, occur sporadically in mafic rocks. Most are not directly associated with faults. Their equivalent in the carbonated ultramafic rock (e.g. near Red Point) has resulted in the recrystallization of the original carbonate minerals alongside the quartz veins, producing a spurious layered appearance.

Major Tectonic Contacts

External Tectonic Contacts: Relationships of Ultramafic Rocks to Fleur de Lys Supergroup Rocks

(a) Eastern

The greater part of the small ultramafic body exposed on the east side of Mings Bight (Fig. 2) is undeformed, variably serpentinized harzburgite. The remainder of the internal part of the body consists of a zone of shear-polyhedra serpentinite and a zone of cleaved talc-carbonate bearing serpentinite. The single cleavage in the latter zone is cut by subparallel phacoidal surfaces. Both are steeply dipping and oriented perpendicular to the elongation of the ultramafic body. These deformed zones were generated by relative movements between adjacent blocks during tectonic emplacement; they are highly oblique to metamorphic schistosities in the Fleur de Lys rocks outside the body. Two porphyritic (plagioclase) dolerite dikes cut the body; the airphoto lineament associated with the wide one is truncated at the contact of the ultramafic body with the Fleur de Lys schists. Kennedy and Phillips (1971, p. 39) call these dikes "sheets of schistose metagabbro with similar textural features to surrounding Fleur de Lys schist." However, the dikes are altered, to an albite - colorless actinolite - epidote assemblage. Apart from a few small inci-pient shear zones (Ramsay and Graham 1970) a few centimetres across, they are undeformed. The exposed southern contact of the large dike (loc. 5)[1] shows no trace of movement; the adjacent ultramafic rock shows strong metasomatic reaction with the development of a 10 cm rind of fibrous white amphibole with fibers oriented perpendicular to the dike wall. The dikes do not resemble penetratively deformed, green hornblende-bearing, coarse-grained metagabbroic rocks that occur in Fleur de Lys schists to the north.

Kennedy and Phillips (1971, p. 39) referring to the contacts of this body, state that in the "strongly schistose marginal zones ... two penetrative schis-

[1]All locality numbers quoted are shown in Fig. 2.

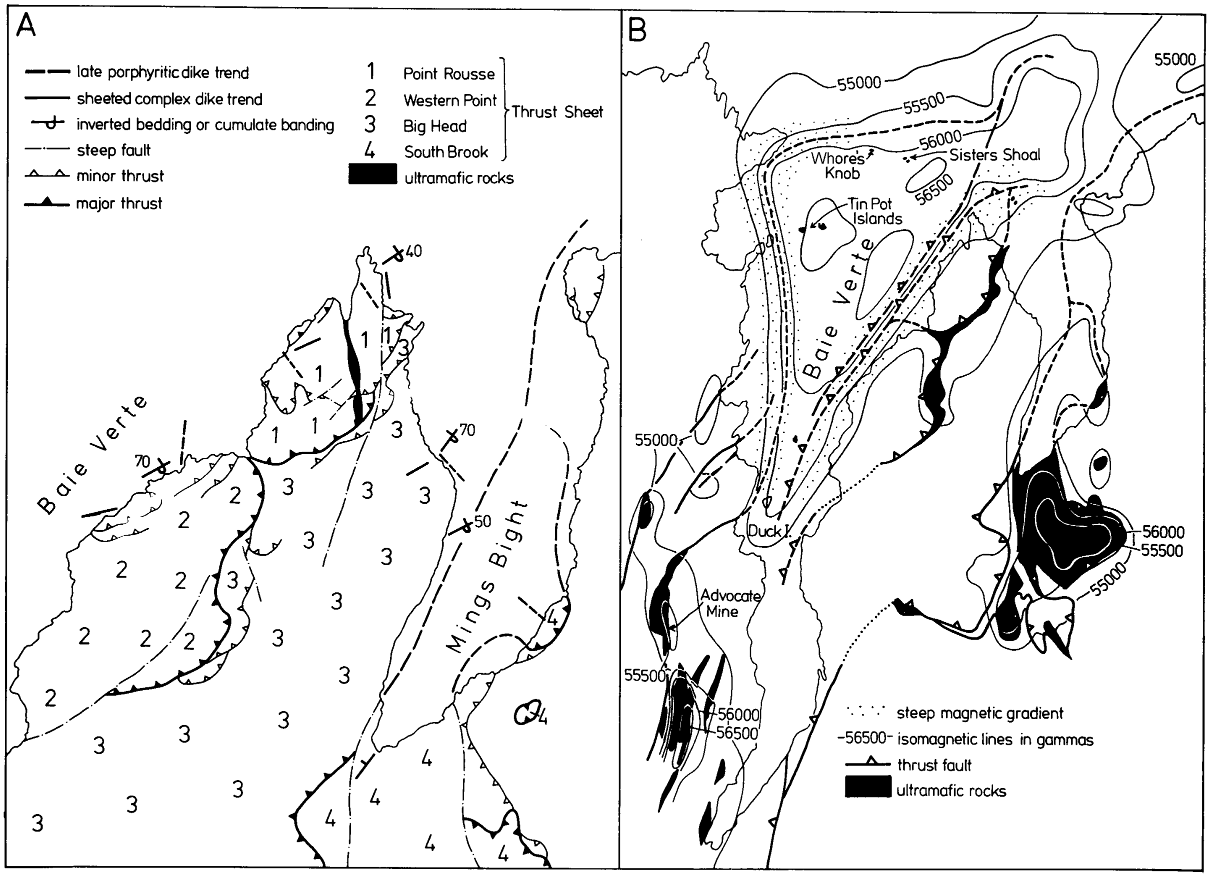

FIG. 3. (A) Outline structural map of the Mings Bight Peninsula, 'ultramafic rocks' refers only to the Devils Cove serpentinite. (B) Map of Baie Verte and Mings Bight showing the relationship between magnetic anomalies, magnetic gradients and ultramafic rocks with an interpretation of the sub-sea continuation of major thrusts. Magnetic data from Geological Survey of Canada (1968).

tosities can be seen of which the later one, S_2, is of similar attitude to the steep regionally developed S_2 schistosity of the surrounding Fleur de Lys rocks". At the northern contact of the body (loc. 6), there is an unexposed gap of about 5 m between Fleur de Lys greenschists and a talc-carbonate rock with one penetrative cleavage and a later nearly coplanar phacoidal surface that, in part, follows the earlier cleavage. Both the cleavage and the later surface are highly oblique to the penetrative composite metamorphic schistosity (S_2) in the adjacent Fleur de Lys rocks (Fig. 2). A common and well-developed crenulation of this main schistosity in the greenschists is not seen in the talc-carbonate 'schist'. Near the southern contact (loc. 4) a thin porphyritic dolerite dike forms a small outcrop. A small gap in outcrop follows and the next exposure shows a rubble of rusty Fleur de Lys psammite blocks with abundant carbonate mineralization. A few small blocks appear to be highly altered serpentinite and the north end of the outcrop shows some carbonated, strongly sheared serpentinite. The crude sheeting in this outcrop dips about 20° more steeply than the flat-lying schistosity in the Fleur de Lys psammites exposed 20 m to the south. Thus, the northern discordant contact and the southern approximately concordant contact show evidence of significant tectonic disruption. On a large scale, the body has the appearance of a relict tongue of a sheet thrust up over the Fleur de Lys rocks to the east. There is no evidence suggesting involvement of this body in Fleur de Lys deformation and metamorphism. The aeromagnetic map (Fig. 3B) suggests that it is connected under Mings Bight with the large ultramafic body exposed from the south shore of Mings Bight inland.

The latter ultramafic body is wholly serpentinized and has significant carbonate alteration (Neale 1958). We have not examined the inland part of this body, except for its eastern contact with Fleur de Lys metasediments and metavolcanics some 2 km up Mings South Brook. Here, the talc-carbonate rock has an intense cleavage and a local, later, subparallel phacoidal surface. The flat-lying (loc. 1) to moderately west-dipping (loc. 3) cleavage (Fig. 2, section EF) is transgressive to the steep composite metamorphic fabric in the local Fleur de Lys rocks. Within a metre or two of the contact, small pods of talc-carbonate 'schist', typically 0.5 m wide by several metres long, occur as septa along the foliation of the Fleur de Lys rocks (loc. 2). The pods have rinds of acicular actinolite perpendicular to the contact when in schistose mafic metavolcanic rocks. The cleavage within the pods can be concordant with, or wholly discordant to, the dominant Fleur de Lys foliation; where discordant, no corresponding foliation is seen in the Fleur de Lys rocks. We interpret the contact as a thrust, wholly later than the Fleur de Lys regional metamorphism and deformation, and the small talc-carbonate pods in the Fleur de Lys rocks as tectonic enclosures derived from the cleaved base of the ultramafic body and developed by the formation of small schuppen in the Fleur de Lys schists immediately beneath the thrust surface during the movement. Supporting evidence for the view that the Baie Verte Group rocks overthrust the Fleur de Lys Schists is seen at locs. 7-9 (Fig. 2), south of Grandtoss Cove Point. Here west-dipping thrust zones, post-kinematic with respect to regional Fleur de Lys structures, occur within the Fleur de Lys rocks. There is local buckling of the main Fleur de Lys foliation into these zones; the orientation of the regional foliation in this area a short distance away from the thrusts is highly oblique to them. Also, in the area east and southeast of South Brook village (Fig. 2), air photographs show a maze of intersecting lineaments in the Fleur de Lys schists adjacent to the contact. This strong disturbance is local, and is not seen elsewhere in the area. It is significant that, here, the exposed schists extend furthest west towards the ophiolitic rocks; the contact is inferred to be modified by a high-angle fault and the thrust surface was not far above the fractured schists immediately to the east.

(b) Western

On the western side of the Baie Verte Lineament (Fig. 3B), the contacts of the large ultramafic body underlying Baie Verte are not exposed. However, the magnetically defined western contact cuts across stratigraphy, structure and several major tectonic dislocations in the Fleur de Lys rocks exposed onshore, and must also be a major tectonic dislocation (J. T. Bursnall, personal communication, 1900). The northern contact is inferred from the aeromagnetic data to be a similar dislocation, because weak magnetic trend lines parallel to the structure in the Fleur de Lys Supergroup curve around and appear to be truncated at the rather diffusely defined contact. Although this does not necessarily mean that the body is younger than the Fleur de Lys rocks, we see no reason to divorce it from the remainder of the ophiolite suite. The same argument applies to the bodies on the eastern side of the area, although *small* tectonic pods of similar ultramafic rock and derivatives, with associated gabbro occur within the Fleur de Lys Supergroup outside this map area (Dewey and Bird 1971; Kennedy 1971; Kennedy and Phillips 1971). These are also 'alpine type' chrome-bearing depleted ultramafic rock, and must also have been derived from oceanic upper mantle.

Major Internal Tectonic Contacts

There are several ages of thrusts within the complex, but, apart from a single minor example at Island Rock Point (loc. 27), all were developed at the same time or later than that cleavage and lineation. The most important of these thrusts are syn-cleavage; they form the boundaries between sheets 2(a + b) and 3, most significantly, because of the younging direction reversal, between thrust sheets 3 and 4 (Fig. 3A). The thrust of ultramafic rocks (sheet 4) over the Fleur de Lys schists on the east side and south of Mings Bight is also most probably of this age. The syn-cleavage age of these thrust zones (tectonic slides) can be demonstrated where they adjoin pillow lavas that contain the regional cleavage and steeply pitching fabric lineation. The syn-cleavage thrusts contain a much more intensely developed regional cleavage and stretching lineation; these fabrics pass by gradation into the same ones in adjacent, normally cleaved rocks, as at Deer Cove (locs. 37-38) between thrust sheets 2a and 3, and southwest of Mings Bight village between thrust sheets 3 and 4. The orientation of these thrusts implies overthrusting from a northwest - westnorthwest direction.

In contrast, at Eastern Point (loc. 36), a thrust of undeformed gabbro over cleaved pillow lava is marked by a thin, moderately north-dipping, weakly cleaved breccia zone. On a large scale, this thrust is very gently dipping, and transgressive to the cleavage in the underlying pillow lava (Fig. 2, section AB) and, therefore, postdates the regional cleavage. The eastern end of this thrust apparently separates sheeted dikes from the altered ultramafic rocks in Hammer Cove. Other thrusts of this age are inferred to cut gabbro and cumulate ultramafic rocks between Whalesback and Deer Cove (Fig. 2). The attitude of these post-cleavage thrusts implies overthrusting from a northerly direction, in contrast to the syn-cleavage thrusts. Two other, small, flat-lying post-cleavage thrusts occur at loc. 13.

The southeastern contact of the large ultramafic body underlying the mouth of Baie Verte is not exposed. However, the intensely deformed mafic rocks at Whalesback (loc. 44) and on Deer Cove Islands (Fig. 2), several small high-angle thrust faults cutting gabbro near Devils Cove (loc. 34) and the aeromagnetic data (Fig. 3B) all suggest that the contact is a major high-angle thrust located just offshore (Fig. 2, section AB). The steep attitude of this thrust, given by the attitude of the cleavage at Whalesback and Deer Cove Islands, and the thrusts near Devils Cove, suggests that it truncates the gently dipping thrust at Eastern Point (Fig. 2, sections AB, CD). Its attitude implies a mainly vertical relative motion across it combined with a component of overthrusting from the northwest.

The attitudes of the regional cleavage and fabric lineation seen in the sections from Deer Cove to Eastern Point and from Big Head to the Government Wharf, and the overall shape of the syn-cleavage thrust zone exposed in Deer Cove, demonstrate the existence of large-scale open kink folds of the cleavage and the syn-cleavage thrusts. These kink folds are probably of about the same age as the gently dipping post-cleavage thrusts, because the straight, steep thrust bounding the ultramafic body offshore from Eastern Point is not affected and is therefore inferred to cut them.

The sequence of structural development outlined above shows abrupt changes in the motion during the overall compressive, horizontal shortening deformation. We do not suggest that these episodes were widely separated in time, but believe that they were accommodations to the changing direction of convergence and (or) geometry of the overall zone of deformation during the strong shortening across the Baie Verte Lineament. The geometry of these structures, taken in conjuction with the subvertical attitude of those in most of the rest of the Baie Verte Lineament (Kidd 1974, 1977), implies a systematic change during the deformation. The initial development of moderately west- to northwest-dipping cleavage, thrusts and the major fold axial surface are now preserved in the Mings Bight area and along the easternmost part of the Baie Verte Lineament to the south. This was followed by oversteepening to a subvertical - steeply west-dipping attitude, and continued horizontal shortening, of the rocks now preserved in the main part of the Baie Verte Lineament. The steep thrust zone bordering the southeast margin of the ultramafic body underlying the mouth of Baie Verte is the tectonic boundary, in this area, between these two contrasting structural zones.

Our interpretation of the thrust sheets does not correspond with that of Norman and Strong (1975), who place a major thrust along the narrow, vertical zone of serpentinite south from Devils Cove. This zone, containing a recognizably unique, recrystallized but unfoliated serpentinite, is not a thrust fault. The diabase dikes cutting the edge of the Devils Cove serpentinite show that it was formed on a fault during construction of the ophiolite complex as oceanic crust. Norman and Strong (1975) do not distinguish the different relative ages and orientations of thrusts, nor do they comment on the fact that the ophiolite stratigraphy is inverted except in the southeastern thrust sheet. They assert that the deformation of the Baie Verte Group is related to early Ordovician ophiolite obduction with some Acadian overprint. This may not be the case, since it can be shown 30 km south along the lineament, that the Baie Verte Group was wholly undeformed prior to deposition of the well-dated early Devonian Mic Mac Lake Group (Kidd 1974, 1977).

Stratigraphy of the Baie Verte Group in the Mings Bight Area

The Mings Bight Ophiolite Complex is included as the basal unit of the Baie Verte Group (c.f. Upadhyay et al. 1971). We do not give the units within the ophiolite complex formal stratigraphic names. Thicknesses quoted are the largest preserved within the essentially intact, tectonically bounded blocks, and are therefore minimum estimates unless otherwise stated. All thicknesses have been estimated from the map. No attempt has been made to remove the effects of ductile compressive deformation on the thicknesses of the sediment unit and the upper pillow lava unit; shapes of deformed pillows indicate that the latter was probably reduced not less than 50% during the deformation. The lithologies and stratigraphy are described starting at the base.

Mings Bight Ophiolite Complex

Non-cumulate Depleted Ultramafic Rocks

Intact ophiolite complexes always show variable thicknesses of non-cumulate, 'Alpine-type' depleted Mg-rich ultramafic rocks at their base (Smith 1958; Wilson 1959; Moores 1969; Reinhardt 1969; Davies 1971). The Mings Bight Ophiolite Complex, although it is dissected by faults, is no exception. The least altered rocks are exposed on Grassy Island (Fig. 2). This island, together with the Tin Pot Islands, the Sisters Shoal and the reef called Whore's Knob (Fig. 3B) are the only outcrops of a large, broadly triangular, ultramafic body, delineated by the aeromagnetic map, underlying the waters of Baie Verte and extending eastward across the mouth of Mings Bight (Fig. 3B). The orange-brown-weathering rocks are more than 99% harzburgite, containing 15-25% orthopyroxene. Less than 1% of dunite forms rare, near-coplanar layers 10-30 cm thick, commonly with one or two seams of chromite, one grain thick, occurring approximately in the center of the layer. Any individual dunite layer may have either slightly gradational or sharp contacts with the harzburgite. Watson (1947) reported collecting a specimen of lherzolite containing 10% diopside from Grassy Island but, despite careful search on this small island, we did not find the locality and the rock must be extremely rare. Obvious megascopic grain-shape orientation of enstatite is rare and weakly developed; however, at one locality on the eastern Tin Pot Island, centimetre-scale layering contains some layers with strongly aligned, elongate olivine grains. Kennedy (1971, Fig. 2) plots several measurements of a Fleur de Lys Supergroup metamorphic schistosity (normally coarse-grained median greenschist to amphibolite facies) on the Tin Pot Islands. No fabric compatible with these P-T conditions was seen and the attitudes plotted are very similar to that of the igneous layering.

Olivine is consistently in the range Fo_{90-95}, and seems to be slightly more magnesian in dunite than in harzburgite. Orthopyroxene is in the range En_{90-95}; rare grains show diopside lamellae. Chromite content of both rocks is less than 1%. An analysis of the harzburgite is given by Watson (1943) and shows the typical depleted composition, with a 9:1 Mg-Fe ratio, of 'Alpine-type' ultramafics. Grain size of olivine and enstatite is similar in both dunite and harzburgite, the range being 0.2-4 mm, averaging 2 mm. This is very consistent in all areas of the undeformed non-cumulate ultramafics. The texture is xenomorphic granular, but with curved grain boundaries; this is shown in outcrop by the typical ovoid shape of enstatite grains. This slight shape anisotropy is responsible for the weak mineral foliation seen in outcrop. Chromite is usually found in harzburgite as ramifying interstitial cuspate grains 0.2-2 mm, averaging 0.8 mm across; in contrast somewhat smaller subhedral to euhedral grains are characteristic of dunite. The harzburgite microstructure is not a simple plutonic crystallization fabric; it is probably highly recrystallized and annealed. The least altered specimen collected from the islands in the mouth of Baie Verte is 10% serpentinized; most are more than 50% serpentinized. The undeformed parts of the other exposed non-cumulate ultramafic bodies in the area are highly altered, ranging from near-total serpentinization (eastern Mings Bight) to partial (southern Mings Bight) to total (Red Point, Red and Flat Island) replacement by various Mg-Fe carbonate - quartz ± minor talc assemblages. However, pseudomorphic textural contrast usually remains in these altered but undeformed rocks; original lithologies can usually be at least tentatively identified by reference to less altered rocks.

The largest minimum thickness of the noncumulate ultramafic rocks is obtained from the inferred extent of the large body underlying the mouth of Baie Verte and is estimated as 1800m.

Cumulate Ultramafics ('Transition Zone')

In well-preserved ophiolite complexes, transition zones occur between non-cumulate depleted ultramafic rocks and gabbros (Moores 1969; Moores and Vine 1971; Upadhyay *et al.* 1971; Davies 1971). We have included in such a zone any layered cumulate rocks containing a dominant quantity of olivine and (or) orthopyroxene and also any section with a dominant proportion of clinopyroxenite layers. Because these rocks are interlayered with gabbros, and are overlain by gabbros containing some clinopyroxenite, the upper contact is somewhat gradational and is difficult to define precisely. The lower contact zone with non-cumulate ultramafics is not exposed in the coastal sections, except where it is faulted. The undeformed ultramafic cumulate rocks of the Mings Bight Ophiolite Complex are generally altered, but some textures and relationships are well displayed. The rocks are mostly medium-grained, with the exception of large post-cumulus 'sieve-poikilitic' oikocrysts in some olivine-rich layers.

The best exposures of these rocks are between Deer Cove and Western Point. At loc. 42, a sequence of altered (carbonate-talc) cumulate ultramafic rocks, originally mostly harzburgites, orthopyroxenites and websterites, are interlayered with some gabbroic rocks. A large xenolith (~3 x 20 m) or autolith of serpentinized dunite within these rocks is displayed in the cliff. Large 'sieve-poikilitic' oikocrysts of pseudomorphs after orthopyroxene up to 3 cm across are found in some of the ultramafic layers. The layers are typically 5-20 cm thick; a few show normal size grading. Several other layers show a gradation from altered poikilitic harzburgite to gabbro, or from serpentinized dunite to gabbro, over a thickness of about 30 cm (loc. 41). A few very fine-grained, grey layers up to 50 cm thick, now composed of talc with some magnesite, are found in the sequence (loc. 41); they are inferred to have been cumulate dunites, and greatly resemble certain rocks in Hammer Cove. Although faulted, the section seems to lose first olivine, and soon after, orthopyroxene, and gain in clinopyroxene and plagioclase upsection to the east. Layered cumulate rocks at loc. 40 consist mainly of websterite and clinopyroxenite with some gabbro. A few spectacular mineral (density) graded layers occur here, one of which also has originally inverted size-grading. Towards loc. 39, cumulate gabbros have tabular plagioclase lying flat in the layering. At loc. 39, clinopyroxenite with crystals up to 10 cm long perpendicular to the layering, occurs within layers of coarse-grained clinopyroxenite. These are probably pegmatites, but could be 'harristic cumulates', where the crystals grew up from the cumulate floor.

At several places south from Caribou Head (between locs. 23 and 25), exposures of altered talcose 'sieve-poikilitic' harzburgite, and other altered olivine and orthopyroxene-rich cumulate ultramafic rocks are intercalated within clinopyroxenites and gabbros; streakily banded gabbro with a very strongly developed 'gneissic' mineral fabric occurs in several places. Another large xenolith of serpentinized dunite occurs within this section (loc. 24), and yet another is found inland just south of Deer Cove Pond, where the surrounding cumulates are mainly websterite and clinopyroxenite. The layered grey talcose altered ultramafic rocks in Hammer Cove resemble the layers at loc. 41 interpreted as cumulate dunite. The occurrence at Hammer Cove may be within a huge xenolith (or autolith) enclosed within the surrounding, now-carbonated, non-cumulate ultramafics. Just south of the Government Wharf to loc. 10, a tectonic sliver of undeformed websterite and clinopyroxenite occurs in association with slivers of sheared serpentinite and strongly gneissic banded gabbro (well displayed at the road by the Government Wharf, loc. 11). The gneissic fabric is a result of deformation at high, sub-magmatic temperatures; it is not due to the compressive deformation that produced such effects as the regional cleavage and the shearing in the serpentinite. These rocks near the Government Wharf are taken to imply the presence of a tectonically shredded transition zone.

The minimum thickness of the transition zone is estimated as 300m, from the section 1.5 km southeast of Deer Cove.

Gabbros

In other intact ophiolite complexes, variable thicknesses of cumulate gabbros, banded and homogeneous gabbros with a planar mineral fabric, homogeneous undeformed, in places intrusive gabbros and fine-grained intrusive gabbros have been described (Smith 1958; Wilson 1959; Reinhardt 1969; Upadhyay *et al.* 1971; Davies 1971). All except the fine-grained gabbro of Davies (1971) are characterized by a distinctive paucity of opaque minerals and are generally rather leucocratic (plagioclase-rich).

The generally medium-grained (1-2 mm) equigranular gabbros in the Mings Bight Ophiolite Complex display well-preserved textures even though the plagioclase is ubiquitously altered, in places to albite, but more commonly to an ultrafine-grained, turbid, zoisite-clinozoisite aggregate. The rare to subordinate orthopyroxene is almost always altered, whereas the clinopyroxene in places remains fresh. Opaque minerals are conspicuously absent. Analyses of such altered rocks should be used with caution and inferences drawn from them (Norman and Strong 1975) may be suspect. However, the macroscopic physical relationships between the different components of the regionally undeformed gabbros are very well preserved and exposed in the coastal outcrop.

Cumulate and Layered Gabbros

Gabbros showing undeformed cumulate textures, apart from those included within the ultramafic cumulates (locs. 39-41, 23-23 and southeast of Deer Cove), have been positively identified at only two other localities, northeast of Eastern Point and north of Cooking Cove (loc. 30). The latter locality displays spectacular graded layers up to 30 cm thick. Other gradationally contrasting layers of cumulate origin, down to about a centimetre thick, are present, ranging in composition from feldspathic clinopyroxenite to anorthositic gabbro. The zone showing unaltered cumulate layering at this locality is overlain and underlain by layered gabbros containing a strong gneissic mineral aggregate foliation coplanar with the layering. This change, which is gradational over about 1-2 m, is accompanied by a reduction in average thickness and continuity of the layers. This foliation is a result of deformation at high, sub-magmatic temperatures and is not due to the deformation that produced the regional cleavage and thrusting. The exposure described above therefore confirms the suspicions of workers in other ophiolite complexes (Moores and Vine 1971, p. 456; Davies 1971, p. 20) that at least some, if not most, of the layered gabbros with a gneissic mineral fabric, that are by far the most abundant constituent of the layered gabbros in the Mings Bight Ophiolite Complex and are commonly found in other ophiolite complexes, are deformed cumulates.

Most of the layered gabbros exposed in the Mings Bight Ophiolite Complex possess this gneissic mineral foliation, generally strongly developed;

lineation accompanying the foliation is extremely rare and weakly developed when observed. Norman and Strong (1975) do not refer to the prominent gneissic foliation and related deformation. In most places it cannot be directly demonstrated that the layering is cumulate, although as discussed above, this is likely. The gneissic gabbros have layers ranging in composition from clinopyroxenite to anorthosite, although most of them are between mafic and felsic clinopyroxene gabbro. Rarely, tight to isoclinal folds of layered gabbro are seen, as on the west side of Point Rousse (Fig. 4E), and at loc. 42 near Western Point; the gneissic foliation is axial planar to these folds. Other structures in the gabbro are related to differential remobilization of not fully consolidated rocks, particularly clinopyroxenites; some demonstrate the rare presence of some genuinely intrusive gabbro within the layered gabbro.

Gabbroic pegmatites are common in the layered gabbros, commonly forming somewhat poddy dikes or sills mostly between 1-20 cm thick that in all cases cut the layering. One example was seen deformed by the gneissic foliation; all others cut it. The pegmatites range from coarse- to very coarse-grained (several centimetres); most appear to have been pyroxene-bearing, although a few, particularly northeast of Eastern Point, seem to contain original hornblende.

Diabase dikes occur in places in the layered gabbros. Although they are very rare in most places, the section from Point Rousse to Cooking Cove contains dikes occupying up to 5% of the distance measured along the layering and perpendicular to the dikes. The dikes, without exception, cut the layering, gneissic foliation, and the pegmatites. They consist of aphyric, altered, fine-grained dolerite identical to the dikes in the sheeted complex; their margins are chilled but are not as fine-grained as in the sheeted complex. Their presence in part of the layered gabbro, is discussed in a later section.

A maximum exposed original thickness of 300 m for the layered gabbro is taken from the section near Point Rousse. As the contact with both the underlying transition zone and the overlying homogeneous gabbros is always faulted in the coastal sections, this is a minimum estimate.

Homogeneous Upper Gabbro and Transition to Sheeted Dikes

More homogeneous clinopyroxene gabbro is found between the layered gabbros and the sheeted dikes. This is exposed in fault-bounded blocks on the coast north of Big Head for about 200 m north from loc. 22, and south from the south side of Cooking Cove. This gabbro is medium-grained and is generally slightly more feldspathic than gabbro crystallized without differentiation from a basaltic melt. Although there are, in places, slight variations in feldspar content, these mostly take place gradually on a large scale, and sharp contacts cannot be seen. Rarely, as at loc. 22 north of Big Head, a weakly defined planar layering is defined by slight compositional variations, the layers ranging from a few millimetres to not more than a few centimetres thick. These weakly layered patches are isolated within homogeneous gabbro and each layered sequence is no more than a metre or two thick. We suggest that this layering is unlikely to be cumulate in origin. It is essentially parallel to the diabase dikes in the section north of Big Head, and we interpret its original orientation to have been near vertical. The homogeneous gabbro does not show any grain-size variation that might indicate its formation as a series of coarse-grained, wide dikes. No gneissic foliation has been seen in the homogeneous gabbros; a weak grain shape orientation present in rare instances is probably an original igneous crystallization fabric. Small gabbro pegmatite dikes and more irregular patches that cut the homogeneous gabbro are not uncommon.

The transition between this gabbro and 100% sheeted diabase dikes is exposed on the coast between Point Rousse and Devils Cove at loc. 31, and between Cooking Cove and Island Rock Point (locs. 27-29). The base of the latter section, seen in Cooking Cove, is a normal fault that separates layered gabbro with about 5% diabase dikes (to the north) from homogeneous gabbro with 5-10% diabase dikes to the south. Dikes are again typically between 20-60 cm wide. The transition from gabbro to dikes involves an upward increase in the proportion of diabase dikes and a complementary decrease in the proportion of gabbro screens between the dikes. Thus, near Island Rock Point the ratio of dikes to gabbro is at least 99:1. However, in detail, the transition is more complex. In the section between loc. 29 and Island Rock Point, zones containing a high proportion (60-90%) of dikes and typically 20-30 m wide alternate with zones of comparable width that contain a low (10-40%) proportion of the dikes. This pattern is repeated on a larger scale; on the west side of Point Rousse (loc. 31) nearly 100% sheeted dikes are found at least 200 m stratigraphically lower in the gabbro than on the east side where (loc. 28) a significant amount of gabbro is still present. It appears that the zones with a high proportion of dikes have depths of approximately the same dimensions as widths. We interpret the surfaces, defined by any given proportion of dikes to gabbro, to have a corrugated form, and the 100% dike surface to have a minimum relief of 200 m. Although dikes generally become less abundant downwards, we have not been able to trace any dike unambiguously to its roots. On the western side of Point Rousse, just north of loc. 31, a local transition from about 5% to greater than 99% dikes takes place over a distance of about 20m. This transition is exceedingly complex; disconnected pieces of dike cannot be continuously traced to their roots, but the rapidity of the change is noteworthy. We emphasize the striking grain size contrast in all exposures of this zone of transition from homogeneous gabbro to sheeted dikes. Rocks within it are either medium-grained gabbro or relatively fine-grained dolerite in dikes; there are no intermediate categories or gradations from one rock to the other. Another general property of this zone is a sheet jointing parallel to the dike orientation that is also seen in the sheeted dikes and even homogeneous gabbros north of Big Head (loc. 22).

Within the zone of transition from homogeneous gabbro to sheeted dikes, and in the lowest part of the 100% sheeted diabase dike complex, a small amount of medium-grained sodic leucogabbro or trondhjemite (in a broad sense) is found in the form of the matrix to net-vein breccia. This is well displayed within homogeneous gabbro north of Big Head (loc. 22), within homogeneous gabbro with diabase dikes on the south side of Cooking Cove, and within sheeted dikes near loc. 32, west of Point Rousse. On the south side of Cooking Cove, the trondhjemitic breccia cuts some diabase dikes but is cut by others, indicating that it is an integral part of the magmatic sequence. The trondhjemite, essentially quartz-free, consists of albitized, originally-zoned sodic plagioclase and minor actinolitized hornblende. At loc. 22, north of Big Head, there is one occurrence consisting of a few tension-gash-type veins filled with fine-grained quartz-rich trondhjemite, in addition to the more widespread quartz-poor trondhjemite net breccia present in this area.

The minimum overall thickness of the zone of homogeneous gabbro and transition into sheeted dikes is about 300 m in the section from Cooking Cove to Hammer Cove. The base of this section is faulted, but the top is structurally intact.

The thickness of layered and homogeneous gabbro together is more than the 600 m obtained from addition of the separate minimum thickness estimates. The extent of the gabbro south of Western Point and southwest of Devils Cove indicates the presence of at least 850 m of gabbro overall.

Sheeted Diabase Dike Complex

The diabase dikes all consist of homogeneous, fine-grained, wholly altered dolerite; they are aphyric and have an intersertal to slightly sub-ophitic texture. The two best exposures of wholly undeformed 100% sheeted dikes are north of Big Head (locs. 20, 21), where a fault-bounded block contains more than 500 m width of 100% dikes, and on the east side of Devils Cove (locs. 31, 32). The sheeted nature of the diabase in locs. 31, 32 is not at all obvious, and a careful search for chilled margins has to be made before it is apparent that the exposures consist of 100% dikes, some of which are several metres wide. The dikes in the Big Head section (locs. 20, 21) range between 10 cm and 1 m wide; most are between 20 and 50 cm wide, the same as in the homogeneous gabbro and pillow lava elsewhere. The dikes in this section stand out because their chilled margins and internal flow laminations are well defined. Some dikes are half dikes with only one chilled margin, having been split by the intrusion of subsequent dikes. One-hundred percent sheeted diabase dikes are also seen just north of Hammer Cove. Exposures of sheeted dikes are also present between Upper and Lower Green Cove, where they are, in a few places, cleaved and cut by small high strain zones. From Lower Green Cove they become highly cleaved greenschists towards a major high strain zone at Whalesback (loc. 44). Dikes in this section, like those near Devils Cove, are hard to see; there appear to be two distinct orientations, one of which is much more common than the other (loc. 45). Dike widths are similar to those in the Big Head section; they may on average be a small amount wider.

A minimum thickness for the 100% sheeted dike complex is 350 m from the section north of Big Head. Dikes in the Big Head section strike about east-west and dip steeply north whereas those near Point Rousse and Upper Green Cove strike north-south and dip steeply west or east.

Transition from Dikes to Pillow Lavas

This is well exposed only in the slightly deformed to undeformed section in Upper Green Cove. A tiny fault slice at Big Head (loc. 20) of apparently homogeneous, partly vesicular lava (with some interstitial maroon chert suggesting that it is partly pillowed), cut by a few dikes, may belong to this interval. The base of this transition is defined by the first screen of pillow lava seen between dikes. The section in Upper Green Cove shows a stratigraphically upward, generally increasing, proportion of such screens of pillow lava, about 30% pillow lava screens being present at the south side of Upper Green Cove. At loc. 46, in Upper Green Cove, there is a spectacular exposure that displays a dike feeding pillows into a hyaloclastite matrix. A minimum thickness of 850 m of this dike - pillow lava transition is present in the Green Cove section.

Pillow Lavas (Ophiolite Sequence)

Pillow lavas belonging to the ophiolite complex, apart from those in screens between dikes in Upper Green Cove, are only definitely identified in the short section at Big Head (Fig. 2, section EF). These pillow lavas are variolitic, wholly altered, and appear to have been aphyric. They are weakly cleaved, contain very little pillow breccia or hyaloclastite except for the upper 2-3 m, and are cut by rare, thin (~30 cm) diabase sills and dikes. They have an intact, but inverted, stratigraphic contact with sedimentary rocks overlying the ophiolite complex at loc. 19, just south of Big Head, where there is less than 100 m preserved. Pillow lavas (locs. 15-12) south of the sediments are stratigraphically above the sediments and, therefore, despite very similar geochemistry to the ophiolite pillows (R. G. W. Kidd, personal communication, 1900; Norman and Strong 1975), are not part of the ophiolite complex proper. The section of rather strongly cleaved pillow lava and fairly abundant pillow breccia seen from Eastern Point to Deer Cove may be part of the ophiolite complex pillow lavas, particularly since it contains rare diabase dikes in its northern part, but this cannot be certainly established since both contacts are faulted. Some maroon chert in boudinaged beds and interstitial to pillows occurs in the southern half of this section; two boudinaged layers, one up to 1 m thick, of dark, pyritic, manganiferous argillite are found on the north side of Deer Cove (loc. 37). About 700 m thickness of pillow lavas are present in this section. The geochemistry (R. G. W. Kidd, personal communication, 1900) of the Big Head pillow lavas shows that at least some are basaltic komatiites; depending on the definition used, most can be defined as such (A. J. Naldrett, personal communication, 1900).

Rocks Overlying the Ophiolite Complex
Big Head Formation (Sediment Sequence)

We propose the name Big Head Formation for the sequence of sedimentary rocks that directly and conformably overlie the pillow lavas of the Mings Bight Ophiolite Complex. The type section designated consists of the structurally inverted section seen in the shore and cliff exposures south from Big Head on the western side of Mings Bight. The base of the Formation is drawn at the base of the lowest bedded sediment that lies on the pillow lavas and hyaboclastite, which, in the type section, is a purplish-maroon chert bed about 30 cm thick containing slump folds. The top of the formation is drawn at the base of the first pillow lava overlying the sediments, seen about 700m south of Big Head in the type section. The formation consists of green, sandy and silty mafic volcaniclastics, some of which are conglomeratic, and subordinate argillite. A few thin diabase sills are present. The thickness of the formation in the type section is about 300m; the section is cut by a number of small faults of unknown displacement.

Above the purplish-maroon slumped chert bed, there are 30 m of quite well-cleaved, mostly homogeneous, sandy, mafic volcaniclastics. A few, thin, purplish-maroon chert lenses are found in the first metre, and the upper 15 m contain a few intercalations of finely-laminated silty to sandy volcaniclastics; a few laminae are lilac-colored and carbonate-rich. The remainder of the formation is essentially uncleaved as seen in outcrop, apart from the top few metres. The clastic components of the green, arenaceous volcaniclastics consist mainly of abraded, albitized plagioclase grains and lithic clasts, mainly of argillite. Mafic grains, and grains smaller than about 0.5 mm are not seen and have been recrystallized to the fine-grained albite - epidote -colorless actinolite matrix. Beds range up to a few metres thick, the thicker beds usually being coarse-grained. Four repetitive fining-upward sequences are seen in the type section. These start with a conglomeratic bed 1-3 m thick. There follows a sequence of conglomeratic beds, generally thinner and with smaller and less abundant clasts than the first, interbedded with sandy mafic volcaniclastic beds. At some point silty volcaniclastics and, subsequently, argillite become interbedded and, at another point, which may be below or above where the argillite forms individual beds, the conglomeratic beds die out. These sequences start at about 30, 100, 180 and 250m above the base of the formation (L, not marked, M, and N, respectively, on section EF, Fig. 2). Some of the larger conglomeratic beds have channeled bases (e.g. loc. 18). In all the conglomeratic beds, the clasts are supported by the sandy volcaniclastic matrix. The lithic clasts in the first sequence are mainly angular lumps, chips and thin (~0.5 cm) slabs up to 10 cm across of greenish-yellow cherty argillite and a few, small, mafic lava clasts. The argillite is identical to that seen as beds higher in the sequence. The lithic clasts in the next two sequences are dominantly aphyric, non-vesicular, mafic volcanic rock, with subordinate cherty argillite and mafic volcaniclastic blocks, and exceedingly rare gabbro. One of the volcaniclastic blocks measures 0.7 x 2.5 m; most volcanic clasts in the coarsest beds are not more than 30 cm across. The mafic volcanic clasts are moderately rounded suggesting a shallow water transit before deposition in deep water. Only a few sandy and conglomeratic beds are graded.

The uppermost conglomeratic sequence contains different and very distinctive dominant lithic clasts of a mafic volcanic rock studded with large (to 1 cm), black, altered clinopyroxene phenocrysts, which are also abundant as isolated grains in the matrix. Subordinate clast types are aphyric mafic volcanic rock, dolerite and pyroxenite, and rare gabbro. Some of the gabbro clasts contain a gneissic fabric. The distinctive black actinolitized pyroxenes occur sparingly in the associated sandy mafic volcaniclastic beds and a 1 cm thick graded bed consisting entirely of them is also exposed.

A few metres of cleaved sandy to silty mafic volcaniclastic rocks and cherty argillite lie at the top of the formation and these are conformably overlain by cleaved pillow lava that becomes increasingly intensely cleaved to the south. A small number of slumped horizons, small diapiric piercements, washouts of argillite and larger channels show that the sediments of the Big Head Formation are consistently inverted in this section. The few slump folds seen consistently indicate downslope movement from an originally easterly-northeasterly direction. The lack of graded bedding and the matrix-supported conglomeratic beds suggest that most of the sediments were last transported by a grain flow mechanism; a minority are turbidites. The lithic clasts, with the exception of the black pyroxene porphyry, were derived from progressively lower levels of the ophiolite complex. The mafic volcanic rock with black pyroxene phenocrysts and, perhaps, the pyroxenite, were probably derived from an island arc volcano. Mitchell (1970) illustrates very similar rocks from an island arc volcanic - volcaniclastic sequence.

Barry - Cunningham Formation (Upper Sequence of Pillow Lavas)

We propose the name Barry-Cunningham Formation for the sequence of deformed pillow lavas and subordinate pillow breccias and hyaloclastite that overlies the sedimentary Big Head Formation. The type section designated consists of the structurally inverted section on the west side of Mings Bight from the Government Wharf to about 700 m south of Big Head. The base of this formation is drawn at the base of the first pillow lava overlying the sediments of the Big Head Formation. The top of this formation is not preserved due to faulting. This formation mostly consists of moderately to highly deformed rocks, originally mainly green mafic pillow lava, with subordinate pillow breccia and hyaloclastite, and (or) minor sandy mafic volcaniclastic sediments. Some slightly deformed, small sills of diabase and large ones of porphyritic dolerite occur in the type section. The apparent minimum preserved thickness of this formation in its present deformed state is about 1000 m in the type section. Some faults and high-strain zones cut the type section; one high-strain zone may have major displacement across it.

Most rocks in the type section are strongly to intensely deformed, have a single regional penetrative cleavage and a stretching lineation pitching at about 90°. The identification of the original lithologies of the cleaved rocks is difficult; only the recurring presence of deformed varioles defining the forms of congruently deformed pillows confirm that the section was mainly pillow lava. Less-deformed rocks with recognizable variolitic pillows occur in a few places, particularly at the north and south ends of the type section. It is probable that a minor part of the section was sandy to silty mafic volcaniclastics as interbeds, and (or) hyaloclastite as tops to pillow lava flows. An exceedingly strongly-cleaved zone of 100-200 m on either side of the old Barry-Cunningham prospect adit does not contain deformed varioles and may have consisted of mafic volcaniclastics; alternatively it may be a large high-strain zone (syn-cleavage thrust) with the lithologies derived from pillow lava. Uncommon thin diabase sills in the sequence, in contrast, are mostly not cleaved except at their fine-grained margins. Two very large (uncleaved) sills, of plagioclase porphyry dolerite, seen elsewhere as late dikes, occur near the Government Wharf and locally contain glomeroporphyritic-altered plagioclase surrounding cores (1-2 cm across) of altered pyroxene-ilmenite intergrowths. Near its northern contact, the northern sill contains sparse altered black pyroxene phenocrysts instead of plagioclase.

In two localities north of the Barry-Cunningham adit, areas of less-deformed rocks show a southward sequence from pillow lava, pillow breccia to hyaloclastite, and then to a sharp contact with pillow lava (e.g. loc. 15), implying an inverted sequence. Sills in the whole section are always concordant with cleavage, and the cleavage has a general northwest-dipping attitude, although the large, open kink folds have locally changed its strike (Fig. 2 The evidence in the coast section suggests that no major fold closure is present as neither the Big Head Formation sediments, nor the two large porphyritic dolerite sills are repeated. The gradual increase of the cleavage intensity on both margins of this high-strain zone lead us to favor the possibility that it is not a zone of major displacement. If it were, we would expect more abruptly defined margins and, probably, a narrower zone of highly strained rocks showing evidence of disrupted foliations like the high-strain zone at Whalesback.

Late Dikes

Porphyritic dolerite, containing abundant, squat, altered plagioclase phenocrysts forms two large sills high in the Barry-Cunningham Formation and dikes cutting the ophiolite complex, in non-cumulate harzburgite on the east side of Mings Bight (locs. 4, 5), cumulate ultramafics and gabbro (loc. 23 and near Western Point), gabbro (loc. 35 and 100 m north of loc. 22) and sheeted dikes (loc. 32). Two of these large dikes have central parts so replete with phenocrysts that they are easily confused with gabbro. Most have porphyritic centers and phenocryst-poor margins and are wide (~20 m) as distinguished from almost all the non-porphyritic dikes. Two wide (~ 10 m) non-porphyritic dolerite dikes (between locs. 22, 23, and near loc. 30) have the same orientation as the porphyritic dolerite dikes and in outcrop have a fresher appearance than the diabase of the sheeted complex group of dikes. We suggest that they are of the same relative age as the porphyritic dolerite. Small non-porphyritic dolerite dikes cut the porphyritic dolerite 100 m north of loc. 22, and are parallel to its margins. The orientation of the porphyritic dolerite dikes is remarkably consistent considering that they are found in all the thrust sheets, except that of the ultramafics (sheet 1) mostly covered by the waters of Baie Verte. In thrust sheets 2a, 2b, 3 and 4 (Fig. 3) they almost all strike within about 10° of a mean 330° azimuth and bisect the two orientation sets of sheeted diabase dikes. A porphyritic dolerite dike at Western Point is chilled against and cuts at a shallow angle a narrow (~50 cm) zone of phacoidal, rudely foliated fault breccia. We conclude that this fault formed while the ophiolite complex was undisturbed oceanic crust.

The other type of late dike is very rare and narrow (~30 cm) but very continuous. It consists of a purplish-buff, fine-grained dolerite with titanaugite. Dikes of this type have been seen at three localities, cutting sheeted dikes (between locs. 20, 21 and at loc. 32) and cumulate ultramafics (loc. 41). They cut across the sheeted dikes obliquely; the dike between locs. 20, 21 strikes northeast-south-west and is steeply dipping, oriented at a high angle to the porphyritic dolerite dikes. It is not known whether such dikes are part of the general ophiolite complex or belong to a wholly younger event.

Some Detailed Aspects of the Ophiolite Complex

Devils Cove Serpentinite-Ocean Floor Fault

Within thrust sheet 2a (Fig. 2, Fig. 3A), a narrow, subvertical, parallel-sided zone of unfoliated but totally recrystallized, knotty-textured serpentinite runs south from Devils Cove separating gabbro on the west from sheeted dikes on the east (Fig. 2, section CD). This serpentinite is wholly distinct from the undeformed, variably serpentinized ultramafic rocks showing pseudomorphed plutonic textures, and from the highly foliated, usually carbonate-rich, derivatives of ultramafic rocks seen in thrust zones.

On the west side of Devils Cove (loc. 33), the exposure of knotty-textured serpentinite is capped by a few metres of a clastic ultramafic rock, which contains pebbles and boulders up to 50 cm across, all consisting of the same serpentinite as below, floating in a crudely cleaved, granular serpentine matrix. The cleavage dips steeply to the north and strikes about east-west, perpendicular to the elongation and contacts of the serpentinite zone but essentially concordant with the regional cleavage attitude as seen in the pillow lava south of Eastern Point. The clastic rock therefore formed before regional deformation and inversion of this thrust sheet. The clasts are slightly deformed and the cleavage dies out abruptly at the sharp contact with the homogeneous, knotty serpentinite below. In detail, the contact is irregular; overall it dips about 25° southward on the eastern, north-south, steep face of the outcrop, and has a deep U-shape on the east-west, vertical south face of the outcrop. Inverted cumulate banding on Point Rousse (and all other younging indicators on the Mings Bight Peninsula) shows that all the rocks here should be rotated through 110° to 140° about strike to obtain a first approximation to the original horizontal. When this rotation is performed, the attitude and shape of the ultramafic clastic rock describes a very steeply plunging half-cylinder. In view of the restored attitude and shape, the monomictic character and the structural environment of the clastic body, we suggest that it is an exposure of part of a gas breccia (or tuffisite) pipe.

The subvertical contact of the serpentinite and the gabbro to the west is also exposed here (loc. 33). The gabbro is rodingitized and weakly brecciated within a metre of the contact; small apophyses of serpentinite behave as cold, tectonic 'intrusions' into cracks in the rodingitized gabbro. Within the serpentinite, in the zone up to 3 m from the contact, there are five or six, 20-50 cm wide, wholly undeformed sheeted diabase dikes with chilled margins, striking parallel with the contact of serpentinite and gabbro. A narrow zone of weakly sheared and brecciated serpentinite occurs between the westernmost dike and the gabbro, and a thin septum of serpentinite is found between two of the dikes. The dikes are chilled against the serpentinite, and were intruded after the shearing within the serpentinite. They are identical to sheeted diabase dikes seen elsewhere in the ophiolite complex, and have the same orientation as those on the east side of Devils Cove.

These relationships show that this zone of serpentinite does not occupy a thrust fault, as asserted by Norman and Strong (1975), and are unambiguous evidence that the zone is a cold, diapiric serpentinite 'intrusion' that was emplaced on a fault between sheeted dikes and gabbro while basaltic magma was still available to form the undeformed diabase dikes that cut the serpentinite. We suggest that the gas breccia pipe within the serpentinite was also formed at this time. We interpret this feature as a small ocean-floor fault (probably a normal fault), containing a cold, diapiric serpentinite 'intrusion', formed very close to the spreading-ridge axis that produced the ophiolite complex. The same structural setting has been interpreted for some serpentinites dredged from the Mid-Atlantic Ridge near 45°N (Aumento *et al.* 1971).

Localities Previously Described as Unconformities

Kennedy and Phillips (1971) claimed that there were two exposures of an unconformity of sediments overlying ultramafic rocks on the coast of the Mings Bight Peninsula. One of these is the exposure on the west side of Devils Cove (loc. 33) described above that we interpret as a gas breccia pipe in a cold serpentinite intrusion on an ocean-floor fault. The other locality is Hammer Cove (loc. 26) where the rocks appear to have been originally fairly complex and to have undergone mineralogical alteration and to be cut by closely spaced small faults associated with the major thrust fault exposed at the base of the cliff on the north side of the cove. The rocks alleged to be 'serpentine sediments' by Kennedy and Phillips (1971), which consist of a few metres of grey, talc-rich rocks that are laminated or layered in places, we interpret, by comparison with a few similar layers seen at loc. 41, west of Deer Cove, to have been olivine-rich dunite cumulates. Specularite-rich layers in the Hammer Cove rocks contain relict chromite in thin section, suggesting that they were cumulate chromite-rich layers. One layer, possibly originally feldspathic dunite, shows an apparent graded bed, facing downwards. Rusty-weathering, altered, but homogeneous ultramafic rocks adjoin the grey, altered ultramafic cumulates and continue to the south around Red Point. They are wholly altered to Mg-Fe carbonates but are not penetratively deformed. The alleged conglomerate in Hammer Cove is found in some places at or very near the contact of the rusty, altered, homogeneous ultramafic rock and the grey, altered cumulates. It seems to have been formed by a mix of igneous and (or) tectonic brecciation. In one place, the contact between the rusty homogeneous rock and the grey altered cumulates appears not to be faulted. It is not clear whether this is an intrusive or a cumulate contact. If it is cumulate, the homogeneous altered ultramafic rocks to the south would have originally overlain the grey, altered cumulates before overturning and transport of the thrust sheet they lie in, and therefore the alleged sediments cannot have overlain those ultramafics, unconformably or otherwise. The ultramafic igneous breccias exposed in the altered ultramafic rocks on the south side of Red Point and in Hammer Cove, suggest that the altered cumulate rocks in Hammer Cove are more likely to be a very large xenolith within the altered, homogeneous ultramafic rock.

Some Detailed Features of the Ophiolite Gabbros and Cumulate Ultramafics

Cumulates

Noticeable layering in the undeformed cumulates is usually less than a few metres thick and ranges down to laminae a single grain thick. Accumulation fabrics, such as tabular plagioclase aligned in the layering plane, are rare. The rocks most obviously showing post-cumulus crystal growth and resorbtion are the harzburgites with large orthopyroxene oikocrysts. Mineral (density) and size grading are seen in a few places. Scour structures and cross-bedding are very rare; a spectacular example of cross-bedding in a 1 m thick cumulate feldspathic dunite layer is exposed about 100 m north of loc. 24.

High Temperature Deformation of the Ophiolitic Plutonic Rocks

The non-cumulate harzburgites are typical of those found in other ophiolite complexes and alpine ultramafic bodies dismembered from them (Loney et al. 1971). Their xenomorphic granular textures, the kinked enstatites and the preferred crystallographic orientation of olivine suggest that they have undergone extensive recrystallization and grain growth at low applied stresses and strain rates. The typical slightly ovoid orthopyroxene grains in most samples show moderate to extreme kinking whereas the olivines show undulose extinction; weakly kinked olivines are rarely seen. Local zones of higher strain, which contain a more obvious foliation, are found in a few places.

In other intact ophiolite complexes (e.g. Bay of Islands), the fabrics due to this deformation affecting the harzburgites continue upward, affecting at least parts of the cumulate ultramafics and gabbros, although it is not seen in the upper levels of the gabbros. Thus, we suggest that the gneissic foliation that affects some zones in the cumulate ultramafites and cumulate gabbros of the Mings Bight Ophiolite Complex, is due to the same deformation that pervasively affects the harzburgites, even though a continuous sequence from harzburgite to cumulates is not seen in the Mings Bight Complex.

Some zones in the cumulates are slightly deformed (e.g. locs. 42-39) whereas others are, in parts, strongly deformed (e.g. locs. 24, 25). This suggests that the deformation was episodic; variation in the rate of consolidation of the different rock types by intercumulus crystallization, if significant, would also have affected the susceptibility of the rocks to deformation. We infer that the deformation was localized in cumulate layers that were not fully consolidated. At loc. 40, west of Deer Cove, a dike about 40 cm wide of gabbro with anorthosite layers (vertical in Fig. 4F) cuts cumulate websterite and clinopyroxenite affected locally by a weak mineral foliation parallel to layering (horizontal in Fig. 4). The gabbro in the dike is also affected by a weak gneissic mineral foliation parallel to the layers in the dike and its margins. This dike appears to have formed by cracking of the pyroxene-rich cumulates and injection of originally underlying layered cumulate gabbro and anorthosite that was not fully consolidated, but sufficiently so to preserve the layering. Hopson (1975) reports similar features from the Point Sal Ophiolite Complex. In the Mings Bight Complex, the weak gneissic mineral foliation in the dike is due to flow of nearly consolidated cumulate material. We infer a similar control on the foliation seen elsewhere in the complex. We suggest that it developed progressively, soon after deposition of any affected rocks as cumulates.

The gneissic foliation, which forms by the deformation and progressive breakdown of original cumulate grains, is almost everywhere parallel to layering. In rare examples, it crosscuts layering at a low angle. North of loc. 30, an outcrop surface only slightly oblique to the layering (Fig. 4A) exposes a contact between more mafic (top) and more felsic gabbro. Here, a diabase dike crosscuts both the layering and the foliation implying the origin of foliation during formation of the ophiolite complex. Another example is found on the southwest part of the exposures at Point Rousse, where the foliation is the axial surface fabric to a tight fold in layered gabbro with thin anorthosite layers (Fig. 4E). At loc. 30, the layered rocks with the mineral foliation are at least partly derived from cumulates; undeformed cumulate layers are overlain and underlain by foliated layered rocks. Pinch-and-swell type boudinage of layers within foliated rocks is seen south of loc. 25; at the south side of the exposures at loc. 30 (Fig. 4C) clinopyroxenite masses within gabbro may be boudinaged layers, or perhaps autoliths.

Unambiguous slump features are rare. Figure 4D illustrates one example in mafic to felsic gabbro that we interpret to be due to surficial breakup of cumulate layers. Some blocks are fractured; others are deformed in a ductile fashion and have a weak mineral foliation. Southwest of loc. 30, another smaller example consists entirely of small pieces of anorthositic gabbro broken in a ductile fashion; both fragments and matrix are deformed and affected by the mineral foliation.

The mineral foliation seems to us to be analogous in nature to the flow-induced fabrics of viscous magma 'mush' known in granite batholiths (Brindley 1954) and the more strongly developed fabrics analogous to those seen in the marginal zones of high strain that occur at the margins of such batholiths.

Intrusive Features in Cumulates

At Point Rousse, an angular xenolith of anorthositic gabbro with gneissic foliation is enclosed by undeformed, medium-grained gabbro (Fig. 4B). In the southwest part of the exposures at Point Rousse, more of this gabbro crosscuts the folded and mineral-foliated layered gabbro (Fig. 4E). The post-foliation gabbro does not seem to form regular dikes or sills, and it does not seem to be very voluminous. Other xenolithic relationships are well displayed about 100 m north of loc. 24, where large (5 m) blocks of layered cumulate ultramafics are enclosed in coarse pegmatitic gabbro; about 30m southwest, rounded, plastically disrupted gabbro xenoliths up to 1 m across are enclosed in the same coarse gabbro. Coarse websterite at this locality also appears to be intrusive. Large (10-20 m), isolated dunite xenoliths are found in three separate localities in the ultramafic cumulate zone. All xenoliths consist of rock types found in the layered, originally cumulate sequence and no xenoliths containing diabase dikes have been seen.

Relatively thick (~1-2 m) dark clinopyroxenite layers appear to have been mobile until very late in the history of the complex. At Point Rousse, these

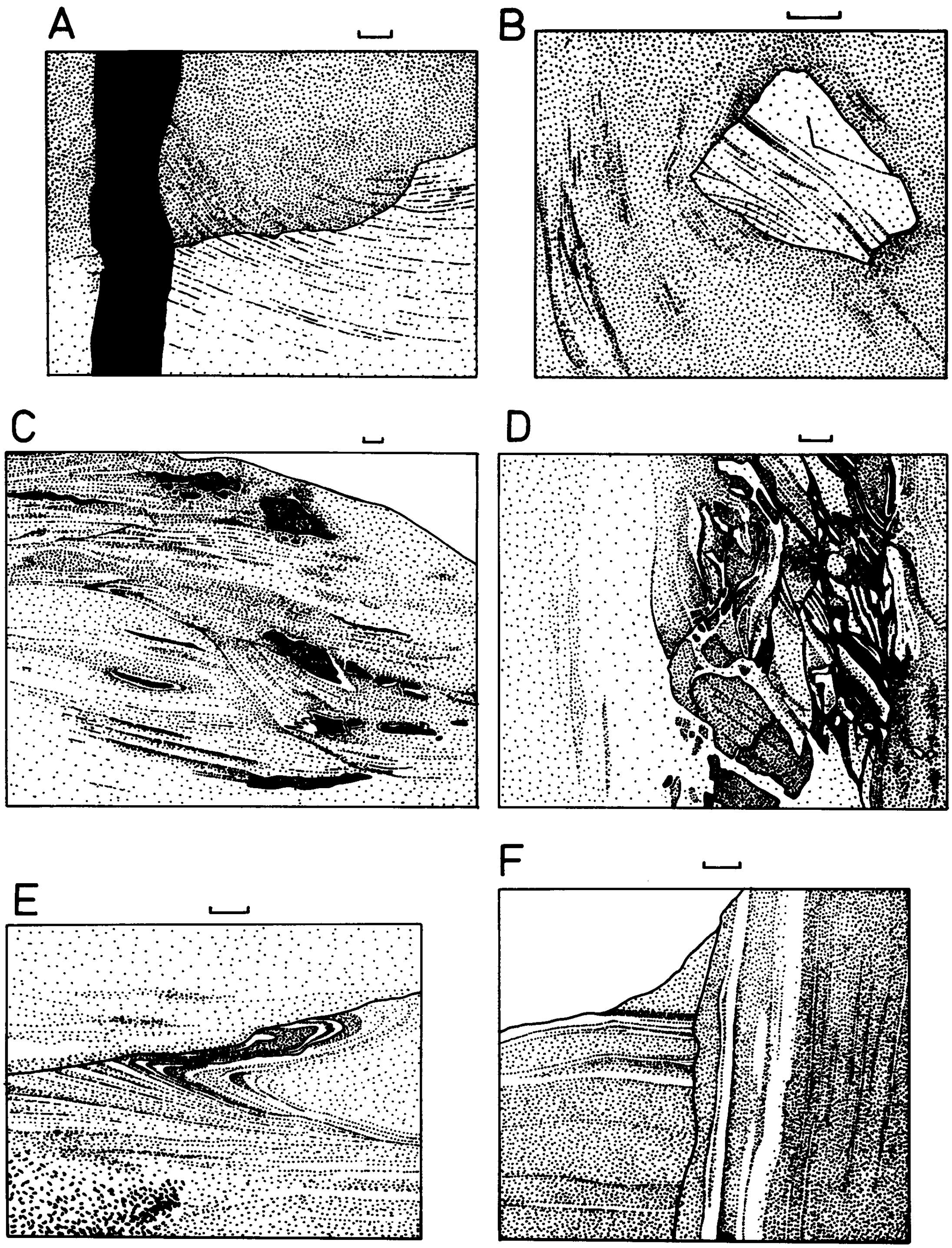

FIG. 4. Structures and relationships in the gabbros and cumulate ultramafics. (A) North of locality 30; (B) Point Rousse; (C) locality 30; (D) between localities 24 and 25; (E) southwest of Point Rousse; (F) locality 40. All sketches except D are oriented their present way up. Present way up is to the left in D. Original way up is the opposite to present way up in all cases. Black (in A), diabase dike; very heavy stipple (in C), clinopyroxenite; heavy stipple, mafic gabbro; medium stipple, gabbro; light stipple, anorthositic gabbro; coarse stipple (in E), pegmatitic gabbro. Rocks in left half of F are clinopyroxenites and websterites. Discussion in text.

layers have been remobilized and have intruded into the surrounding layered rocks. At least one has been remobilized after it was cut by two diabase dikes occurring as disrupted segments in the clinopyroxenite.

The diabase dikes that cut cumulate and deformed layered gabbros are most abundant near Point Rousse and between Big Head and Caribou Head. They never exceed 5% of the width of the section in the layered gabbros and always cut the layering at a high angle. Figure 5A shows a diabase dike that cuts thinly layered gabbros with gneissic foliation. The foliation is cut by a narrow pegmatite dike that is itself cut by the diabase dike. On the eastern side of Point Rousse at (Fig. 5C) and 50 m north (Fig. 5B) of loc. 30, diabase dikes cutting gabbro (layered in the latter case) display partially detached flaps of host rock that have been bent into, and across (Fig. 5C) the diabase dikes. Diabase dikes, near loc. 25 (Fig. 5D) show clear evidence of the fracturing of the gabbro that allowed their intrusion. The sections of layered gabbro west of Western Point, and west of Devils Cove, contain very few diabase dikes. The sections of ultramafic cumulates contain even fewer; the one seen in the section west of Deer Cove is believed to be a late alkalic dike. The late porphyritic dolerite dikes are known to cut the full thickness of the complex, and to be associated with non-porphyritic dikes. Therefore, it cannot be established whether the diabase dikes cutting the layered and cumulate gabbros are of the same age as the porphyritic dolerite dikes, the sheeted complex dikes, or an intermediate age. In the two sections where they are most abundant (Point Rousse and south of Caribou Head), they have the same orientation and appearance as the nearby sheeted dikes; if, as this suggests, they are of the same age as the sheeted complex dikes, their apparent increasing scarcity downwards from the cumulate and layered gabbros into the ultramafic cumulates implies that they are either fed from sills or that the apparent increasing scarcity is fortuitous. The dikes have chilled margins and consist of fine-grained dolerite but no sills with fine-grained dolerite margins have been seen in the layered rocks and it is concluded that the apparent increasing scarcity is fortuitous.

Orientation of Dikes

The sheeted dikes in thrust sheet 2a (Fig. 3), from Point Rousse to Hammer Cove, and in sheet 2b in Upper Green Cove, strike north-south and are subvertical (Fig. 2). Those in thrust sheet 3 from Caribou Head to Big Head strike east-west and dip steeply north, orthogonal to those in thrust sheets 2a and 2b. The consistent orientation of the thick, late, porphyritic dolerite dikes in thrust sheets 2a, 2b, 3 and 4 indicates that there has not been significant rotation about a vertical axis of the thrust sheets relative to one another, particularly since these dikes do not follow the older fracture sets parallel to the sheeted dikes. The two orthogonal sheeted dike orientations, combined with the porphyritic dolerite dike orientation that bisects them, very closely confine the possible orientation of a rotation axis that will remove the inversion of the thrust sheets and restore the cumulate layers and bedding to the original horizontal, while leaving all three dike sets near vertical. It is perhaps not a coincidence that this axis is essentially horizontal and parallel to the strike of cumulate layering near Deer Cove and Point Rousse, and to the strike of bedding near Big Head. This rotation, when performed, leaves the porphyritic dolerite dikes striking about northwest-southeast; the present east-west set of sheeted dikes become north-south striking and conversely the presently north-south striking set become east-west striking. The two orthogonal sets of sheeted dikes thus restored are most easily interpreted as being from two episodes of spreading with a radical change in spreading direction from the earlier to the later episode. We have not seen any evidence that shows one set to be older than the other. Alternatively, the orthogonality of diabase dikes of the sheeted suite cutting layered and cumulate gabbros and the layering suggests that the Point Rousse and Caribou Head sections have been tectonically rotated relative to one another, either during the thrusting or by ocean-floor faulting. Therefore, there may have been only one original orientation set of sheeted dikes; taking the set in thrust sheet 3 north of Big Head as anomalously rotated, because two other thrust sheets (2a, 2b) both show the same orientation, then the original orientation of the sheeted dikes before regional deformation would have been east-west, at a high angle to the length of the Baie Verte Lineament (Fig. 1).

Anomalies in the Ophiolite Complex

The Mings Bight Ophiolite Complex is, in its large-scale lithologic zonation and relationships, apparently very similar to other well-preserved ophiolite complexes. However, our observations on other Newfoundland ophiolite complexes, together with generalized descriptions of well-preserved ophiolite complexes elsewhere (Wilson 1959; Reinhardt 1969; Moores 1969; Davies 1971) indicate that two features of the Mings Bight Ophiolite Complex are perhaps anomalous. One is

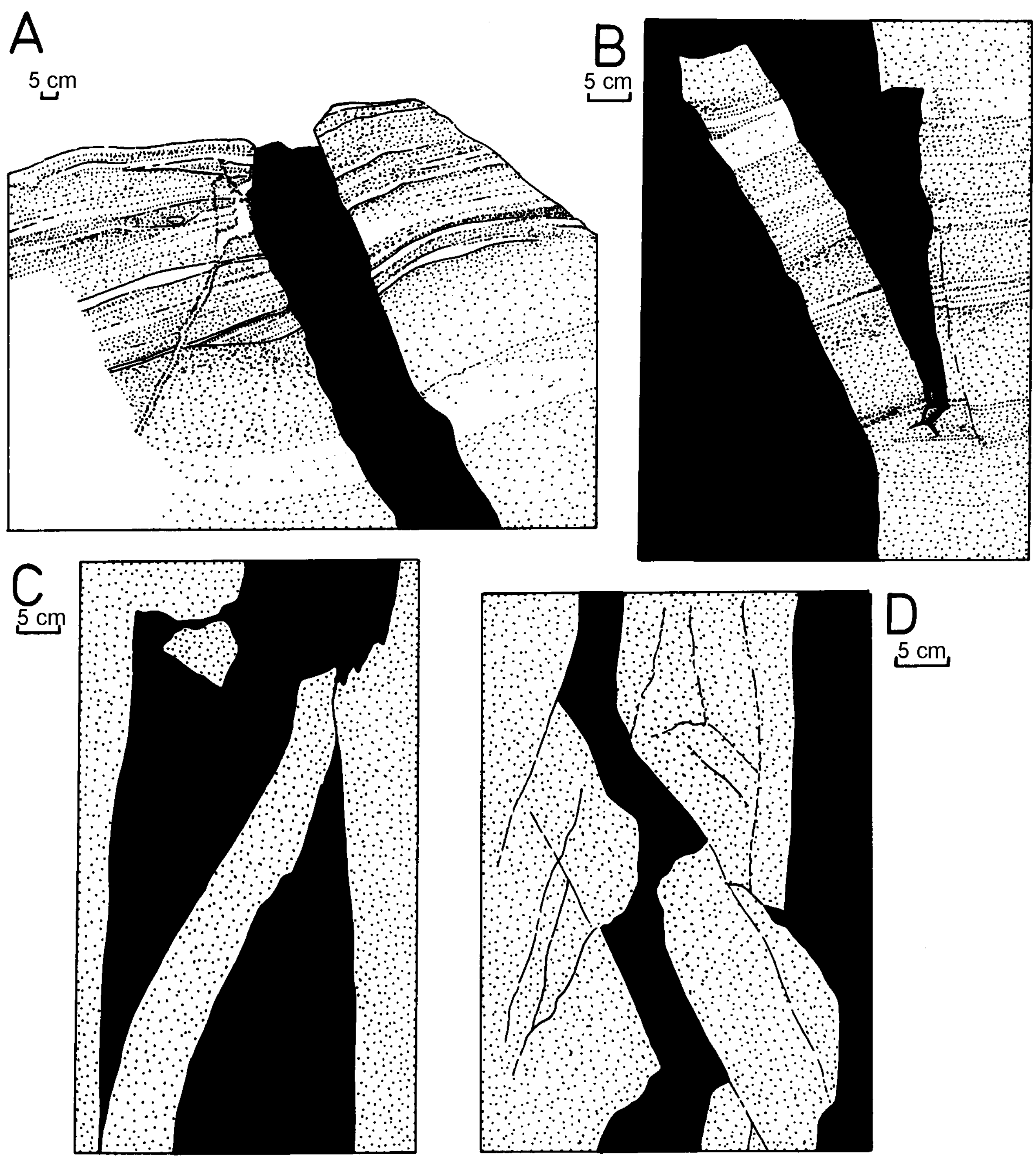

FIG. 5. Relationships between gabbros and dikes. (A) Point Rousse; (B) north of loc. 30; (C) loc. 30; (D) between loc. 24 and 25. All sketches oriented their present way up. Original way up is opposite to present in all sketches. Solid black, diabase dike; heavy-medium-light stipple, mafic-normal-anorthositic gabbro. P, gabbro pegmatite.

the occurrence of up to 5% width of diabase dikes, that seem to be of the sheeted suite, cutting layered and cumulate gabbro. The other is the fairly common occurrence of xenolithic relations in the layered gabbros and cumulate ultramafics, and, in one place (Red Point) in undeformed but completely altered homogeneous ultramafic rock.

The geology of the layered and cumulate zone appears to have been rather complex even before pervasive mineralogical alteration and disruption

TABLE 1. Minimum 'stratigraphic' thicknesses of the Mings Bight Ophiolite Complex and overlying sediments and volcanics

	Lithologic unit – formation	Minimum thickness (m)	Location of section
	Barry–Cunningham Formation (upper pillow lavas)	1000	South of Big Head
	Big Head Formation (volcaniclastic sediments)	300	South of Big Head
← Mings Bight Ophiolite Complex →	Pillow lavas	100	Big Head
	Pillow lavas(?)	(700)	South of Eastern Point
	Pillow lavas + dikes	850	Upper Green Cove
	100% sheeted diabase dikes	350	North of Big Head
	Homogeneous gabbro + dikes	300	Cooking Cove – Hammer Cove
	Overall gabbro	850	Western Point to south (inland)
	Cumulate gabbros	300	Point Rousse
	Cumulate ultramafics	300	Southeast of Deer Cove (inland)
	Non-cumulate harzburgite	1800	Tin Pot Island – Grassy Island

NOTE: Bracket indicates conformable sequence. Contacts of other units are mostly tectonic (see map, Fig. 2) for these sections, but intact contacts between all units except those between non-cumulative harzburgite and cumulate ultramafics, and between cumulate and homogeneous gabbros, are seen in one or more places. Spaces indicate places where tectonic excision is present.

by faulting. Because of this disruption we are unable to identify the cause(s) of the anomalous features. It has been suggested (Kidd 1974, 1977) that the rocks of the Baie Verte Group, to the west of the Mings Peninsula around Advocate Mine and south along the Baie Verte Lineament, contain evidence of an oceanic transform fault - fracture zone history, in particular in the occurrence of a basal fault-scarp sedimentary megabreccia and in the nature of the deformation of the gabbroic lithologies comprising all the blocks in it. If this inference is correct, then the apparently anomalous features can be explained by the formation of the ophiolite complex exposed around Mings Bight near (within a few kilometres) an oceanic transform fault - fracture zone. Some of the possible complexities resulting from this proximity are discussed by Karson and Dewey (1978).

Summary and Conclusions

Ophiolite Complex

Although the Mings Bight Ophiolite Complex is dissected by thrusts and steep faults, the original sequence of lithologic units can still be established, and the contact relationships of all the units, except those between non-cumulate and cumulate ultramafics and between layered and homogeneous gabbro, are exposed in the continuous coastal section. The reconstructed, minimum stratigraphic thicknesses are given in Table 1.

The section consists, at the base, of 1800 m of non-cumulate, residual, tectonite harzburgite containing minor dunite layers. This must have formed a floor to the overlying cumulate rocks, and must be a residue left by the extraction of a mafic partial melt that now constitutes the gabbros and basalts. Layered cumulate ultramafic rocks, consisting mainly of clinopyroxenite and websterite, with subordinate gabbro, dunite and harzburgite succeed the non-cumulate harzburgites. These grade up into layered, originally cumulate gabbros, which are affected in many places by a gneissic mineral foliation. Homogeneous, somewhat leucocratic gabbro overlies the layered, foliated gabbros; it is locally cut by net-vein breccias of trondhjemite that are also uncommonly found in the lowest part of the succeeding 100% sheeted diabase dike complex. Parallel diabase dikes cut the homogeneous gabbro and some cut the trondhjemite; the dikes increase in abundance upward at the expense of the gabbro until the rock consists entirely of parallel diabase dikes. Pillow lava screens occur above at least 350 m of sheeted dikes. These increase in abundance upward at the expense of the dikes, so that, at the top of the ophiolite complex, the rock consists entirely of pillow lava. The minimum reconstructed thickness of the complex from the base of the ultramafic cumulates to the top of the pillow lava is about 2500 m. A 1300 m thick section of coarse mafic volcaniclastic sediments and pillow lavas conformably overlies the ophiolite complex; this was probably at least 2.5 km thick before deformation, and an original stratigraphic top to the section is not seen.

We emphasize the occurrence of the high-temperature mineral foliation in parts of the layered, originally cumulate gabbros and, to an apparently lesser extent, in the ultramafic cumulates.

The foliation is not ubiquitous and is zonal; there are rapid variations in its intensity across the layering to which it is almost everywhere parallel. This contrasts with the Bay of Islands Complex, where (personal observations) a strong lineation dominates in the zonal high-temperature deformation of the cumulate ultramafics and lower layered gabbros. The deformation is accompanied, in the harzburgites of the Mings Bight Complex, by a preferred orientation of the medium- to coarse-grained olivine, which has undergone recrystallization and grain growth, implying low applied stresses. The cutting of foliated layered gabbros by unfoliated gabbro, by a few diabase dikes of the sheeted suite and by the late porphyritic dolerite dikes, which also cut the foliation in the harzburgite, shows that the mineral foliation was formed during the generation of the ophiolite complex. An explanation of the origin of the foliation has been suggested by Dewey and Kidd (1977). We suggest that the mostly moderate foliation developed in the Mings Bight Complex as opposed to the strong lineation in the Bay of Islands Complex resulted from different spreading rates, higher in the case of the Bay of Islands. It may be possible to calibrate this proposed spreading rate indicator by work on Mesozoic and younger ophiolites and by detailed seismic studies of present spreading ridges.

The occurrence of diabase dikes of the sheeted suite, and fairly common xenolithic relationships in the layered, originally cumulate ultramafics and gabbros, are apparently anomalous. We suggest that they may indicate that the ophiolite complex was formed near (but not in) a transform fault-fracture zone in the developing oceanic lithosphere. The late porphyritic dolerite dikes, which cut the whole ophiolite complex, may be contemporary with the upper pillow lavas (Barry-Cunningham Formation) because they feed thick sills within the latter; they clearly pre-date the regional deformation. Karson and Dewey (1978) show that plagioclase porphyry dikes are characteristically associated with a transform fault zone preserved in the Lewis Hills massif of the Bay of Islands Complex. It is possible that this is also the case for the dikes in the Mings Bight Complex.

The relationships between sheeted dikes and homogeneous gabbro, and sheeted dikes and pillow lavas, show how these parts of the complex were built. That dikes become less abundant downwards into the homogeneous gabbro requires that gabbro be progressively plated onto a roof that, at the exact spreading axis, is the base of the 100% sheeted dike layer. The lower levels of the plated gabbro, formed further from the exact spreading axis, but within the zone of dike injection, will then be cut by fewer dikes than at higher levels (Dewey and Kidd 1977). The fact that dikes cut gabbro, not the reverse, and that no xenoliths of sheeted dikes or of gabbro containing diabase dike segments are seen in the layered gabbros supports the idea that gabbro plates progressively onto the roof and also shows that blocks of the roof do not seem to be susceptible to falling into the magma chamber. The zones of high dike density that penetrate further than normal down into the homogeneous gabbro are interpreted as temporarily preferred sites of dike injection. As a result of these preferred injection sites, the form of the base of the 100% dike unit is not planar but has an original relief of up to 250 m. The rare, diffuse banding seen in the homogeneous gabbro cut by dikes is interpreted as a roof plating-banding. Small areas of trondhjemite net-vein breccia that cut the homogeneous gabbro and some diabase dikes, but are cut by other diabase dikes, show that it formed in the zone of dike injection during the generation of the oceanic crust. These small volumes of trondhjemite are probably derived from residual interstitial liquid formed in the crystallization of the homogeneous roof gabbros. The contrast in grain size between the dikes and gabbro shows that the gabbro must be very rapidly cooled after plating; it has been suggested (Dewey and Kidd 1977) that hydrothermal circulation in the dike and pillow lava roof is responsible. The gradual upward increase of pillow lava screens between dikes shows that the pillow lavas also were progressively accumulated in the zone of dike injection. The lowest lavas would have been formed at the exact spreading axis immediately above the top of the 100% sheeted dike complex; younger lavas extruded away from the exact spreading axis will be cut by fewer dikes (Dewey and Kidd 1977).

Although the sheeted dikes are readily seen in the Big Head section, they are very difficult to detect in Green Cove and near Devils Cove, even though these are clean, fully exposed coastal outcrop. This may be of significance given reports that some ophiolite complexes contain large areas of homogeneous diabase but no sheeted dike complex (e.g. Davies 1971). If, in any ophiolite complex, diabase dikes are seen in pillow lava or in gabbro, we suggest that it is likely that 100% sheeted dike complex was once present, and that apparently homogeneous diabase is very likely to consist of sheeted dikes that are difficult to discern.

Tectonic Setting

The thick volcaniclastic sediment and pillow lava sequence that lies conformably on the ophiolite complex is difficult to reconcile with an origin at an oceanic accreting plate boundary like those of the present-day spreading ridges. Thin pelagic sediment sequences cover present-day oceanic crust, as they do some well-preserved ophiolite complexes (Moores and Vine 1971; Reinhardt 1969; Moores 1969). The Betts Cove Complex (Upadhyay *et al.* 1971), however, does have a thick, conformably overlying volcaniclastic sediment and intercalated volcanic sequence very similar to that overlying the Mings Bight Complex. As with the Betts Cove Complex, we interpret this to mean that the Mings Bight Ophiolite Complex was generated as part of the floor to a marginal basin just to the rear, or within, a volcanic island arc situated above a major subduction zone (Bird *et al.* 1971; Dewey and Bird 1971; Kidd 1974, 1977). The sedimentary structures in the coarse volcaniclastic sediments of the Big Head Formation indicate that transport was dominantly by a grain-flow mechanism, and that the section represents a deep water, base-of-slope environment. One of the clast types strongly suggests the presence of a nearby island arc, which would have been to the east because slump structures show an original east to west palaeoslope. The coarseness of the sediments and the small proportion of argillaceous sediment show, given present-day slopes and sedimentation patterns at the island-arc sides of marginal basins (e.g. Karig 1971), that this section is unlikely to have been more than about 30 km from the edge of the arc platform.

Because the Baie Verte Group in the Mings Bight area (and elsewhere) has an overall synclinal structure, and because there is no evidence in the form of tectonic melanges, it does not appear that subduction was involved in the Acadian deformation of this belt. Thus, the original width of the Baie Verte Lineament can be estimated (Kidd 1977) by attempting to unstack the thrusts and the major synclinal fold. Allowing a generous estimate for the amplitudes of the fold and the thrust displacements, this gives an original width of oceanic crust of not more than 30-50 km. This estimated width and the apparent length (100-200 km) of the basin suggests (Kidd 1977) that it was most comparable in type and scale to the narrow discontinuous basins that dissect the New Hebrides island arc (Karig and Mammerickx 1972).

Regional Geology

Relationships exposed on the east side and south of Mings Bight show clearly that the regional deformation and lowest greenschist facies alteration and metamorphism of the Baie Verte Group wholly post-dates that of the polyphase-deformed and metamorphosed Fleur de Lys schists. The overall structure of the Baie Verte Group in the Mings Bight area is a tight, moderately northwest-plunging syncline disrupted by similarly dipping thrusts that moved towards the southeast-eastsoutheast. A late high-strain zone just offshore from the northwest coast of the Mings Peninsula marks the eastern edge of the oversteepened, subvertical belt that contains most of the rocks in the remainder of the Baie Verte Lineament to the west and south. Evidence seen south of Flatwater Pond shows (Kidd 1974, 1977) that the regional deformation, lowest greenschist facies metamorphism and the thrust-disrupted synclinal form of the rocks in the lineament involve early Devonian rocks that overlie the Baie Verte Group with very slight (a few degrees) angular unconformity. The deformation is therefore probably Acadian. The late, steep faults that consistently downthrow to the north and west in the Mings Bight area can be matched south of Flatwater Pond with late structures showing the same kinematics, which are opposite to those of the main deformation and thrusting. These structures indicate a relaxation of the convergent motion and a 'fall-back' of the rocks into the steeply-dipping pinched zone from which they had just been expelled.

The evidence cited above that the Baie Verte Group was almost flat-lying, and was not at all penetratively deformed in the early Devonian suggests that it is very unlikely to be regionally allochthonous. This suggestion is confirmed by the presence, south of Flatwater Pond, of conglomerate clasts in sediments low in the Baie Verte Group stratigraphy (which also contain slumps showing an east-to-west palaeoslope) that can be matched with rocks now exposed in the terrain to the east. Some of these quartzo-feldspathic clasts contain strong pre-depositional foliations (Kidd 1974, 1977) and therefore show that the Fleur de Lys deformation predated development of the Baie Verte Lineament as a marginal basin.

Williams (1977) argues that the Mings Bight Ophiolite Complex and the more disrupted ophiolite complex rocks along the western side of the Baie Verte Lineament were obducted at the same time as the allochthonous Bay of Islands Complex, and that this obduction was responsible for the deformation and metamorphism of the Fleur de Lys Supergroup. He avoids the evidence of the predepositionally deformed clasts in the Baie Verte Group sediments and the evidence that the Baie Verte Group was flat-lying long after obduction by proposing that the rocks of the Baie Verte Group around and south of Flatwater Pond are a younger, separate group from those near Mings Bight and Advocate Mine at the northern end of the Lineament, and by proposing abandonment of the term Baie Verte Group. However, a rather precise stratigraphic correlation can be made between the sediments and volcanics of the Baie Verte Group south of Flatwater Pond and (data from J. T. Bursnall, personal communication, 1900) the section overlying the Advocate Mine ophiolite (location on Fig. 2). Furthermore, the sediments south of Flatwater Pond are also essentially identical in lithology and are affected by the same, single Acadian cleavage as these rocks at Advocate and those that *conformably* overlie the Mings Bight Ophiolite Complex.

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