Forearc extension and sea-floor spreading in the Thetford Mines Ophiolite Complex

JEAN-MICHEL SCHROETTER[1], PHILIPPE PAGÉ[1], JEAN H. BÉDARD[2], ALAIN TREMBLAY[3] & VALÉRIE BÉCU[2]

[1]*Institut National de la Recherche Scientifique—Eau, Terre et Environnement, 880 Chemin Sainte-Foy, Québec, PQ, Canada, G1S 2L2*

[2]*Natural Resources Canada, Geological Survey, 880 Chemin Sainte-Foy, Québec, PQ, Canada, G1S 2L2 (e-mail: jbedard@nrcan.gc.ca)*

[3]*Département des Sciences de la Terre, Université du Québec à Montréal, CP 8888, Succursale centre ville, Montréal, PQ, Canada, H3C 3P8*

Abstract: The Ordovician Thetford Mines Ophiolite Complex (TMOC) is an oceanic terrane accreted to the Laurentian margin during the Taconic Orogeny and is affected by syn-obduction (syn-emplacement) deformation and two post-obduction events (Silurian backthrusting and normal faulting, and Acadian folding and reverse faulting). The southern part of the TMOC was tilted to the vertical during post-obduction deformation and preserves a nearly complete cross-section through the crust. From base to top we distinguish cumulate Dunitic, Pyroxenitic and Gabbroic Zones, a hypabyssal unit (either sheeted dykes or a subvolcanic breccia facies), and an ophiolitic extrusive–sedimentary sequence, upon which were deposited sedimentary rocks constituting the base of a piggy-back basin. Our mapping has revealed the presence of subvertically dipping, north–south- to 20°-striking faults, spaced *c.* 1 km apart on average. The faults are manifested as sheared or mylonitic dunites and synmagmatic breccias, and correspond to breaks in lithology. The fault breccias are cut by undeformed websteritic to peridotitic intrusions, demonstrating the pre- to synmagmatic nature of the faulting. Assuming that rhythmic cumulate bedding was originally palaeo-horizontal, kinematic analysis indicates that these are normal faults separating a series of tilted blocks. In the upper part of the crust, the north–south-striking fault blocks contain north–south-striking dykes that locally constitute a sheeted complex. The faults correspond to marked lateral changes in the thickness and facies assemblages seen in supracrustal rocks, are locally marked by prominent subvolcanic breccias, and have upward decreasing throws suggesting that they are growth faults. The base of the volcano-sedimentary sequence is a major erosional surface in places, which can penetrate down to the Dunitic Zone. The evidence for coeval extension and magmatism, and the discovery of a locally well-developed sheeted dyke complex, suggest that the TMOC formed by sea-floor spreading. The dominance of a boninitic signature in cumulate and volcanic rocks suggests that spreading occurred in a subduction zone environment, possibly in a forearc setting.

Ophiolites are fragments of oceanic lithosphere transported onto a continental margin during orogenesis (Coleman 1971) and they may preserve evidence of the tectono-magmatic evolution of the fossil oceanic crust. It is now known that many ophiolites formed in suprasubduction zone environments (Robinson *et al.* 1983; Pearce *et al.* 1984;Robinson & Malpas 1990; Bloomer *et al.* 1995). Because boninitic magmatism seems to characterize forearc environments, ophiolites that contain abundant boninitic rocks are interpreted as having formed in forearcs (Lytwyn & Casey 1993; Bédard *et al.* 1998). In such settings, the overriding plate can consist of older oceanic crust, an older arc, or can be a tectonic collage of off-scraped material and accreted micro-terranes (Dickinson & Seely 1979; Dickinson *et al.* 1988), and its stress field can be compressional or extensional (Uyeda & Kanamori 1979; Hamilton 1988, 1995). Typically, the forearc basement is hidden beneath thick accumulations of volcanogenic sediments derived from the active arc (Stern & Bloomer 1992), hindering reconstructions of the previous geological history. The ease of access and quality of exposure available in forearc ophiolites afford an opportunity to understand the processes associated with these complex tectonic environments. However, the Ordovician ophiolites of the Canadian Appalachians have been overprinted by Ordovician (Taconic), Silurian (Salinic)

From: DILEK, Y. & ROBINSON P. T. (eds) 2003. *Ophiolites in Earth History*. Geological Society, London, Special Publications, **218**, 231–251. 0305-8719/03/$15

and Devonian (Acadian) events, which make it difficult to recognize pre-obduction structures. None the less, pre-obduction structures have been recognized in the Betts Cove (Tremblay *et al.* 1997; Bédard *et al.* 2000) and Bay of Islands (Karson *et al.* 1983; Bédard 1991; Berclaz *et al.* 1998; Suhr & Cawood 2001) ophiolites of Newfoundland. At Betts Cove, forearc extension appears to have proceeded to the point where seafloor spreading was initiated (Bédard *et al.* 1998; compare Stern & Bloomer 1992). The Thetford Mines Ophiolite Complex (TMOC) of southern Québec also has abundant boninitic lavas (Church 1977, 1987; Hébert & Laurent 1989; Laurent & Hébert 1989) and cumulates derived from boninites (Bédard *et al.* 2001), and appears to represent another fragment of forearc crust. This paper presents field evidence for an extensive and complex pre-obduction history for the TMOC.

Geological setting

The northern Appalachians are the result of successive Palaeozoic orogenic pulses. The Mid- to Late Ordovician Taconic Orogeny is thought to be due to the accretion of oceanic terranes (i.e. the Dunnage Zone) against the continental margin of Laurentia (i.e. the Humber Zone) (see inset map in Fig. 1). The tectono-stratigraphic zones of the Canadian Appalachians were defined on the basis of the Cambro-Ordovician geology (Williams 1979). The Baie-Verte–Brompton Line (BBL) separates the Humber and Dunnage Zones, is characterized by numerous ophiolite complexes, and has been interpreted as a continental–oceanic suture zone (Williams & St-Julien 1982). The Late Silurian, Mid-Devonian Acadian Orogeny is attributed to the final closure and destruction of Iapetus Ocean, when the Composite Avalonian Terrane collided with the Taconian Orogen developed upon the Laurentian margin (Osberg 1978; Williams & Hatcher 1983; Robinson *et al.* 1998; van Staal *et al.* 1998).

In southern Québec, the Dunnage Zone contains three principal ophiolite complexes (Thetford Mines, Asbestos and Orford) which are shown on existing maps as slivers embedded in the Saint-Daniel Mélange, interpreted by Cousineau & St-Julien (1992) as the vestiges of a subduction zone accretionary complex. Other important components of the Dunnage Zone in southern Québec are the Ascot Complex (*c.* 460 Ma), interpreted as a volcanic arc by Tremblay (1992), and the Upper Ordovician Magog Group, interpreted as a forearc sedimentary basin that developed during continuing convergence between the Laurentian margin and adjacent oceanic terranes (Cousineau & St-Julien 1992). The Humber Zone is characterized by a first generation of NW-verging thrusts of Mid-Ordovician age (469–460 Ma), and coeval regional metamorphism associated with ophiolite emplacement and crustal thickening (Pinet & Tremblay 1995; Tremblay & Castonguay 2002). Subsequently, Late Silurian to Early Devonian (*c.* 430–415 Ma; Castonguay *et al.* 2001) backthrusts and associated normal faults (e.g. the St-Joseph fault) exhumed Laurentian margin metamorphic rocks (Pinet *et al.* 1996a, 1996b; Castonguay *et al.* 2001; Tremblay & Castonguay 2002). The ophiolites of southern Québec occur in the hanging wall of the St-Joseph normal fault, which separates panels with a distinct metamorphic history, and which forms a composite structure with the BBL (Tremblay & Castonguay 2002).

The Thetford Mines (TMOC) and Asbestos (AOC) Ophiolite Complexes are composed of thick mantle and crustal sections, whereas only upper-crustal rocks are preserved in the Mt Orford Ophiolite Complex (MOC) (Oshin & Crocket 1986; Church 1987; Laurent & Hébert 1989; Hébert & Bédard 2000). Lavas and dykes in the MOC have backarc to forearc geochemical affinities (Harnois & Morency 1989; Laurent & Hébert 1989; Huot *et al.* 2002). Zircons from a trondhjemite intruded into the gabbroic crust contain inherited Proterozoic cores, and the least contaminated fractions give a U/Pb age of 504 ± 3 Ma (David & Marquis 1994). The TMOC and AOC, on the other hand, are dated at 479 ± 3 Ma and 478–480 $+3/-2$ Ma (Dunning *et al.* 1986; Whitehead *et al.* 2000), respectively, and are dominated by boninitic (forearc) lavas and dykes (Church 1977, 1987; Hébert & Laurent 1989; Laurent & Hébert 1989), and by cumulates derived from boninites (Hébert & Bédard 2000; Bédard *et al.* 2001).

The Thetford Mines Ophiolite Complex (TMOC)

The TMOC can be divided into the Thetford Mines Massif (TMM) in the north and the Mount Adstock–Ham Massif (AHM) in the south (Laurent 1975). The TMM has a thick mantle section (*c.* 5 km thick) and a thin crust (Fig. 1; Laurent *et al.* 1979). Pre-emplacement structures are difficult to recognize in the TMM, because they have been reworked by three types of syn- and post-emplacement structures (Brassard & Tremblay 1999; Schroetter *et al.* 2000, 2002), as follows.

(1) Syn-emplacement structures are associated with dynamo-thermal metamorphism, with an inverted metamorphic gradient ranging from greenschist grade at the base to upper amphibolite

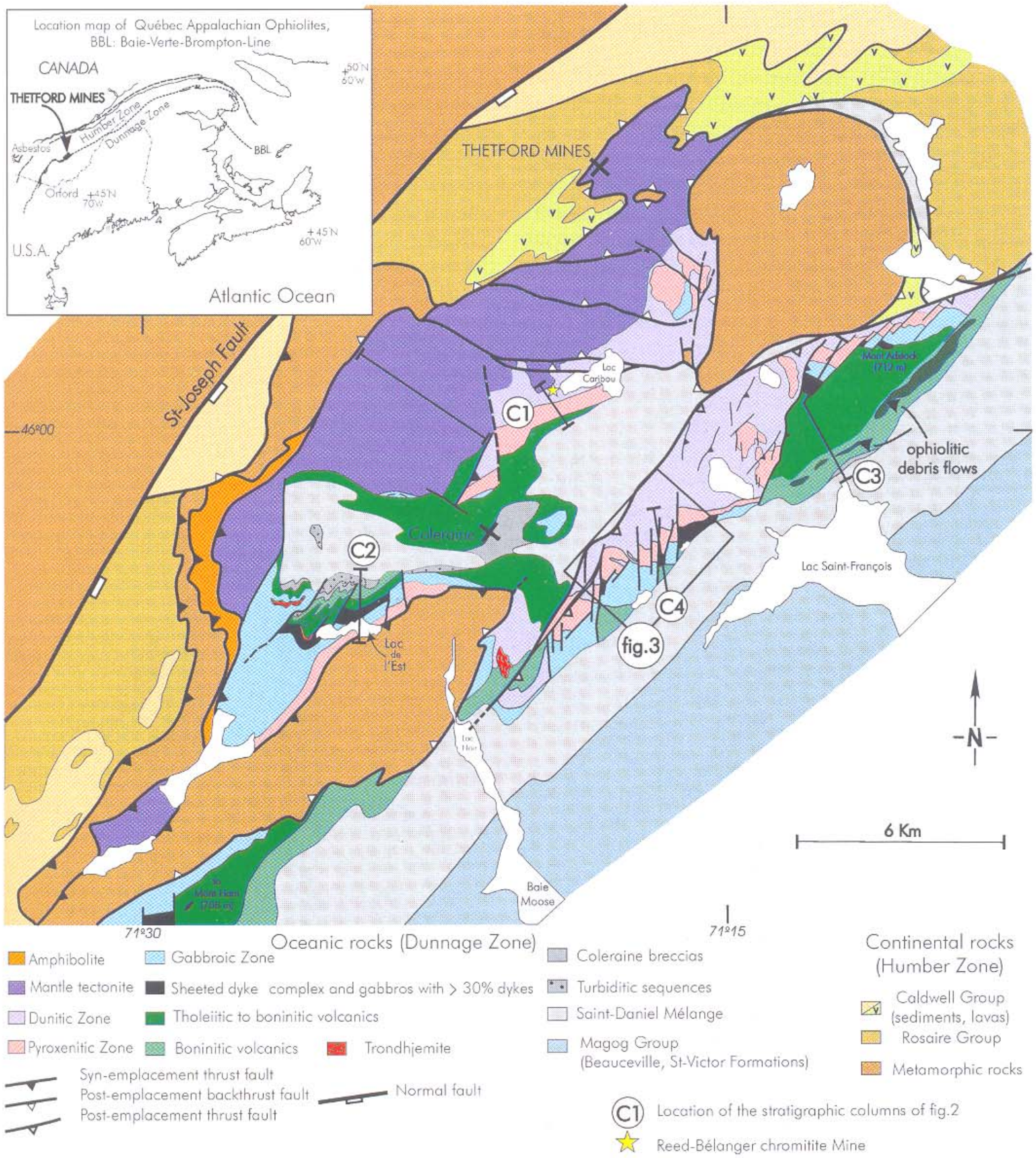

Fig. 1. Geological and structural map of the Thetford Mines Ophiolitic Complex, showing the location of the stratigraphic columns of Fig. 2, and the location of Fig. 3. The map is based on three summers of fieldwork (2000–2002) by our team, with additional data collected in previous field campaigns, complemented by data from previous maps (Cooke 1938; Hébert 1983; Pinet 1995; Brassard & Tremblay 1999).

grade at the contact with the mantle units (Feininger 1981). The most recent age for the metamorphic aureole is 477 ± 5 Ma (Ar–Ar, Whitehead *et al*. 1995); an age that is interpreted to reflect emplacement (Taconian Orogeny) onto the continental margin (Pinet & Tremblay 1995; Tremblay & Castonguay 2002).

(2) Post-emplacement structures include: (a) WSW–ENE- to east–west-striking, SE-directed backthrusts and associated folds, which are probably coeval with the Upper Silurian–Early Devonian backthrusts that characterize the Humber Zone; (b) subvertical NE–SW-striking, Mid-Devonian Acadian reverse faults and folds (Schroetter *et al*. 2000, 2002; Tremblay & Castonguay 2002). The original stratigraphy of the TMM has been tilted to the vertical by these events, and the ophiolitic rocks now occupy the hinge of a major

(multi-kilometre), overturned, SE-verging fold (Laurent 1975; St-Julien 1987).

According to Cousineau & St-Julien (1992), the black shales and pebbly mudstones of the Saint-Daniel Mélange represent a subduction zone accretionary complex, and its contact with the upper part of the TMOC is a fault. Others have interpreted this contact as stratigraphic and depositional (Dérosier 1971; Hébert 1983; Schroetter *et al.* 2000, 2002), implying rather that the Saint-Daniel Mélange is a piggy-back basin, and that it represents the lowermost unit of the forearc basin sequence, the Magog Group (Schroetter *et al.* 2000, 2002) (Fig. 2, column C4). Our field observations indicate that the northern margin of the AHM is indeed tectonic. In the Mt Adstock area (AHM, Fig. 1), an Acadian reverse fault juxtaposes ophiolitic rocks (SE side) against sedimentary rocks of the Saint-Daniel Mélange (NW side; Schroetter *et al.* 2002), whereas in the Mt

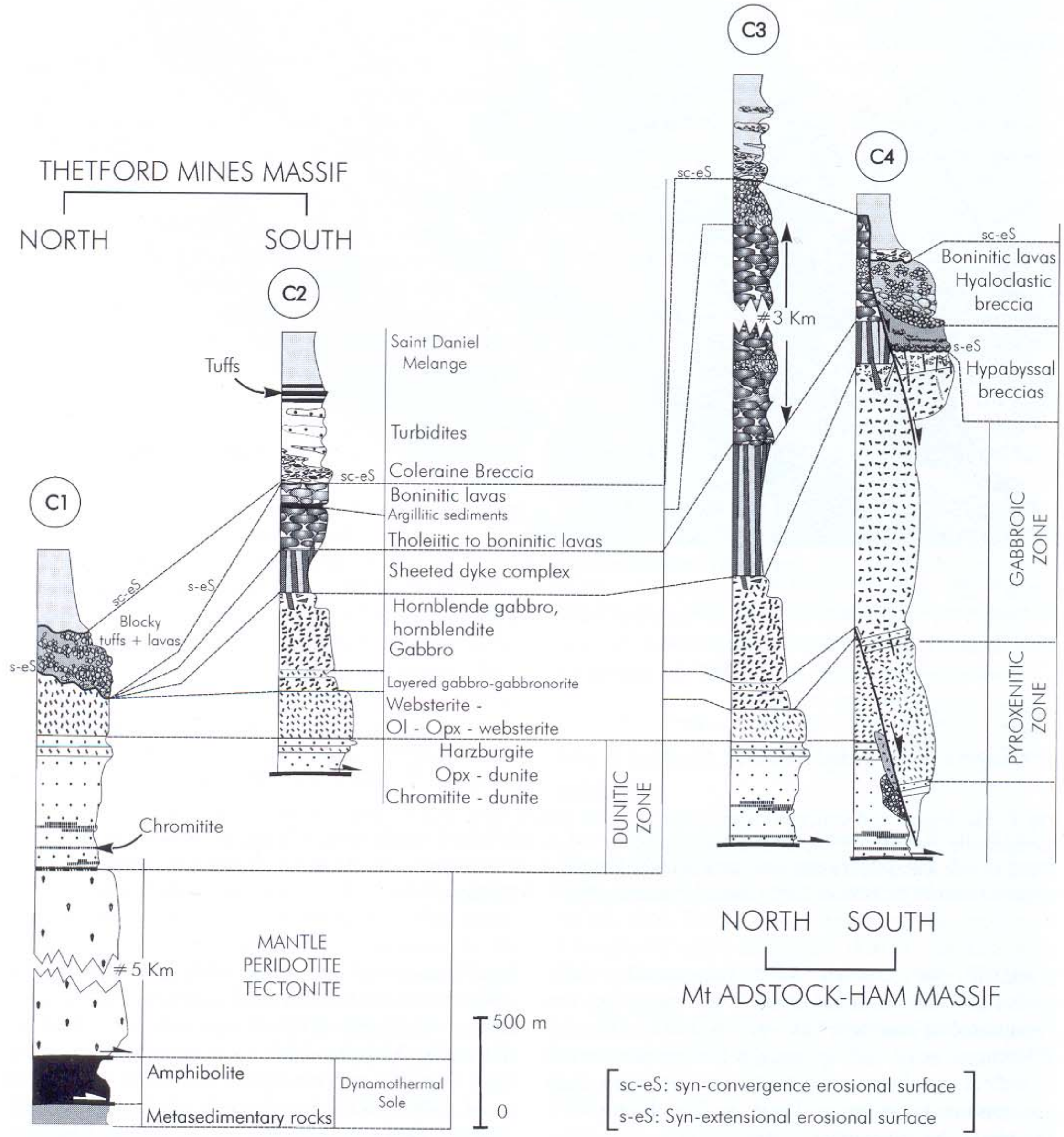

Fig. 2. Stratigraphic columns. Column C1, Thetford Mines Massif (TMM), Caribou Lake. C2, TMM, Lac de l'Est. C3 and C4 are from the northern and southern (Bisby Lake) parts of the Adstock–Ham Massif, respectively.

Ham area, metamorphosed continental margin rocks are backthrust onto the ophiolite. However, the southeastern contact between the Saint-Daniel Mélange and the AHM is depositional, and is marked by an erosional unconformity (Fig. 2).

The AHM is composed of ultramafic and mafic cumulates, layered to massive gabbroic rocks, ultramafic to mafic dyke swarms that locally grade into a sheeted dyke complex, tholeiitic to boninitic lavas and felsic pyroclastic rocks, upon which were deposited detrital sediments (Coleraine Breccia and the Saint Daniel Mélange). Late structures are of the same type as in the TMM (type 2 described above), but are less intensely developed. Our mapping shows that these post-emplacement structures are superimposed upon pre-existing, north–south- to NNE–SSW-trending, brittle–ductile to brittle faults, which we attribute to a pre-emplacement (pre-obduction) extensional phase (Figs 1 and 3).

Stratigraphy

The stratigraphy of the TMOC is shown in the four columns of Figure 2. Columns C1 and C2 are from the northern and southern parts of the TMM, respectively, and columns C3 and C4 are from the northern and southern parts of the AHM, respectively. The main points to note are: (1) the existence of significant lateral variations in the thicknesses of cumulate, volcanic and sedimentary facies; (2) the presence of an erosional surface (s-eS in Fig. 2) within the ophiolitic crust that penetrates down to the Dunitic Zone in column C1, and to the Gabbroic Zone in column C4 (Fig. 2); (3) sedimentary rocks of the Saint-Daniel Mélange are in depositional contact with underlying ophiolitic rocks, with the basal member (Coleraine Breccia) infilling a palaeo-topography, with coarse debris flow deposits above thick lava sequences, and thin-bedded siltstones and mudstones above thin lava sequences.

In the next section we describe typical plutonic, hypabyssal and supracrustal facies of the AHM, with only brief mention of equivalent facies from the TMM. The early structures (faults) that dissect the AHM crust are then described. Swarms of hydrothermal veins containing amphibolite and greenschist mineral assemblages are common throughout the section, but are not evenly distributed. Most rocks are affected by pervasive intra-oceanic hydrothermal metamorphism, but textural pseudomorphism and the absence of younger penetrative deformation generally allow primary features to be identified; for this reason we omit the prefix 'meta-' in the following.

Mantle section

The harzburgite tectonites of the TMM show near-ubiquitous planar and linear fabrics defined by alignment and elongation of orthopyroxene and chromite grains. Occasional compositional layering, pyroxenite dykes, podiform dunites and chromitites are discordant to this high-temperature tectonite fabric (Laurent *et al.* 1979). Emplacement-related fabrics are developed near the contact with the dynamo-thermal sole. Mineral chemical and whole-rock geochemical data imply that the harzburgite is residual from extensive partial fusion (Hébert 1985; Hébert & Laurent 1989). Peraluminous two-mica granite intrusions have high $^{87}Sr/^{86}Sr$ initial ratios and igneous zircons with low $^{208}Pb/^{206}Pb$ ratios, and have yielded 469 ± 4 to 470 ± 5 Ma crystallization ages (U/Pb on zircon, Whitehead *et al.* 2000), suggesting that these granites were derived by fusion of continental margin sediments during emplacement of the hot ophiolite (Clague *et al.* 1985; Whitehead *et al.* 2000). The transition from mantle tectonite to cumulate crust is rarely observed, and where exposed (north of Lac Caribou), it is a post-emplacement backthrust fault (Schroetter *et al.* 2000).

Plutonic crustal section

We divide the plutonic crustal section into three zones.

The Dunitic Zone is up to 500 m thick (Fig. 1 and column C1 in Fig. 2). Above a thin (50 m) basal dunite subzone, it is common to find rhythmic alternations of dunite and massive chromitite beds. Chromitites are typically 1–10 cm thick, with a maximum of 12 m observed in the Reed–Bélanger deposit (see Figs 3 and 4d). Chromitite beds may show modal grading but do not define a consistent polarity. Localized development of schlieren textures and tight isoclinal folds suggest an early, high-temperature deformation event (Kacira 1971). The origin of the rhythmic chromitite–dunite bedding is uncertain. It could reflect the interplay of fractional crystallization, magmatic currents, wallrock assimilation and chamber replenishment. In any case, it appears to have originated as near-horizontal primary bedding, and we use these orientations to define the palaeo-horizontal in the Dunitic Zone. It is noteworthy that immediately below the main chromitite layer of the Reed–Bélanger Mine, there are at least four pyroxenite layers (10–30 cm thick) composed of orthopyroxenitic clots in a dunitic matrix. These pyroxenite layers are size-graded, are oriented subparallel to the chromitites, and all give the same, normal, way up. They are inter-

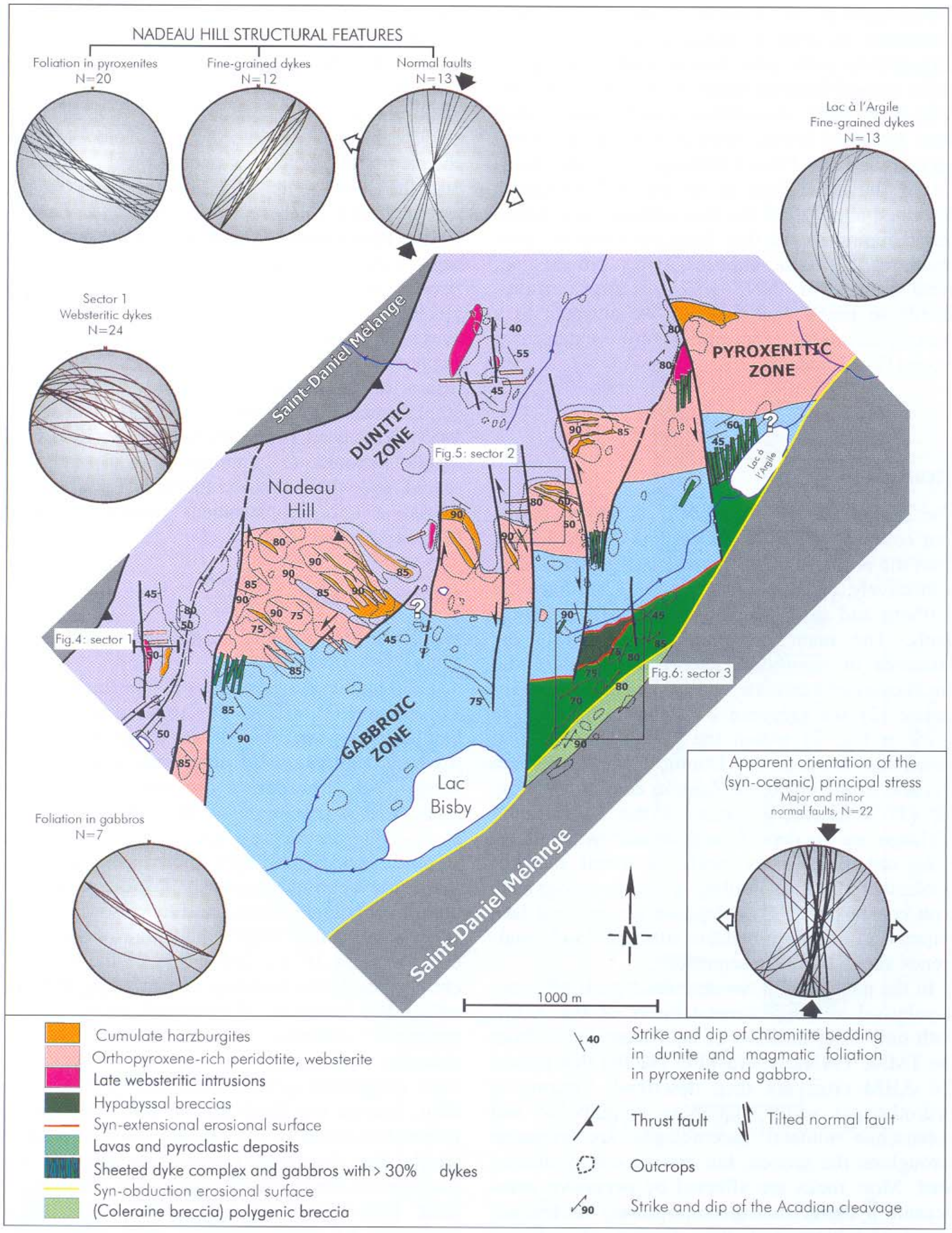

Fig. 3. Geological map of part of the Adstock–Ham Massif, with equal area stereograms (Schmidt) of structural features, and the locations of Figs 4–6. It should be noted that because of Siluro-Devonian tilting to the vertical, this map-view is essentially a cross-section through the oceanic crust.

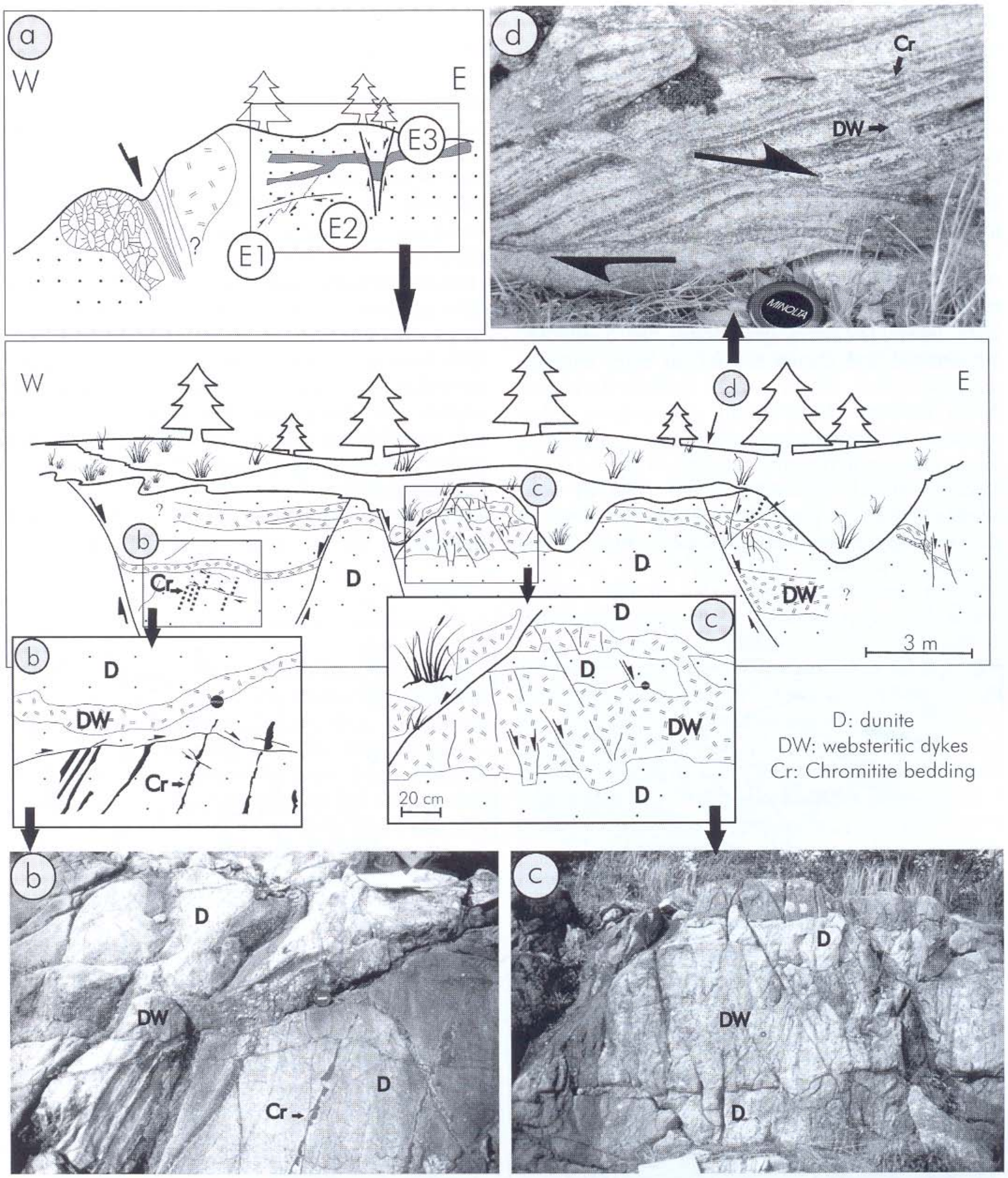

Fig. 4. Photographs and sketches illustrating syn-oceanic deformation features from the lower crust in Sector 1. (**a**) Schematic cross-section summarizing the different tectonic events and illustrating the relation between the major normal fault, the chromitite bedding, the minor and major intrusions, and deformation. (**b**) Chromitite beds are tilted to the west and crosscut by websterite dykes (E2 event). (**c**) Websterite dykes are truncated by sub-north–south normal faults, forming a horst-and-graben pattern (E3 event). (**d**) Chromitite beds are tilted and crosscut by peridotite dykes (E2 event); high-temperature synmagmatic(?) deformation (E1 event) parallel to chromitite beds should be noted.

preted to represent some type of depositional event, and show no evidence of tectonic inversion or refolding. The disappearance of chromitite beds higher in the section marks the transition from the rhythmic dunite–chromitite subzone to the massive dunite subzone, which is dominated by dunites having only 1–1.5% of disseminated chromite. The top of the Dunitic Zone is characterized by the presence of disseminated orthopyroxene (*c.* 5%) and minor chromite (c. 1%).

Fine-grained websterite dykes (Fig. 4b and c) are typically oriented sub-east–west (in the AHM) with subhorizontal to shallow dips.

The Pyroxenitic Zone has a maximum thickness of 600 m but is commonly beheaded by an internal erosional surface (s-eS in Fig. 2, column C1). A basal rhythmic subzone (5–15 m) is characterized by alternations of harzburgite, orthopyroxenite and websterite layers (30–60 cm) (Fig. 5b1), which are used to define the palaeo-horizontal. In the harzburgite, cumulus orthopyroxene grains may be aligned parallel to layering. The pyroxenite layers (3–8 mm grain size) are commonly boudinaged and transposed by an early tectono-magmatic event, which is more strongly developed in the Thetford Mines Massif. Above these orthopyroxene + olivine-dominated rocks there is a thick, massive pyroxenite subzone, which is dominated by thickly layered (typically 1–3 m) websterites. Layering generally reflects variations in the orthopyroxene/clinopyroxene ratio. Some layers exhibit modal grading, with a concentration of cumulus-textured clinopyroxene at the top. Thin olivine-rich 'layers', schlieren or veins may separate pyroxenite layers. Embedded within these massive websterites, there are repetitions of rhythmically layered harzburgite–websterite sequences (10–15 m thick) that are very similar to those of the basal rhythmic subzone. The nature of these embedded harzburgite–websterite sequences is uncertain. They may be the result of magma chamber replenishments, or they could represent sills affected by the high-temperature tectono-magmatic deformation. Sills of massive dunite (20–30 m thick), and of undeformed harzburgite–lherzolite (2–3 m thick) intrude the massive websterites. A size-graded lherzolite sill gives a way-up direction towards the SSW. Fine-grained dykes oriented perpendicular to layering are locally abundant.

The Gabbroic Zone is up to 1200 m thick (depending on the depth of penetration of the

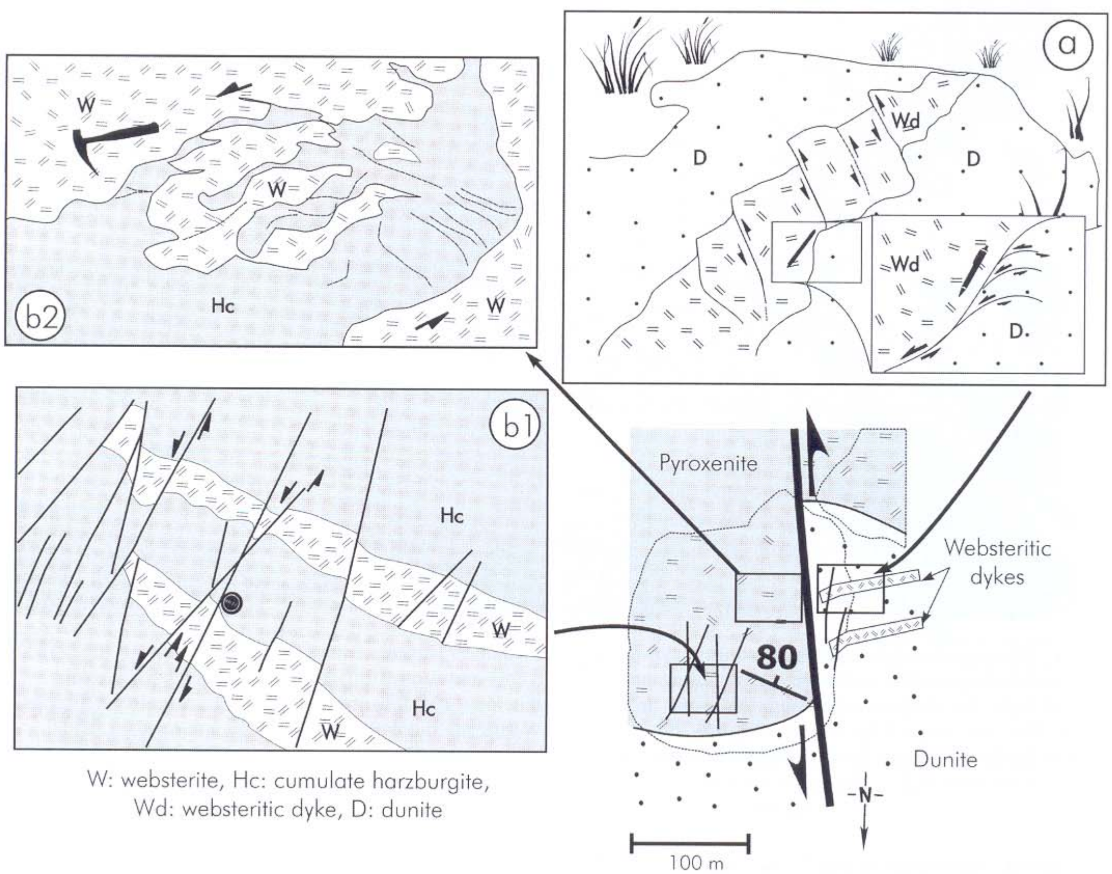

Fig. 5. Sketches illustrating syn-oceanic deformation features from the lower crust in Sector 2. (**a**) Dunitic Zone, websterite dyke in dunite is cut by north–south normal faults oriented parallel to the main fault structure. (**b1**) Pyroxenitic Zone, alternation of cumulate websterite and harzburgite layers, truncated by north–south normal faults that define a horst-and-graben pattern. (**b2**) Near the main fault, websterite–harzburgite layers are more chaotic.

denudational surface) and is organized on a gross scale, with interlayered norites and gabbronorites at the base, gabbros *sensu stricto* in the middle, and an upper complex composed of hornblende gabbro, hornblendite, dykes, cataclasites, trondhjemitic intrusions and breccia veins. The norites, gabbronorites and gabbros are fine to medium grained, and thin, laterally discontinuous melagabbro to pyroxenite (1–5 cm) layers are common. The medium-grained to pegmatitic hornblende gabbros are either layered or chaotically varitextured, and are associated with coarse-grained hornblendite pods and layers. Trondhjemite forms intrusions of various size (1 cm to 30 m) within the uppermost gabbros, and may form a breccia fill between angular gabbroic clasts; but trondhjemites also occur to a lesser extent as dykes within the Dunitic and Pyroxenitic Zones. Zircons extracted from TMOC trondhjemites gave U/Pb ages of 478 +3/−2 and 480 ± 2 Ma, which were interpreted as crystallization ages by Whitehead *et al.* (2000).

Hypabyssal facies

Two types of hypabyssal facies rocks occur above the Gabbroic and Pyroxenitic Zones: dyke swarms and breccias. The dykes (30 cm to *c.* 1 m thick) are mafic to ultramafic, locally orthopyroxene-phyric, but more commonly show microgabbroic or aphanitic textures. Microgabbro dykes have thin (1–2 cm) aphanitic chilled margins. Dyke margins are locally brecciated by late hydrothermal vein networks. Dykes are most commonly oriented north–south, and cut the magmatic foliation in host plutonic rocks at *c.* 60–90°. In some sectors, the dykes constitute 40–100% of the outcrop over hundreds of metres, and are mapped as a sheeted dyke complex (Figs 1 and 3).

The breccia facies was not previously recognized in the TMOC. The breccias reach a maximum of 150 m in thickness and separate plutonic and volcanic sequences. Where we have studied them in detail (north of Lac Bisby, Fig. 3; column C4 in Fig. 2, and Fig. 6), the breccia facies caps the gabbroic sequence and is overlain by boninitic lavas and volcaniclastic deposits. In this area, the brecciated hypabyssal facies is characterized by alternations of breccia (0.5–3 m thick) and amygdaloidal sills (Fig. 6c). The breccia is composed of angular clasts (90–40%, 1–10 cm) of aphanitic 'dolerite', microgabbro and gabbro. The angularity of clasts indicates only limited transport, and jigsaw-puzzle textures imply *in situ* brecciation. The matrix is typically igneous (microgabbro), but hydrothermal assemblages are also prominent locally. Intra-breccia sills (Fig. 6b and d) (1 cm–2 m) have very irregular shapes, contain clasts of adjoining breccia at top and bottom, are commonly amygdaloidal and locally orthopyroxene porphyritic. Similar breccia facies occur above ultramafic rocks of the Dunitic Zone near Lac Caribou (Fig. 2, column C1), but in this instance, the clasts in the breccia are composed of peridotites and pyroxenites, and the overlying debris flows and pyroclastic deposits are rich in gabbroic and pyroxenitic fragments. Because platinum group element (PGE) mineralization occurs within the Pyroxenitic Zone (e.g. Star Chrome showing: Pagé *et al.* 2001), erosion that penetrates to the Dunitic Zone implies that palaeo-placer PGE deposits might represent a new type of exploration target in the area.

Volcano-sedimentary facies

The volcanic and volcaniclastic rocks of the ophiolite exhibit marked lateral changes in thickness and lithology (Fig. 2). North of Lac Bisby (Fig. 1), the volcaniclastic rocks are made of blocky tuffs (2–20 m thick) containing rounded pillow-lava fragments (10 cm average, with a few larger blocks), in a sandy volcaniclastic matrix (Fig. 2, column C4). Vesicular pillow lavas of 1–1.5 m in size alternate with smaller pillows (0.5 m average), with intercalated massive flows, hyaloclastite breccias, and possible submarine talus breccias. South of Lac Caribou there are abundant pyroclastic flow breccias containing rounded clasts of dacite, gabbro and pyroxenite, with intercalated fine-grained dacitic tufs (1–2 m thick) and argillites (Fig. 2, column C1). At the Lac de l'Est section, a 1–2 m red argillite separates a lower volcanic unit composed of tholeiites and boninites from an upper unit dominated by boninites (Hébert 1983; Hébert & Bédard 2000). The volcanic and volcaniclastic rocks of the TMOC are everywhere capped by a thick-bedded polygenic breccia (the Coleraine Breccia) that contains ophiolitic and metasedimentary fragments in an epiclastic matrix, which appear to represent submarine debris flows (Hébert 1981). The Coleraine Breccia grades up and laterally into the Saint-Daniel Mélange (Schroetter *et al.* 2002).

Deformation

A major problem in the structural analysis of ophiolites is the identification of marker surfaces allowing the geometry of deformation to be constrained. We used the orientation of chromitite bedding as palaeo-horizontal markers in the Dunitic Zone, of harzburgite–websterite rhythmic layering and mineral foliation as palaeo-horizontal markers in the Pyroxenitic Zone, and of pyroxenite–gabbro rhythmic layering and mineral folia-

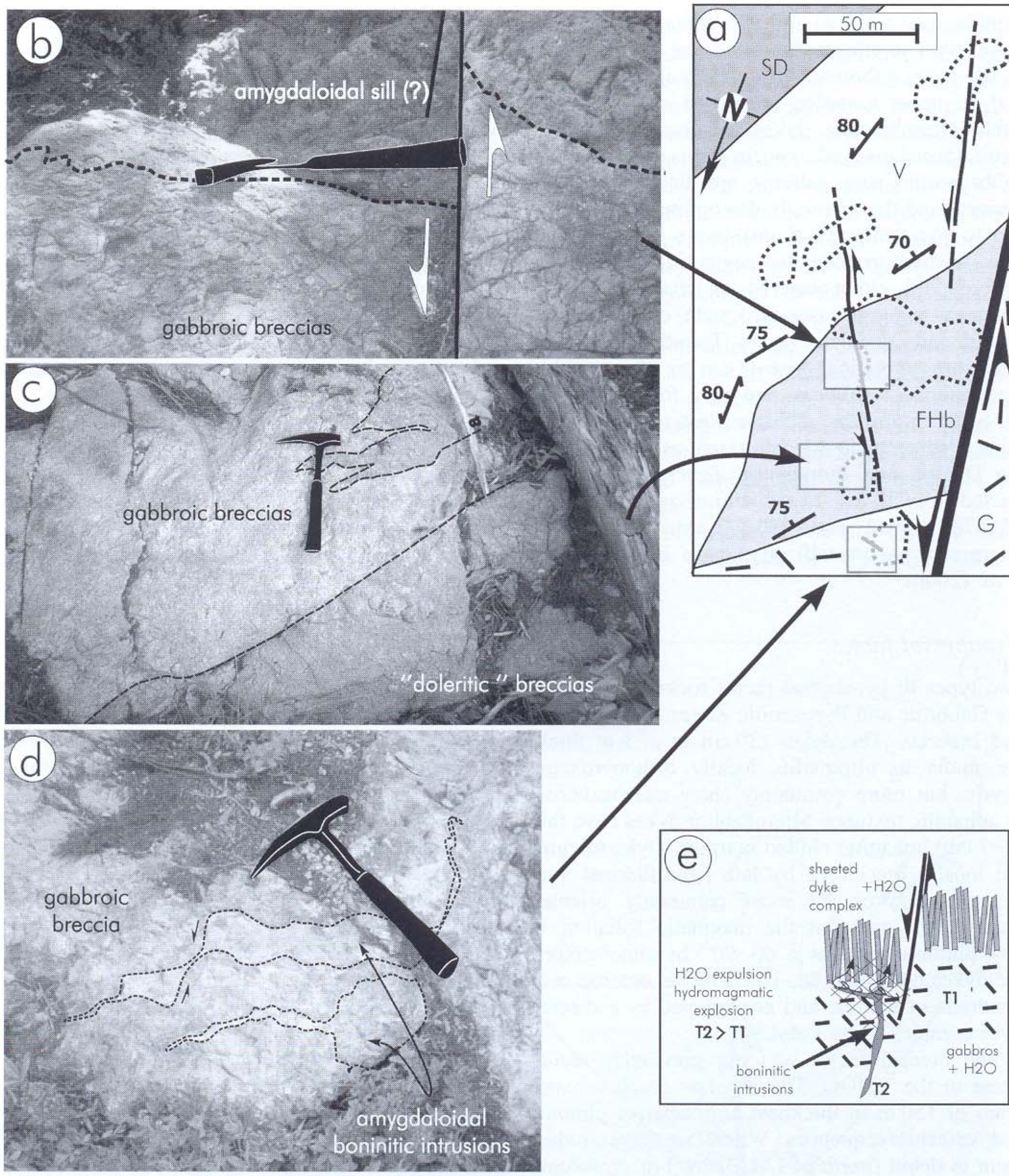

Fig. 6. Photographs and sketches illustrating syn-oceanic deformation features from the upper crust. (**a**) Detailed map of Sector 3. V, volcanic rocks; FHb, 'doleritic' and gabbroic breccias; G, upper gabbro; SD, Saint-Daniel Mélange. (**b**) Alternation of amygdaloidal mafic sills and gabbro to microgabbro breccias, all crosscut by a brittle normal fault. (**c**) Alternation of breccia composed of fine-grained mafic clasts (grey) and gabbro to microgabbro clasts (light grey). (**d**) Boudinaged boninitic sill emplaced in microgabbro breccia. (**e**) Interpretative sketch showing formation of the brecciated hypabyssal facies. T1, oceanic crust temperature; T2, intrusion temperature.

tion as palaeo-horizontal markers in the Gabbroic Zone. Dyke contacts were considered to have been originally palaeo-vertical and oriented perpendicular to the spreading direction (Cann 1974), allowing the determination of the palaeo-stress field in the upper crust. Finally, bedding in fine-grained sedimentary rocks occurring in the volcano-sedimentary superstructure was assumed to originally have been horizontal. Although none of these assumptions are fail-safe, the fact that they all provide similar answers suggests that they are reasonable.

In the AHM (Fig. 1), there are two main generations of faults. Younger reverse faults are associated with open folds and a regionally distributed, vertically dipping fracture cleavage trending *c.* NE–SW. An older set of subvertically dipping faults that trend north–south to NNE–SSW had not been recognized hitherto. The field relations we describe next lead us to believe that these older structures are related to intra-oceanic extension and sea-floor spreading. The origin of the older, high-temperature, tectono-magmatic deformation that affects much of the lower crust will not be discussed in this paper, and remains a major unresolved problem.

Considered on the scale of the AHM, chromitite beds do not display a consistent orientation, with each major fault block (see below) having its own attitude. Near major north–south faults, chromitites appear to be transposed into parallelism with the fault (Figs 3 and 4d). The magmatic foliation in the uppermost Dunitic Zone and Pyroxenitic Zone is much more regular, trending *c.* 125–130°, and dipping steeply (85–90°) to the SSW. Most dykes from the AHM are oriented *c.* north–south, dip steeply to the west (Fig. 3), and cut the foliation and layering in host plutonic rocks.

In contrast to the great regularity shown by the Pyroxenitic Zone, the magmatic foliation in the Gabbroic Zone is much more variable, although always very steeply dipping. North of Lac Bisby, the foliation in the gabbros strikes between 40° and 170°, whereas at Lac de l'Argile, it strikes *c.* 60° (Fig. 3). At Mt Adstock (Fig. 1), the layering in the gabbros is more regular, striking *c.* 110°. Dykes at Mt Adstock dip subvertically and strike 80°, as does the pyroxenite–gabbro contact. Cross-cutting relationships suggest that the pyroxenite–gabbro contact is, at least locally, an intrusive one, with pyroxenites cutting older gabbros.

Our mapping shows that faults striking from north–south to 20° commonly terminate the along-strike extension of rocks of the Dunitic, Pyroxenitic and Gabbroic Zones, or separate plutonic from volcanic rocks (Figs 1 and 3). These north–south- to 20°-striking faults are locally reactivated by post-emplacement (Silurian and Acadian) deformational events. We have mapped in some detail an area that appears to be largely free of post-emplacement deformation (Fig. 3) to illustrate the distribution of lithologies, and the orientation of the various marker surfaces. It should be noted that, because this part of the AHM was tilted to the subvertical during post-emplacement deformation (Silurian and Acadian), the map view (Fig. 3) is effectively a cross-section through the crust. Offsets between unit boundaries indicate an apparent sinistral motion in map view (Fig. 3). Apparent fault throws decrease from the base (north) to the top (south), and are capped by deposition of synconvergence sediments of the Saint-Daniel Mélange. The apparent throw calculated from the Dunitic–Pyroxenitic Zone boundary offset is *c.* 300 m on average, with a maximum of *c.* 1 km for the fault on the western flank of Nadeau Hill (Fig. 3). It should be noted also that the thickness of various lithological packages varies to either side of these faults. The map patterns indicate that the upper part of the crust in this part of the AHM was dissected by these north–south faults into a series of 100 m to 1 km blocks. Detailed observations from three sectors (labelled 1, 2 and 3 in Fig. 3) allow a kinematic analysis of this deformation. Sectors 1 and 2 are from the Dunitic and Pyroxenitic Zones, which are characterized by higher temperatures, whereas Sector 3 is from the cooler and more brittle upper crust.

Lower-crustal deformation

We now describe the two sectors we have analysed in detail, and then present some generalizations. Sector 1 (Fig. 4, location shown in Fig. 3) is entirely contained within the Dunitic Zone. From west to east, massive dunite with disseminated chromite gives way (over 0.5 m) to a thick (hundreds of metres) breccia composed of angular, 1–10 cm clasts of dunite, locally with chromitite beds within them, in an orthopyroxenitic stockwork. As the main fault plane is approached, sheared serpentinites appear in the dunite, culminating in a 2–3 m wide serpentinite mylonite that marks the core of the fault. These breccias and mylonites are cross-cut by undeformed, tabular websterite and lherzolite intrusions (30–50 m wide), which are oriented north–south, parallel to the main fault. The overall geometry is illustrated by Figure 4a. The fault kinematics are clearly expressed in outcrops located *c.* 50 m east of the main fault plane, where chromitite beds strike north–south and dip 50° to the west (Fig. 4b and d). The chromitites are cut by shallowly dipping, east–west-striking websterite dykes. The websterite dykes lack chilled margins against their dunitic hosts, suggesting that these rocks were still hot at the time of dyke emplacement. A series of faults associated with serpentinite veins are parallel to dyke contacts and offset chromitite beds to the east (Fig. 4b). On the same outcrop, these same websterite dykes are chopped up into centimetre-scale horst-and-graben structures by a series of conjugate, steeply dipping, north–south-striking normal faults (Fig. 4c).

The history of deformation in Sector 1 can thus be divided into three increments (E1–E3, Fig. 4a). The early event (E1) corresponds to the localized development of a high-temperature layering-parallel fabric (chromitite schlieren and isoclinal folds (Kacira 1971)). Restoration to the horizontal of the chromitite beds (50° tilts, Fig. 4b and d) gives the websterite dykes and parallel faults of increment E2 a subvertical orientation, and so E2 faults are interpreted as having originally been steeply dipping, normal faults. The last increment of deformation (E3) defines a horst-and-graben system (Fig. 4c), marking a continuation of extension.

Sector 2 (Fig. 5, location shown in Fig. 3) contains a fault that crosses the Dunitic–Pyroxenitic Zone contact. The Dunitic Zone here is very similar to that of Sector 1 as described above, whereas the Pyroxenitic Zone (Fig. 5b1) is composed of steeply dipping websterite and harzburgite layers that strike 125°. The main fault plane is highlighted by the presence of strongly sheared blue dunite. Immediately to the west of the main fault (in the Dunitic Zone), east–west-trending, shallowly south-dipping websterite dykes are offset sinistrally by steeply dipping, north–south-striking faults oriented parallel to the main fault plane (Fig. 5a). In the Pyroxenite Zone to the east of the main fault plane, websterite layers embedded in harzburgite are stretched and boudinaged, and a mineral foliation is developed parallel to layering. Immediately adjacent to the main fault plane, layering becomes more chaotic, with the development of disharmonic folds (Fig. 5b2). The high-temperature fabrics are cut into horst-and-graben structures by steeply dipping faults (Fig. 5b1).

Restoration to the horizontal of the websterite–harzburgite layering converts the main fault of Sector 2 into a steeply dipping normal fault, with downthrow of the eastern compartment. Shear-sense indicators and the orientation of faults affecting rocks of the Pyroxenitic and Dunitic Zones of Sector 2 have the same orientation as faults of episode E3 in Sector 1, and we infer a correlation. However, it is not clear whether the high-temperature fabric in Sector 2 corresponds to E2 or E1 in Sector 1.

To summarize, both sectors register several extensional deformation events. As fault blocks are variably rotated by events E2 and E3, this explains why the early tectono-magmatic fabric (E1) is not uniform in the AHM. The E3 event is particularly prominent, and affects most of the plutonic crust of the AHM. Restoration to palaeo-horizontal of marker horizons implies that the principal E2 and E3 structures were originally steeply dipping normal faults that controlled the map-scale distribution of lithologies. Our mapping shows that massive, undeformed intrusions of peridotite and pyroxenite were commonly emplaced in the immediate vicinity of the main faults (Fig. 3), in some cases being injected within the fault breccias, or along fault planes (Fig. 4d). This demonstrates the synmagmatic nature of the faults, and suggests that they guided the ascent of magmas.

Upper-crustal deformation

Upper-crustal deformation (Gabbroic Zone, Hypabyssal and Extrusive sequences) is characterized by brittle fractures and brecciation (Fig. 6a and c), and by a structural control on the orientation of dykes (Fig. 3). Brittle to brittle–ductile faults in the upper crust are steeply dipping and oriented roughly north–south. They cut rocks of the Gabbroic Zone, the sheeted dyke complex, and the hypabyssal breccia facies (Fig. 6b). Most have small throws (tens of centimetres) and show an apparent sinistral sense of motion, similar to the faults in the lower crust. In the hypabyssal breccia facies, amygdaloidal intrusions injected into the microgabbroic breccia are stretched and offset by the faults (Fig. 6b and d), suggesting that faulting and magmatism were coeval. The faults must have played a role in brecciation, because some breccias have a cataclastic matrix, and our mapping shows that the brecciated hypabyssal facies is preferentially developed along the extension of the major north–south normal faults described in the previous section. However, the breccia matrix is generally igneous, and the jigsaw-puzzle morphology of the rocks (Fig. 6b) seems more compatible with some type of magmatic hydro-fracture mechanism, perhaps complemented by phreatomagmatic explosions caused by ascent of magma into rocks impregnated with seawater, and possibly by volume expansion caused by vesiculation of ascending, water-rich magmas (Fig. 6e). In any case, the faults would still have played a role in brecciation, as they would represent pathways for the ascent of magma and infiltration of seawater.

Throughout the TMOC, dykes are typically oriented approximately north–south, subparallel to the synmagmatic normal faults described above. An exception is a dyke swarm located south of Nadeau Hill (Fig. 3), where the main dyke trend is *c.* 40°. It is difficult to know if this difference in the trend of the dykes from Nadeau Hill is an original extensional feature (suggesting progressive rotation of fault blocks; see Dilek *et al.* 1998), or if it reflects a syn- or post-emplacement reactivation. None the less, the general parallelism between dykes and normal faults throughout the AHM suggests a genetic link, whereby both features would have formed synchronously during east–west crustal extension (sea-floor spreading).

A preliminary model for the structural and magmatic evolution of the TMOC

The style and sequence of syn-oceanic deformation we have documented (Figs 4 and 5) allow us to propose a possible evolutionary scenario for the main, boninitic phase of the TMOC (Figs 7 and 8). The starting point is a layered plutonic crust composed of dunite at the base, pyroxenite in the middle, and gabbro at the top, into which is rooted a sheeted dyke complex that fed overlying lavas (Fig. 8a). During a first, high-temperature, syn-magmatic deformation event, many of the plutonic rocks were transposed and recrystallized along planes subparallel to original bedding planes. The cause of this first event is not certain. It may reflect shear between crust and mantle driven by a diapir (Nicolas 1992; Rabinowicz *et al.* 1993), extensional shear as crust slides away from the ridge axis, or may represent the way deformation

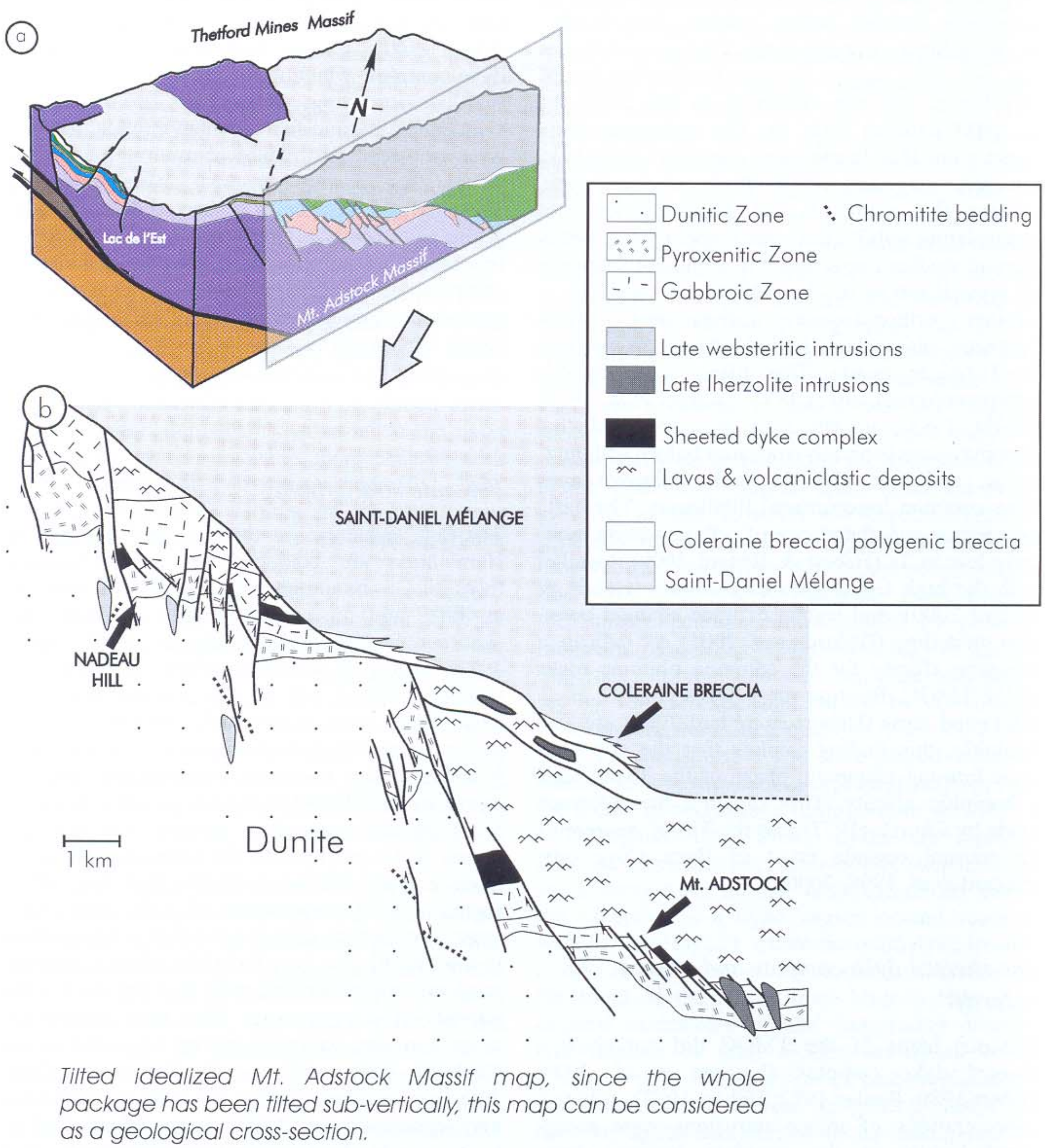

Fig. 7. (**a**) Interpretative palaeogeographical reconstruction of the Thetford Mines Ophiolitic Complex before its emplacement. Colours are the same as in Fig. 1. (**b**) Idealized map cross-section of the Adstock–Ham Massif.

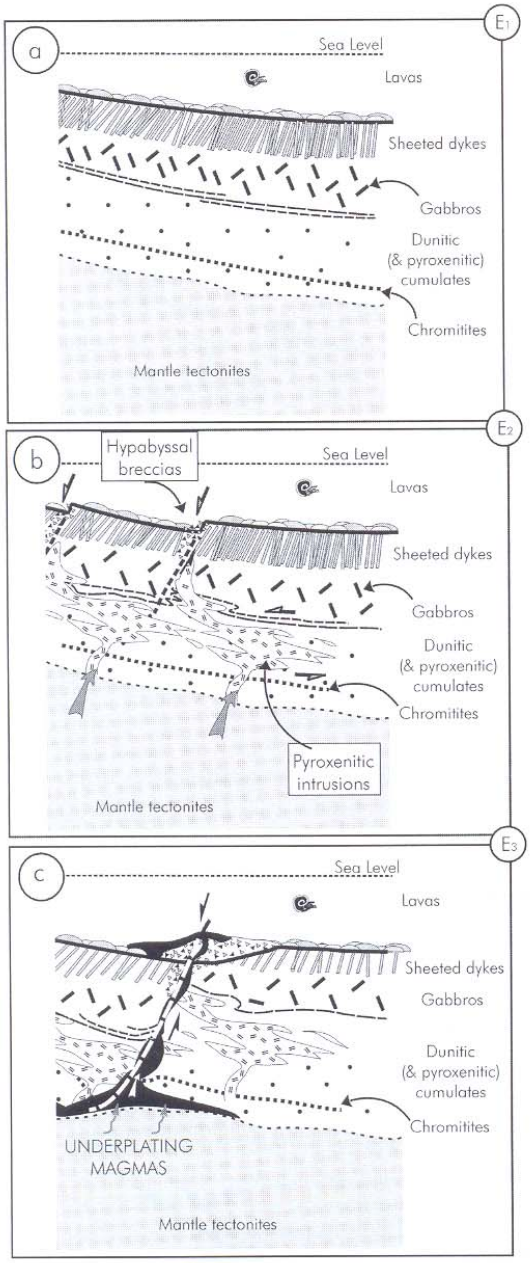

Fig. 8. Schematic illustrations of a possible evolutionary scenario for the main, boninitic, crust-forming event of the Thetford Mines Ophiolitic Complex. (**a**) Event E1. Development of a high-temperature foliation in cumulate rocks. (**b**) Event E2. Synkinematic pyroxenitic intrusions (grey arrows) are emplaced and transposed by continuing high-temperature tectonomagmatic deformation, as are harzburgite and websterite cumulates. Faults begin to form when magmatism wanes. Subvolcanic and talus(?) breccias form in the uppermost crust. (**c**) Event 3. Main stage of faulting, with only minor late-kinematic intrusions and associated lavas. Crust breaks up into horsts and grabens, and blocks are tilted along a basal décollement. The tops of tilted blocks are eroded, locally removing much of the upper crust.

is partitioned in an extending ductile crust at the spreading centre (Tapponier & Francheteau 1978; Harper 1985).

The overall regime remains tensional, however, and continuing magmatism is expressed as dykes of peridotite and pyroxenite (E2, Fig. 8b), some of which were probably emplaced as sills within pre-existing cumulates. The extent to which rocks of the Dunitic and Pyroxenitic Zones belong to the early cumulate 'stratigraphy', and the extent to which they represent under- and intra-plating intrusions is not certain (seeThy 1990; Bédard 1991, 1993). In many cases, these intra-cumulate intrusions would have been transposed and recrystallized during continuing high-temperature deformation, and so would no longer be recognizable as late intrusions. The presence of pyroxenitic debris within graded layers near the base of the Dunitic Zone implies that some dunites are cumulates formed within intra-plating or under-plating sills, with the pyroxenitic debris being derived from disaggregation of a pre-existing roof.

The high-temperature foliation is truncated by the major north–south faults we have documented. On outcrop and map scales, these faults dissect the crust into horst-and-graben structures. Kilometre-scale tilted blocks develop in the rigid upper plutonic crust and overlying extrusive rocks. An upward decrease in throw on these faults indicates that they are growth faults, propagating from bottom to top. Associated intrusions (Figs 3 and 6) imply that these faults were coeval with magmatism and played a role in magma ascent. The Dunitic Zone is only partly affected by these faults, and the lower part of the Dunitic Zone could represent a décollement surface accommodating movement of the tilted blocks (e.g. Harper 1985). The presence of a brecciated hypabyssal facies in areas where the sheeted dyke complex is absent suggests that erosion and/or tectonomagmatic destruction of upper-crustal facies was probably an important part of the geological history of the TMOC (Fig. 8c).

The TMOC, a slow-spreading environment?

The morphology of the axial rift zone of spreading centres is dependent on spreading rate, with prominent fault scarps and a deep axial valley characterizing slow-spreading environments (Macdonald 1982). Seismic data imply that magma chambers are ephemeral at slow-spreading ridges, with the depth of seawater penetration, and the depth of the brittle–ductile transition depending on the presence or absence of magma chambers (Harper 1985; Toomey *et al.* 1988; Dick *et al.* 1992; Dilek & Eddy 1992; Dilek *et al.* 1998). The deep graben of the TMOC, the deep crustal

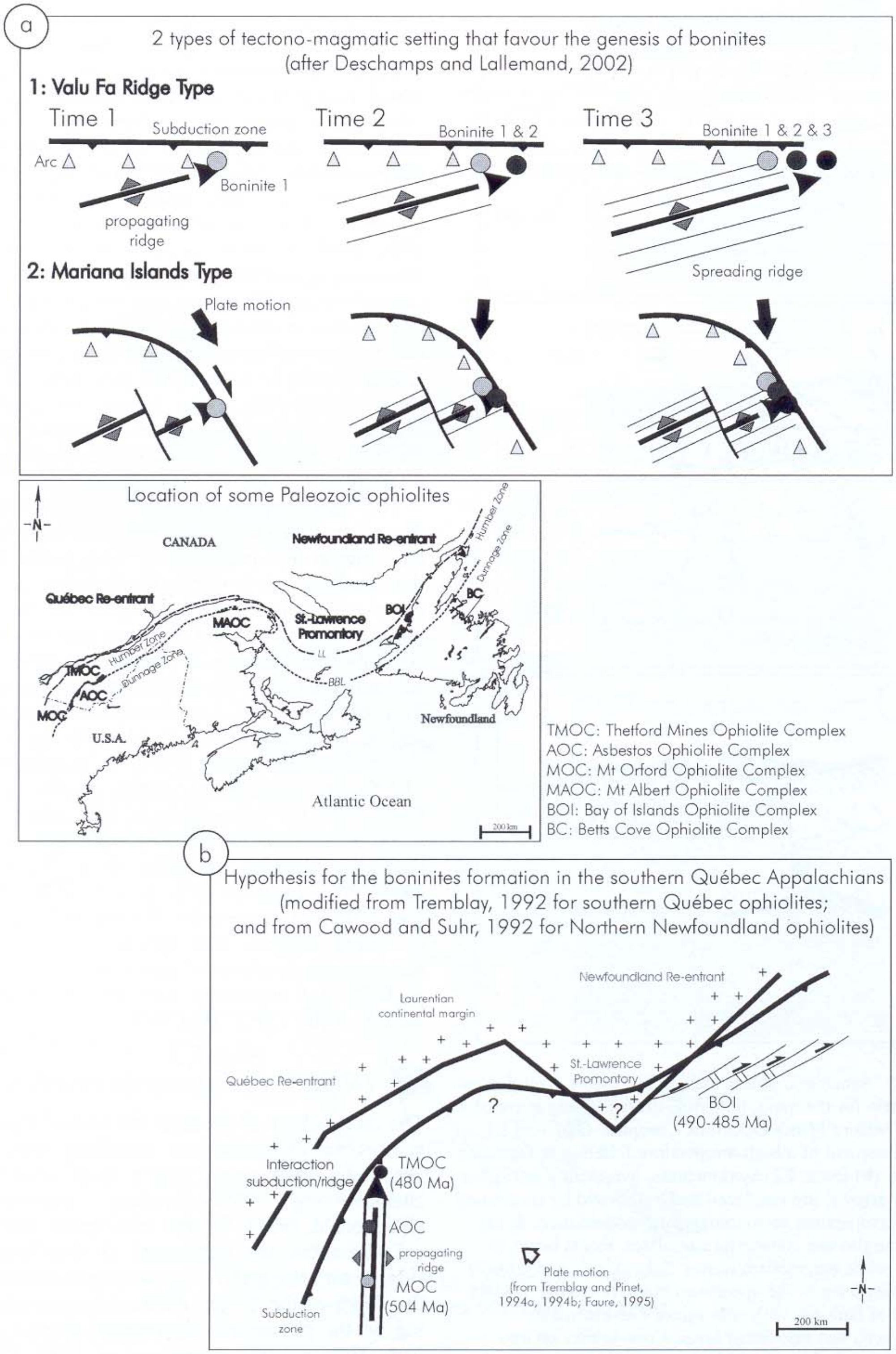

Fig. 9. (**a**) Schematic illustrations of a possible configuration for the genesis of boninites (after Deschamps & Lallemand 2003). (**b**) Model of boninite formation for southern Québec Appalachian ophiolites.

level to which synmagmatic extensional faults penetrate (Figs 1 and 3), and the evidence for near-pervasive lower-crustal hydrothermal metamorphism are all features compatible with a slow spreading rate.

An unresolved problem is the nature and location of the accommodation zone (Harper 1985) for the normal faults we have documented (Figs 1 and 3). We have been able to follow the faults down into the Dunitic Zone, but because of the poor exposure and common post-emplacement reactivation of Dunitic Zone rocks, we cannot be certain whether rotation of upper-crustal fault blocks occurred along a décollement zone, whether it penetrated down into the mantle, or whether the deformation was accommodated by a wide, ductile, lower crust. The tectonic exhumation of lower-crustal and mantle rocks along shallowly dipping extensional faults seems to characterize slow-spreading environments (e.g. Dick *et al.* 1992; Dilek *et al.* 1998). In the TMOC, the deposition of lavas and sediments directly upon mantle or lower-crustal rocks (Fig. 1) could perhaps be explained in this manner.

Formation of boninitic crust in the northern Appalachians

The formation of boninite lavas requires a hot, depleted mantle affected by subduction zone metasomatism (Crawford *et al.* 1989). Several tectonic environments have been suggested to explain boninite genesis, including: subduction of a spreading ridge; plume–subduction zone interactions; subduction zone initiation; propagation of a backarc spreading centre into a subduction zone; slab rollback caused by a decrease in convergence rate (e.g. Crawford *et al.* 1989; Stern & Bloomer 1992; Bédard *et al.* 1998; Macpherson & Hall 2001; Kim & Jacobi 2002; Deschamps & Lallemand 2003). The Taconian ophiolites of the northern Appalachians are unusual in that boninites are extremely abundant, may constitute the dominant magmatic suite over extensive regions and are roughly coeval (490–480 Ma, e.g. Church 1977; Bédard *et al.* 2000; Bédard & Kim 2002; Kim & Jacobi 2002). Any genetic model must explain the development of extensive boninitic magmatism at this juncture.

The identification of sub-north–south extensional oceanic structures in the TMOC suggests that the spreading centre had a north–south trend and that extension and sea-floor spreading took place along an east–west flow direction. North–south lineaments have also been reported from other oceanic terranes in southern Québec and northern New England (Doolan *et al.* 1982; Tremblay & Malo 1991; Tremblay 1992), suggesting that this may be the orientation of the extensional (spreading) axis at this time. Ordovician palaeogeographical reconstructions (Tremblay 1992) suggest a SE-dipping subduction zone, which would produce a diachronous intersection between spreading centre and subduction zone (Fig. 9), a configuration compatible with the genetic model proposed for boninites by Deschamps & Lallemand (2003). However, this assumes that the north–south trend documented for the TMOC spreading axis has not been affected by rotation of the obducted oceanic terrane, as has been documented in other ophiolites (e.g. Perrin *et al.* 1993). Another objection to this model is that it requires a very stable geometry to account for development of similar, roughly coeval, boninitic oceanic crust along >1500 km of strike length in the northern Appalachians. In contrast, models involving collision of the Taconic arc against an irregular continental margin and coeval formation of forearc spreading centres in re-entrants (Fig. 9: Cawood & Suhr 1992; Bédard & Kim 2002; Kim & Jacobi 2002) may be more compatible with the apparently synchronous development of anomalous boninitic crust of Betts Cove type (Church 1977) over such an extensive region.

Conclusions

Our data highlight the links that exist between the structural and magmatic history of the Thetford Mines Ophiolitic Complex (TMOC; Fig. 7). We have documented the manner in which synmagmatic normal faults dissected the upper crust into tilted fault blocks and controlled deposition of lavas and sediments. These observations imply that the ophiolite was partially dismembered by extensional tectonics before its emplacement onto the continental margin (Fig. 7a). The dominance of a boninitic signature in cumulates and lavas suggests that the TMOC is a forearc-related ophiolite. The evidence for coeval extension and magmatism, and the discovery of a locally well-developed sheeted dyke complex suggest that the TMOC formed when sea-floor spreading was initiated in a forearc, with slow magmatic extension being fed by boninitic, rather than tholeiitic magmas. Emplacement onto the continental margin followed very quickly after the ophiolite formed, and a piggy-back forearc basin was developed.

Thanks are due to R. Hébert, R. Laurent and the late P. St-Julien for introducing us to the area and sharing their knowledge during earlier field seasons; to B. Brassard and Ressources Allican Inc. for initiating the project; to N. Pinet, Y. Hébert and F. Huot for numerous discussions; to P. Cousineau and the late lamented G. Kessler for volcanological and sedimentological insights; and to Y. Dilek and P. Robinson for insightful and lightning-fast reviews. This project has been supported by the Geological Survey of Canada, by Canadian National Science and Engineering Research Council grant ESS 233685-99, by a Diversification de l'Exploration Minérale au Québec grant provided by Valorisation-Recherche Québec (project 2201-133), and by Ressources Allican Inc. This is Geological Survey of Canada Contribution 2002206.

References

BÉDARD, J.H. 1991. Cumulate recycling and crustal evolution in the Bay of Islands ophiolite. *Journal of Geology*, **99**, 225–249.

BÉDARD, J.H. 1993. The oceanic crust as a reactive filter: multiple syn-kinematic intrusion, hybridization and assimilation in an ophiolitic magma chamber. *Geology*, **21**, 77–80.

BÉDARD, J.H. & KIM, J. 2002. Boninites in NE Appalachian Notre-Dame Subzone ophiolites: tectonic models for Iapetus terranes. *Geological Society of America, Abstracts with Programs*, **34**, A60.

BÉDARD, J.H., LAUZIÈRE, K., TREMBLAY, A. & SANGSTER, A. 1998. Evidence for forearc seafloor-spreading from the Betts Cove ophiolite, Newfoundland: oceanic crust of boninitic affinity. *Tectonophysics*, **284**, 233–245.

BÉDARD, J.H., LAUZIÈRE, K., TREMBLAY, A., SANGSTER, A., DOUMA, S.L. & DEC, T. 2000. *Betts Cove Ophiolite and its Cover Rocks, Newfoundland*. Geological Survey of Canada Bulletin, **550**.

BÉDARD, J.H., TREMBLAY, A., SCHROETTER, J.-M., BÉCU, V. & PAGÉ, P. 2001. Structural and magmatic evolution of the Thetford Mines Ophiolite: preliminary results from new mapping and analysis. *Geological Association of Canada & Mineral Association of Canada Joint Annual Meeting. St-John's, NF, Canada*. Program with Abstracts, **26**, 11.

BERCLAZ, A., BÉDARD, J.H. & HÉBERT, R. 1998. Importance and significance of deformation processes in the genesis of oceanic crust in a slow-spreading environment: evidences from the Bay of Islands Ophiolite, Newfoundland. *Geological Association of Canada & Mineralogical Association of Canada Joint Annual Meeting, Québec, QC, Canada*. Program with Abstracts, **23**, A-15.

BEULAC, R.X. 1982. *Étude pétrologique du Complexe ophiolitique du lac Nicolet*. MSc thesis, Université Laval, Québec.

BLOOMER, S.H., TAYLOR, B., MACLEOD, C.J., STERN, R.J., FRYER, P., HAWKINS, J.W. & JOHNSON, L. 1995. Early arc volcanism and the ophiolite problem: a perspective from drilling in the western Pacific. *In:* TAYLOR, B. & NATLAND, J. (eds) *Active Margins and Marginal Basins of the Western Pacific*. Geophysical Monograph, American Geophysical Union, **88**, 1–30.

BRASSARD, B. & TREMBLAY, A., 1999. *Synthèse géologique et métallogénique de la M.R.C. de l'Amiante*. Rapport final préparé pour la société d'aide au développement de la collectivité de l'Amiante (SADC), 1/50 000 scale map, northern and southern sheets of the Amiante MRC.

CANN, J.R. 1974. A model for ocean crustal structure developed. *Geophysical Journal of the Royal Astronomical Society*, **39**, 169–187.

CASEY, J.F., KARSON, J.A., ELTON, D.L., ROSENCRANTZ, E. & TITUS, M. 1983. Reconstruction of the geometry of accretion during formation of the Bay of Islands Ophiolite Complex. *Tectonics*, **2**, 509–528.

CASTONGUAY, S., RUFFET, G., TREMBLAY, A. & FÉRAUD, G. 2001. Tectonometamorphic evolution of the southern Quebec Appalachians: $^{40}Ar/^{39}Ar$ evidence for Middle Ordovician crustal thickening and Silurian–Early Devonian exhumation of the internal Humber Zone. *Geological Society of America Bulletin*, **113**, 144–160.

CAWOOD, P.A. & SUHR, G. 1992. Generation and obduction of ophiolites: constraints from the Bay of Islands Complex, western Newfoundland. *Tectonics*, **11**(4), 884–897.

CHURCH, W.R. 1977. The ophiolites of southern Quebec: oceanic crust of Betts Cove type. *Canadian Journal of Earth Sciences*, **14**, 1668–1673.

CHURCH, W.R. 1978. The ophiolites of southern Quebec: oceanic crust of Betts Cove type: Reply. *Canadian Journal of Earth Sciences*, **15**, 1882–1883.

CHURCH, W.R. 1987. The geochemistry and petrogenesis of ophiolitic volcanic rocks from Lac de l'Est, Thetford Mines Complex, Quebec, Canada: Discussion. *Canadian Journal of Earth Sciences*, **24**, 1270–1273.

CLAGUE, D., FRANKEL, C.S. & EABY, J.S. 1985. The age and origin of felsic intrusions on the Thetford Mines ophiolite, Québec. *Canadian Journal of Earth Sciences*, **22**, 1257–1261.

COLEMAN, R.G. 1971. Plate tectonic emplacement of upper mantle peridotites along continental edges. *Journal of Geophysical Research*, **76**, 1212–1222.

COOKE, M.H.-C. 1938. *Région de Thetford, de Disraëli et de la Moitié orientale de Warwick (Québec)*. Commission Géologique du Canada, Mémoire, **211**, 176.

COUSINEAU, P.A. & ST-JULIEN, P. 1992. The Saint-Daniel Mélange: evolution of an accretionary complex in the Dunnage terrane of the Québec Appalachians. *Tectonics*, **11**, 898–909.

CRAWFORD, A.J., FALLOON, T.J. & GREEN, D.H. 1989. Classification, petrogenesis and tectonic setting of boninites. *In:* CRAWFORD, A.J. (ed.) *Boninites and Related Rocks*. Unwin Hyman, London, 1–49.

DAVID, J. & MARQUIS, R. 1994. Géochronologie U–Pb dans les Appalaches du Québec: application aux roches de la zone de Dunnage. *Revue Géologique du Québec*, **1**, 16–20.

DÉROSIER, C. 1971. *Étude géologique des brèches de la région de Thetford Mines*. PhD thesis, Université de Paris.

DESCHAMPS, A. & LALLEMAND, S. 2003. Geodynamic setting of Izu–Bonin–Mariana boninites. *In:* LAR-

TER, R.D. & LEAT, P.T. (eds) *Intra-Oceanic Subduction Systems: Tectonic and Magmatic Processes*. Geological Society of London, Special Publications, **219**, 163–185.

DICK, H.J.B., ROBINSON, P.T. & MEYER, P.S. 1992. The plutonic foundation of a slow-spreading ridge. *In:* DUNCAN, R.A. ET AL. (eds) *The Indian Ocean: A Synthesis of results from the Ocean Drilling Program*. Geophysical Monograph, American Geophysical Union, **70**, 1–92.

DICKINSON, W.R. & SEELY, D.R. 1979. Structure and stratigraphy of forearc regions. *AAPG Bulletin*, **63**, 2–31.

DICKINSON, W.R., KLUTE, M.A., HAYES, M.J., JANECKE, S.U., LUNDIN, E.R., MCKITTRICK, M.A. & OLIVARES, M.D. 1988. Paleogeographic and paleotectonic setting of Laramide sedimentary basins in the central Rocky Mountain region. *Geological Society of America Bulletin*, **100**, 1023–1039.

DILEK, Y. & EDDY, C.A. 1992. The Troodos (Cyprus) and Kizildag (S. Turkey) ophiolites as structural models for slow-spreading ridge segments. *Journal of Geology*, **100**, 305–322.

DILEK, Y., MOORES, E.M. & FURNES, H. 1998. Structure of modern oceanic crust and ophiolites and implications for faulting and magmatism at oceanic spreading centers. *In:* BUCK, R., KARSON, J., DELANEY, P. & LAGABRIELLE, Y. (eds) *Faulting and Magmatism at Mid-Ocean Ridges*. American Geophysical Union, Geophysical Monograph, **106**, 219–266.

DOOLAN, L.D., GALE, M.H., GALE, P. & HOAR, R. 1982. Geology of the Québec Re-entrant: possible constraints from early rifts and the Vermont–Québec serpentine belt. *In:* ST-JULIEN, P. & BÉLAND, J. (eds) *Major Structural Zones and Faults of the Northern Appalachians*. Geological Association of Canada, Special Papers, **24**, 87–115.

DUNNING, G.R., KROGH, T.E. & PEDERSON, R.B. 1986. U–Pb zircon ages of Appalachian–Caledonian ophiolites. *Terra Cognita*, **6**, L51.

FEININGER, T. 1981. Amphibolite associated with the Thetford Mines Ophiolite Complex at Belmina Ridge, Québec. *Canadian Journal of Earth Sciences*, **18**, 1878–1892.

HAMILTON, W.B. 1988. Plate tectonics and island arcs. *Geological Society of America Bulletin*, **100**, 1503–1527.

HAMILTON, W.B. 1995. Subduction systems and magmatism. *In:* SMELLIE, J.L. (ed.) *Volcanism Associated with Extension at Consuming Plate Margins*. Geological Society, London, Special Publications, **81**, 3–28.

HARNOIS, L. & MORENCY, M. 1989. Geochemistry of Mount Orford Ophiolite Complex, Northern Appalachians, Canada. *Chemical Geology*, **77**, 133–147.

HARPER, G.D. 1985. Tectonics of slow spreading mid-ocean ridges and consequences of a variable depth to the brittle/ductile transition. *Tectonics*, **4**, 395–409.

HÉBERT, R. 1980. *Étude pétrologique des roches d'Asbestos et du Mont Ham (Ham Sud), Québec*. MSc thesis, Université Laval, Québec.

HÉBERT, R. 1981. Conglomérats polygéniques ophiolitiques: anciens éboulis de talus de fond océanique. *Canadian Journal of Earth Sciences*, **18**, 619–623.

HÉBERT, R. 1985. *Pétrologie des roches ignées océaniques et comparaison avec les complexes ophiolitiques du Québec, de Chypre et de l'Apennin*. PhD thesis, Université de Bretagne occidentale, Brest, France.

HÉBERT, R. & BÉDARD, J.H. 2000. Les ophiolites d'avant-arc et leur potentiel minéral: exemple des complexes ophiolitiques du sud du Québec. *Chronique de la Recherche Minière*, **539**, 101–117.

HÉBERT, R. & LAURENT, R. 1989. Mineral chemistry of ultramafic and mafic plutonic rocks of the Appalachian ophiolites, Québec, Canada. *Chemical Geology*, **77**, 265–285.

HÉBERT, Y. 1983. *Étude pétrologique du Complexe Ophiolitique de Thetford Mines, Québec*. PhD thesis, Université Laval, Québec.

HUOT, F., HÉBERT, R. & TURCOTTE, B. 2002. A multistage magmatic history for the genesis of the Orford ophiolite (Quebec, Canada): a study of the Mt Chagnon massif. *Canadian Journal of Earth Sciences*, **39**, 1201–1217.

KACIRA, N. 1971. *Geology of chromitite occurrences and ultramafic rocks of the Thetford Mines-Disraëli area, Québec*. PhD thesis, University of Western Ontario, London.

KARSON, J.A., ELTHON, D. & DELONG, S.E. 1983. Ultramafic intrusions in the Lewis Hills Massif, Bay of Islands Ophiolite Complex, Newfoundland: implications for the igneous processes at oceanic fracture zones. *Geological Society of American Bulletin*, **94**, 15–29.

KIM, J. & JACOBI, R.D. 2002. Boninites: characteristics and tectonic constraints, northeastern Appalachians. *Physics and Chemistry of the Earth*, **27**, 109–147.

LAURENT, R. 1975. Occurrences and origin of the ophiolites of the Southern Quebec, Northern Appalachians. *Canadian Journal of Earth Sciences*, **12**, 443–455.

LAURENT, R. 1978. The ophiolites of southern Quebec: oceanic crust of Betts Cove type: Discussion. *Canadian Journal of Earth Sciences*, **15**, 1880–1881.

LAURENT, R. & HÉBERT, Y. 1977. Features of submarine volcanism in ophiolites from the Québec Appalachians. *In:* BARAGAR, W.R. (ed.) *Volcanic Regimes in Canada*. Geological Association of Canada, Special Papers, **16**, 91–109.

LAURENT, R. & HÉBERT, R. 1989. The volcanic and intrusive rocks of the Québec Appalachian ophiolites (Canada) and their island-arc setting. *Chemical Geology*, **77**, 287–302.

LAURENT, R., HÉBERT, R. & HÉBERT, Y. 1979. Tectonic setting and petrological features of the Québec Appalachian ophiolites. *In:* MALPAS, J. & TALKINGTON, R.W. (eds) *Ophiolites of the Canadian Appalachians and Soviet Urals*. International Geological Correlation Project, **39**, 53–77.

LYTWYN, J.N. & CASEY, J.F. 1993. The geochemistry and petrogenesis of volcanics and sheeted dikes from the Hatay (Kizildag) ophiolite, southern Turkey: possible formation with the Troodos ophiolite, Cyprus, along fore-arc spreading centers. *Tectonophysics*, **223**, 237–272.

MACDONALD, K.C. 1982. Mid-ocean ridges: fine scale tectonic, volcanic and hydrothermal processes within the plate boundary zone. *Annual Review of Earth and Planetary Sciences*, **10**, 155–190.

MACPHERSON, C.G. & HALL, R. 2001. Tectonic setting of Eocene boninite magmatism in the Izu–Bonin–Mariana forearc. *Earth and Planetary Science Letters*, **186**, 215–230.

NICOLAS, A. 1992. Kinematics in magmatic rocks with special reference to gabbros. *Journal of Petrology*, **33**, 891–915.

OSBERG, P.H. 1978. Synthesis of the geology of northeastern Appalachians. *In:* TOZER, E.T. & SCHENK, P.E. (eds) *Caledonian–Appalachian Orogen of the North Atlantic Region*. Geological Survey of Canada, Papers, **78-13**, 137–147.

OSHIN, I.O. & CROCKET, J.H. 1986. The geochemistry and petrogenesis of ophiolitic volcanic rocks from Lac de l'Est, Thetford Mines Complex, Quebec, Canada. *Canadian Journal of Earth Sciences*, **23**, 202–213.

PAGÉ, P., BÉCU, V., TREMBLAY, A., BÉDARD, J.H. & BRASSARD, B. 2001. Occurrences of platinum–palladium-enriched chromitites in the Thetford Mines ophiolite, southern Québec. *Geological Society of America, Northeastern Section, Abstracts with Programs*, **33**(1), 58–59.

PEARCE, J.A., LIPPARD, S.J. & ROBERTS, S. 1984. Characteristics and tectonic significance of suprasubduction ophiolites. *In:* KOKELAAR, B.P. & HOWELLS, M.F. (eds) *Marginal Basin Geology*. Geological Society, London, Special Publications, **16**, 77–94.

PERRIN, M., PRÉVOT, M. & BRUERE, F. 1993. Rotation of Oman ophiolite and initial location of the ridge in the hotspot reference frame. *Tectonophysics*, **229**, 31–42.

PINET, N. 1995. *Cartographie du secteur Lac Bisby–Petit Lac Saint-François, Propriété Coleraine*. QERDEM B Dossier d'exploration minière, Ressources Minières Coleraine Inc., **GM 53715**, 33.

PINET, N. & TREMBLAY, A. 1995. Tectonic evolution of the Québec–Maine Appalachians: from oceanic spreading to obduction and collision in the north Appalachians. *American Journal of Science*, **295**, 173–200.

PINET, N., CASTONGUAY, S. & TREMBLAY, A. 1996a. Thrusting and back thrusting in the Taconian internal zone, southern Quebec Appalachians. *Canadian Journal of Earth Sciences*, **33**, 1283–1293.

PINET, N., TREMBLAY, A. & SOSSON, M. 1996b. Extension versus shortening models for hinterland-directed motions in the southern Québec Appalachians. *Tectonophysics*, **267**, 239–256.

RABINOWICZ, M., ROUZO, S., SEMPERE, J.C. & ROSEMBERG, C. 1993. Three-dimensional mantle flow beneath mid-ocean ridges. *Journal of Geophysical Research*, **98**, 7851–7869.

ROBINSON, P.T. & MALPAS, J. 1990. The Troodos ophiolite of Cyprus: new perspectives on its origin and emplacement. *In:* MALPAS, J., MOORES, E.M., PANAYIOTOU, A. & XENOPHONTOS, C. (eds) *Ophiolites, Oceanic Crust Analogues, Proceeedings of Symposium 'TROODOS 1987'*. Geological Survey Department, Nicosia, 13–26.

ROBINSON, P.T., MELSON, W.G., O'HEARN, T. & SCHMINCKE, H.-U. 1983. Volcanic glass compositions of Troodos ophiolite, Cyprus. *Geology*, **11**, 400–404.

ROBINSON, P., TUCKER, R.D., BRADLEY, D., BERRY, H.N. & OSBERG, P.H. 1998. Paleozoic orogens in New England. *Geologiska Föreningens i Stockholm Förhandlingar*, **120**, 119–148.

ST-JULIEN, P. 1987. *Géologie des Régions de Saint-Victor et de Thetford-Mines (moitié est)*. Ministère de l'Énergie et des Ressources du Québec, Québec, **MM86-01**.

SCHROETTER, J.-M., BÉCU, V., PAGÉ, P., TREMBLAY, A. & BÉDARD, J.H. 2000. *Chromitites Ophiolitiques de la Région de Thetford MinesRapport des Activités de Terrain de la Saison Été*. Ministère de l'Énergie et des Ressources du Québec, Québec, **GM 58649**.

SCHROETTER, J.-M., PAGÉ, P., TREMBLAY, A. & BÉDARD, J.H. 2002. Structural evolution of the Thetford-Mines Ophiolitic Complex, Québec: from synoceanic rifting to obduction and post-obduction deformation. *Geological Society of America, Northeastern Section, Abstracts with Programs*, **34**(1), A21.

STERN, R.J. & BLOOMER, S.H. 1992. Subduction zone infancy: examples from the Eocene Izu–Bonin–Mariana and Jurassic California arcs. *Geological Society of America Bulletin*, **104**, 1621–1636.

SUHR, G. & CAWOOD, P.A. 2001. Southeastern Lewis Hills (Bay of Islands Ophiolite): geology of a deeply eroded, inside-corner, ridge–transform intersection. *Geological Society of America Bulletin*, **113**, 1025–1038.

TAPPONIER, P. & FRANCHETEAU, J. 1978. Necking of the lithosphere and the mechanics of slowly accreting plate boundaries. *Journal of Geophysical Research*, **83**, 3955–3970.

THY, P. 1990. Cryptic variation of a cumulate sequence from the plutonic complex of the Troodos ophiolite. *In:* MALPAS, J., MOORES, E.M., PANAYIOTOU, A. & XENOPHONTES, C. (eds) *Ophiolites, Ocean Crustal Analogues, Proceedings of the Symposium 'TROODOS 1987'*. Publisher, Nicosia, 165–172.

TOOMEY, D.R., SOLOMON, S.C. & PURDY, G.M. 1988. Microearthquakes beneath the median valley of the Mid-Atlantic Ridge near 23°N: tomography and tectonics. *Journal of Geophysical Research*, **93**, 9093–9112.

TREMBLAY, A. 1992. Tectonic and accretionary history of Taconian oceanic rocks of the Québec Appalachians. *American Journal of Science*, **292**, 229–252.

TREMBLAY, A. & CASTONGUAY, S. 2002. The structural evolution of the Laurentian margin revisited (southern Quebec): implications for the Salinian Orogeny and Appalachian successor basins. *Geology*, **30**(1), 7–82.

TREMBLAY, A. & MALO, M. 1991. Significance of brittle and plastic fabrics within the Masawippi Lake fault zone, southern Canadian Appalachians. *Journal of Structural Geology*, **13**(9), 1013–1023.

TREMBLAY, A., BÉDARD, J.H. & LAUZIÈRE, C. 1997. Taconian obduction and Silurian exhumation of the

Betts Cove ophiolite, Canadian Appalachians. *Journal of Geology*, **105**, 701–716.

UYEDA, S. & KANAMORI, H. 1979. Back-arc opening and the mode of subduction. *Journal of Geophysical Research*, **84**, 1049–1061.

VAN STAAL, C.R., DEWEY, J.F., MACNIOCAILL, C. & MCKERROW, W.S. 1998. The tectonic evolution of the Northern Appalachians and British Caledonides: history of a complex, west and southwest Pacific-type segment of Iapetus. *In:* BLUNDELL, D.J. & SCOTT, A.C. (eds) *Lyell: the Past is the Key to the Present*. Geological Society, London, Special Publications, **143**, 199–242.

WHITEHEAD, J., REYNOLDS, P.H. & SPRAY, J.G. 1995. The sub-ophiolitic metamorphic rocks of the Québec Appalachians. *Journal of Geodynamics*, **19**, 325–350.

WHITEHEAD, J., DUNNING, G.R. & SPRAY, J.G. 2000. U–Pb geochronology and origin of granitoid rocks in the Thetford Mines ophiolite, Canadian Appalachians. *Geological Society of America Bulletin*, **112**, 915–928.

WILLIAMS, H. 1979. Appalachian Orogen in Canada. *Canadian Journal of Earth Sciences*, **16**, 792–807.

WILLIAMS, H. & HATCHER, R.D. JR 1983. Appalachian suspect terranes. *In:* HATCHER, R.D. JR, WILLIAMS, H. & ZIETZ, I. (eds) *Geophysics of Mountain Chains*. Geological Society of America, Memoirs, **158**, 33–53.

WILLIAMS, H. & ST-JULIEN, P. 1982. The Baie Verte–Brompton Line: Early Paleozoic continent–ocean interface in the Canadian Appalachians. *In:* ST-JULIEN, P. & BÉLAND, J. (eds) *Major Structural Zones and Faults of the Northern Appalachians*. Geological Association of Canada, Special Papers, **24**, 177–207.