Pergamon

Journal of Structural Geology, Vol. 17, No. 4, pp. 475 to 491, 1995

0191-8141/95 $9.50+0.00

0191-8141(94)00076-X

Syn-tectonic magmatism and the development of compositional layering, Ungava Orogen (northern Quebec, Canada)

STEPHEN B. LUCAS and MARC R. ST-ONGE

Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario, Canada K1A 0E8

(*Received* 26 *July* 1993; *accepted in revised form* 14 *June* 1994)

Abstract—Layer- and foliation-parallel emplacement of granitic veins was an important process in the regional development of compositional layering in now-exhumed middle crustal sections of both Archean and Paleoproterozoic age in the northern Ungava peninsula, Quebec (Canada). In the Archean Superior Province, diorite and tonalite plutons were penetratively deformed and metamorphosed at granulite-facies conditions coeval with voluminous granitic magmatism. The Paleoproterozoic Narsajuaq arc contains evidence for contemporaneous magmatism, transpressional deformation and granulite-facies metamorphism prior to its collision with the Superior Province basement. In both plutonic domains, regional compositional layering is defined by (1) metre to kilometre-scale alternation of generally well foliated tonalite and quartz diorite bodies; and (2) centimetre to kilometre-scale, variably deformed granitic veins and sheets that lie parallel to layering/tectonic foliation in the host rocks. Syn-tectonic intrusion of a substantial portion of the veins along extension fractures sub-parallel to layering/foliation (i.e. at high angle to the regional shortening direction) is interpreted to have occurred due to the combination of a strong anisotropy and high magma pressures. Compositional layering generated and/or enhanced by this process may contribute to the overall seismic reflectivity of the middle and lower crust.

INTRODUCTION

Compositional layering is common in metamorphic rocks (Talbot & Hobbs 1968, Williams 1972, 1990, Meyers 1978) and is often defined or accentuated by layer-parallel veins and/or segregations. Such veins or segregations commonly comprise quartz, carbonate and/or feldspars, and are often inferred to be contemporaneous with deformation (e.g. Hanmer 1988, Hudleston 1989, Hanmer & McEachern 1992). Compositional layering in metamorphic rocks is variably described as segregated, differentiated, gneissic, *lit-par-lit* or migmatitic layering, and is generally parallel to grain-scale foliations within individual layers (e.g. Sawyer & Robin 1986, Williams 1990). Study of exposed sections of middle to lower continental crust has indicated that the crust is compositionally layered at a variety of scales, and has underlined the importance of plutonic rocks in defining or enhancing that layering (Fountain & Salisbury 1981, Davidson 1984, Zingg *et al.* 1990). However, the origin of compositional layering in metamorphic rocks has long troubled geologists (Eskola 1932, Turner 1941, Misch 1949, Ramberg 1952, Shelley 1974, Robin 1979, Williams 1990). Both natural and experimental observations and theoretical considerations have shown that externally and internally-controlled variables can play important roles in its development (see Misch 1968, Yardley 1978, Olsen 1983, Sawyer 1994).

In situ mechanisms for generating layering in metamorphic rocks include subsolidus, stress-induced diffusive mass transfer (metamorphic segregation or differentiation, Robin 1979, Sawyer & Robin 1986, Gratier 1987, Williams 1990), anatexis or partial melting (Johannes & Gupta 1982, Johannes 1988, Sawyer 1994), and mechanical processes such as transposition (Myers 1978, Passchier *et al.* 1990) and differentiation of minerals of differing ductility (Shelley 1974, Jordan 1987). Other, *externally-driven* mechanisms include *lit-par-lit* injection of granitic magmas, long considered as a possible migmatization mechanism (Misch 1949, 1968, Olsen 1983), and fluid influx or metasomatism, which may facilitate stress-induced diffusive mass transfer or partial melting (Misch 1968, Yardley 1978, Olsen 1983). In this paper we investigate the importance of *externally-introduced* melts in contributing to the development of compositional layering in both Archean and Proterozoic high grade metamorphic rocks from the northern Ungava peninsula, Quebec (Fig. 1; Lucas & St-Onge 1992, St-Onge *et al.* 1992).

The geological observations from both Archean and Paleoproterozoic gneisses support a model for the origin of layer-parallel granitic veins in which they are emplaced along extension fractures (cf. Etheridge 1983, Clemens & Mawer 1992) sub-parallel to the tectonic foliation/layering during regional deformation. We believe that this results from the combination of a well-developed mechanical anisotropy and high melt pressures (see Gratier 1987, Wickham 1987, Holliger & Levander 1994a). If correct, our model for syn-tectonic layer-parallel veining has important implications for the interpretation of layered (gneissic) middle to lower crust, especially where it is identified indirectly from seismic reflection data (Smithson *et al.* 1986, Green *et al.* 1990). Recent two-dimensional stochastic modeling of the seismic response of lower crustal sections, such as the Strona-Ceneri and Ivrea zones (cf. Zingg *et al.* 1990), have shown that its reflective character can be accounted for in its scale-independent, felsic–mafic compositional layer-

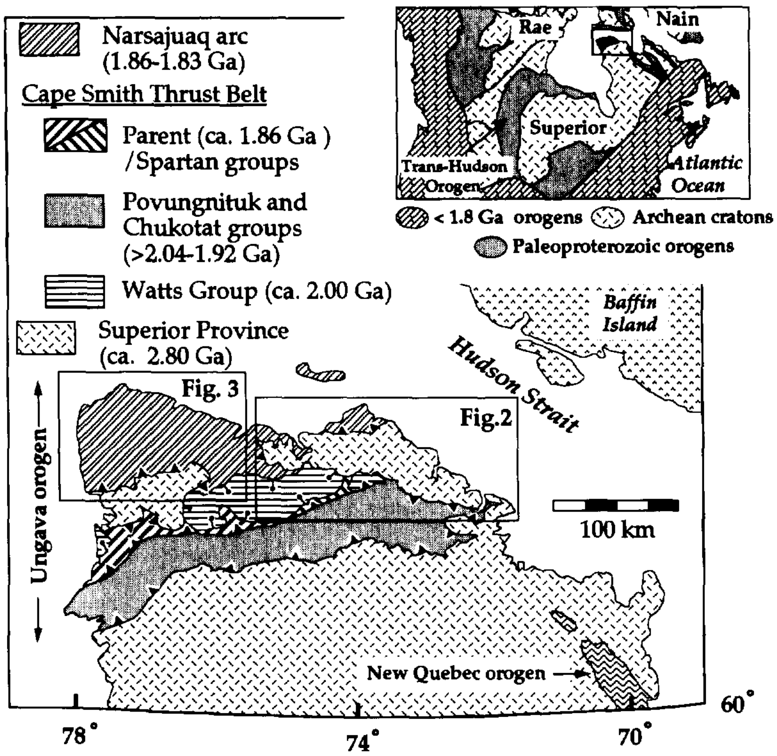

Fig. 1 Tectonic map of the Ungava orogen (after St-Onge *et al.* in press). Barbs along tectonic contacts represent major thrust faults; 'ticks with ball' represent a regionally-extensive normal fault. Locations of Figs. 2 and 3 indicated. Inset map of North America (after Hoffman 1989) shows location of the Ungava orogen (box) relative to the Trans-Hudson Orogen and other Paleoproterozoic orogens.

ing. This produces discontinuous, bimodal velocity–density distributions with shorter characteristic (fractal) scales relative to more compositionally homogeneous upper crust (Levander & Holliger 1992, Holliger & Levander 1994b).

REGIONAL GEOLOGY AND TECTONIC HISTORY

Three principal tectonostratigraphic domains are represented on the northern Ungava peninsula of Quebec (Lucas *et al.* 1992, St-Onge *et al.* 1992): (1) Archean Superior Province basement; (2) Paleoproterozoic divergent continental margin/ocean basin(s); and (3) a Paleoproterozoic convergent margin. The Archean Superior Province (Figs. 1 and 2) represents the stratigraphic basement to domain 2 and the structural basement to both domains 2 and 3. It is characterized principally by amphibolite to granulite-grade plutonic gneisses (2.7–2.9 Ga, Parrish 1989, personal communication 1992, St-Onge *et al.* 1992), interpreted as the mid-crustal core of a magmatic arc on the basis of pluton composition and tectonomagmatic history (St-Onge *et al.* 1992). The principal structural element in the Superior Province basement is a pervasive, steeply inclined layering defined by compositional banding at both map (Fig. 2) and outcrop-scales (see below). This layering developed during 2.9–2.7 Ga diorite–tonalite–granodiorite–granite plutonism and deformation, possibly in a magmatic arc setting (Lucas & St-Onge 1992, St-Onge *et al.* 1992). The Archean layering in the basement is truncated by an unconformity and/or a thrust fault at the base of the Paleoproterozoic cover sequence (Fig. 2).

The Paleoproterozoic divergent margin units include (1) a 2.00 Ga ophiolite (Watts Group, Fig. 1; Parrish 1989, Scott *et al.* 1991), (2) a 2.04–1.96 Ga continental rift sequence of sedimentary and volcanic rocks (Povungnituk Group, Fig. 1; Parrish 1989, St-Onge & Lucas 1990, St-Onge *et al.* 1992, Machado *et al.* 1993), and (3) a sequence of oceanic-like basalts inferred to be approximately 1.92 Ga (Chukotat Group, Fig. 1; Francis *et al.* 1983, St-Onge & Lucas 1990). The convergent margin units are bracketed between 1.90 and 1.83 Ga by U–Pb zircon geochronology (Parrish 1989, St-Onge *et al.* 1992, Machado *et al.* 1993). These units comprise supracrustal sequences inferred to represent arc and forearc deposits (Parent and Spartan groups, respectively; Fig. 1), and plutonic and metasedimentary rocks thought to be a mid-crustal segment of the Narsajuaq magmatic arc (Figs. 1 and 3; Lucas *et al.* 1992, St-Onge *et al.* 1992). This arc was built in part on Watts Group oceanic crust and in part on older continental crust (Dunphy & Ludden 1994).

The tectonic history of the Paleoproterozoic Ungava orogen has been recently described by Lucas & St-Onge (1992) and St-Onge *et al.* (1992), and only the relevant details will be reviewed here (see also Lucas 1989, 1990, St-Onge & Lucas 1990, 1991). The history is characterized by structural–metamorphic episodes that are inferred to both pre-date and post-date a collision (1.82–1.80 Ga) between the Narsajuaq arc terrane (upper-

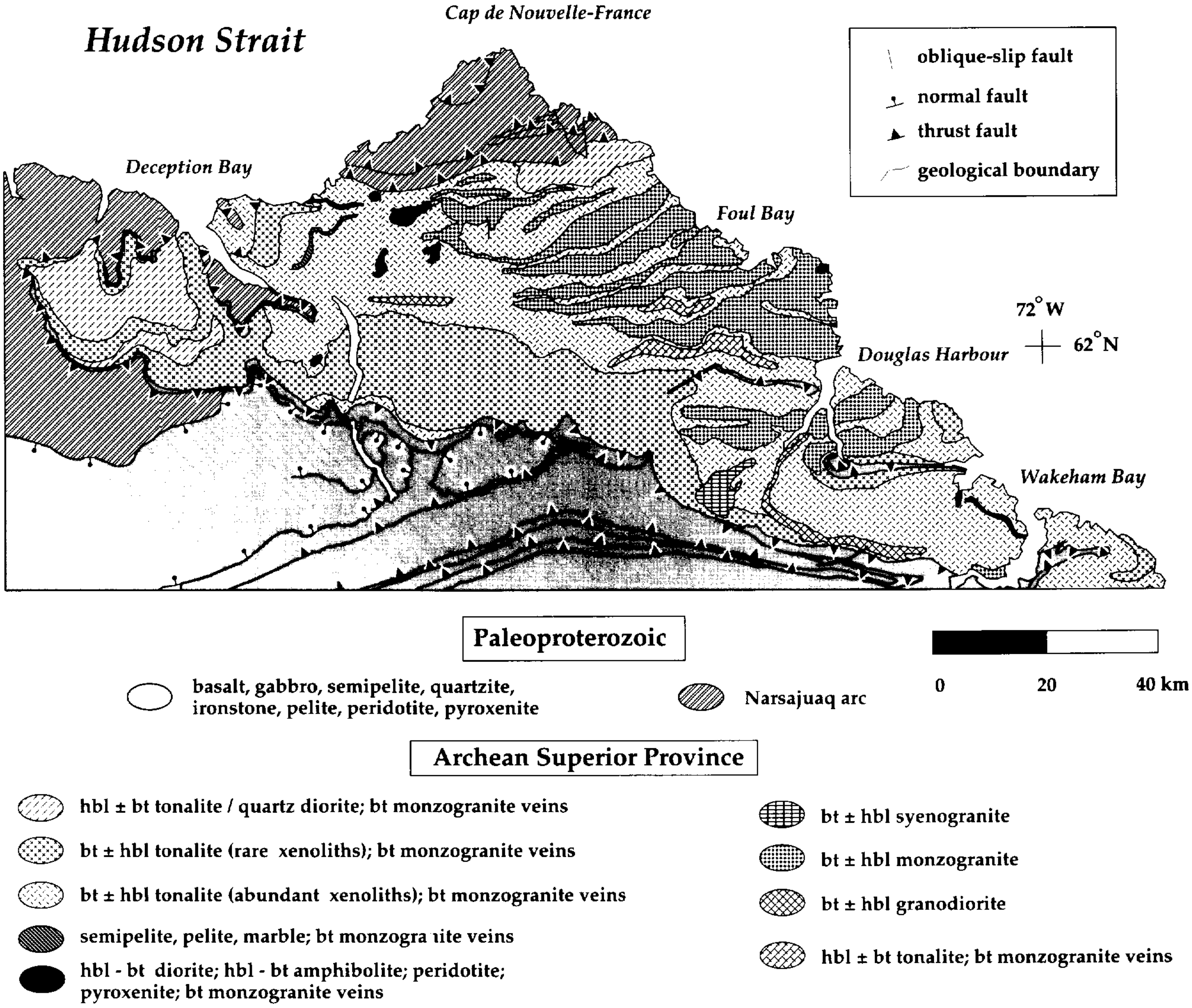

Fig. 2 Geological compilation map for the northeastern part of the Ungava orogen, highlighting the distribution of plutonic rocks in the Archean Superior Province basement.

plate) and the northern continental margin of the Superior Province (lower-plate). The lower-plate units preserved in the external part of the orogen (Cape Smith Belt) record the development of a thrust belt at *ca.* 1.87 Ga characterized by S-verging faults ramping up from a basal décollement located at the basement-cover contact (Lucas 1989). The Narsajuaq arc contains evidence for an intra-arc tectonic episode characterized by granulite-facies metamorphism and dextral transpression that was coeval with 1.86–1.83 Ga arc magmatism (see below).

The ophiolitic and arc units were accreted at *ca.* 1.82–1.80 Ga (St-Onge *et al.* 1992, R. Parrish personal communication 1992) onto the older thrust belt along S-verging faults that re-imbricated the external thrust belt. Thickening and consequent exhumation resulted in relatively high-P, greenschist–amphibolite-facies metamorphism of lower-plate cover units (St-Onge & Lucas 1991, Bégin 1992), and in the retrogression of high-T assemblages in Narsajuaq arc rocks (Lucas & St-Onge 1992). The regional gneissic foliation in the Archean basement was penetratively reworked immediately adjacent to the contact with Povungnituk Group metabasalts and metasedimentary rocks, resulting in the development of an amphibolite-grade mylonitic foliation and an extension lineation parallel to that in the cover rocks (Lucas 1989, 1990).

Continued shortening followed the cessation of thrusting and resulted in folding of both the allochthonous rocks and the footwall basement (Fig. 4; Lucas & Byrne 1992, Lucas & St-Onge 1992). Our geological mapping indicates that the Superior Province rocks exposed north of the Cape Smith Belt lie in the core of a major, regional-scale antiform ('antiformal axis', Fig. 4). Cross-folding of this structure has generated a regional fold-interference geometry in which the antiformal axis now defines a U-shape (strike-parallel section, Fig. 4). Along the antiformal axis, a depression in the central part of the orogen passes up-plunge into the two basement-cored culminations. Thus mid-crustal rocks of both the Superior Province basement and Paleoproterozoic Narsajuaq arc are exposed in oblique section in the Ungava orogen, due to the post-collisional folding episodes (Lucas 1989, Lucas & Byrne 1992).

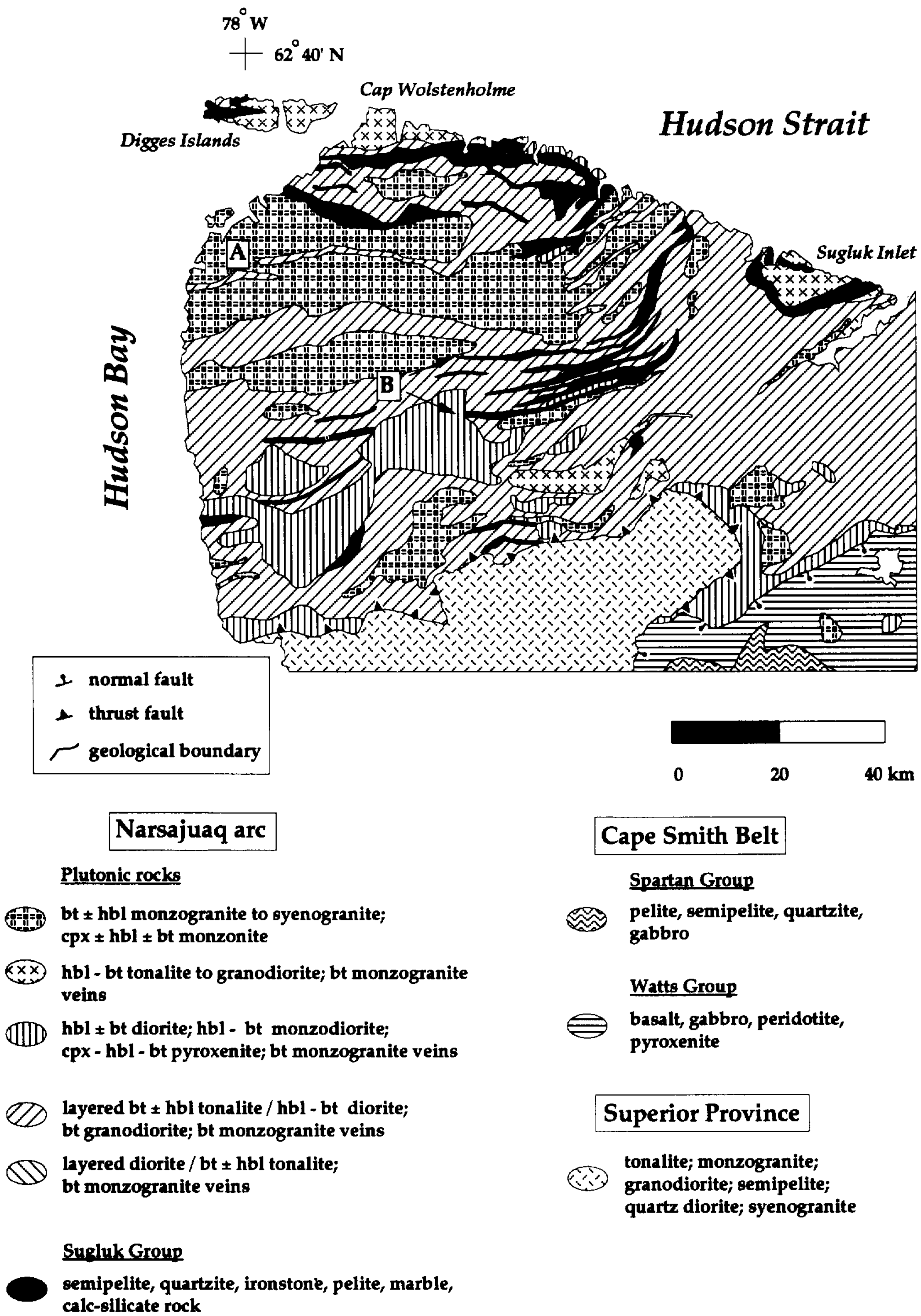

Fig. 3. Geological compilation map for the northwestern part of the Ungava orogen, highlighting the distribution of plutonic and metasedimentary rocks in the Paleoproterozoic Narsajuaq arc. 'A' (south of Digges Islands, near Hudson Bay coast) indicates major, W-trending screens of country rocks (older layered unit) in younger granite pluton(s). Elongate screens of country rock, paralleling regional foliation/layering, suggests that younger plutons were emplaced as sheet-like bodies along the pre-existing anisotropy. 'B' (southeast of 'A') indicates younger plutons that crosscut both outcrop- and regional-scale layering but are parallel to the regional foliation/layering trend.

NATURE OF THE COMPOSITIONAL LAYERING

Compositional layering is a ubiquitous feature of the mid-crustal, high-grade plutonic rocks of both the Archean Superior Province basement and the Paleoproterozoic Narsajuaq arc. This layering is related to two distinct tectonomagmatic events (2.9–2.7 Ga in the basement and 1.86–1.83 Ga in the Narsajuaq arc). As a significant proportion of the plutonic rocks in the basement and the Narsajuaq arc contain a tectonic foliation, the prefix 'meta' applies but will be dropped from the ensuing text for simplicity. In addition, we will employ the term 'granitic' to describe vein compositions within these two domains that span the compositional range

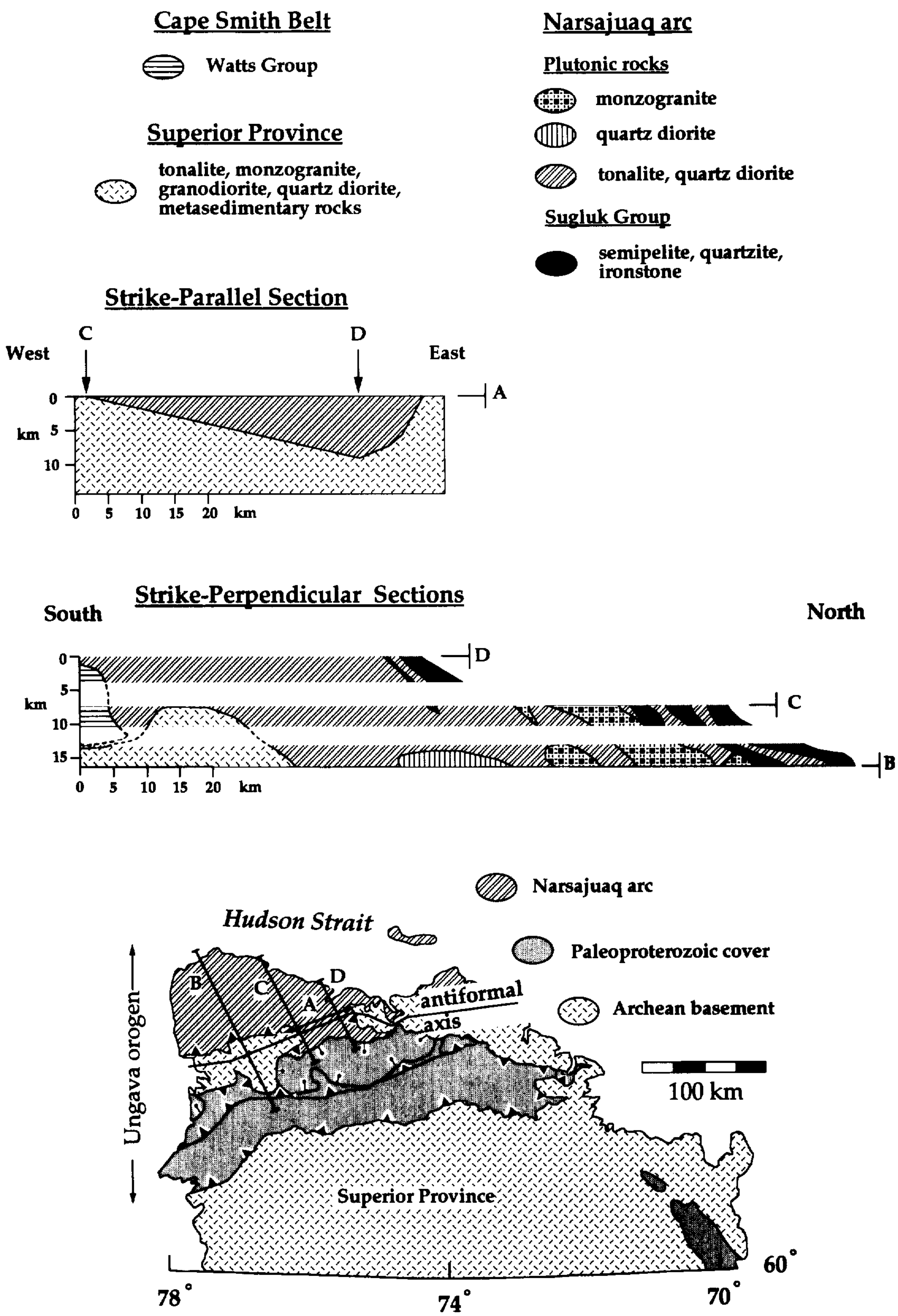

Fig. 4. Vertical, strike-parallel section (A) and down-plunge, strike-perpendicular sections (B–D) of the central and western parts of Narsajuaq arc; simplified geologic map shows the section lines. Section A derived by projecting the Archean basement–Narsajuaq arc contact down the plunge of the antiformal axis within a vertical plane parallel to the local axis trend. Sections B–D generated by projecting the Archean basement–Narsajuaq arc contact down the mean plunge of the antiformal axis onto a plane oriented orthogonal to that axis, assuming plunge continuity ('cylindricity') over the projection interval (see Lucas 1989). Internal structure of Narsajuaq arc on sections B–D based on simple extrapolation of surface dips along each section line on the projection plane.

from syenogranite and leucogranite to tonalite and quartz diorite, following Misch (1968).

Archean Superior Province

Tonalite, dated at 2780±4 Ma and 2882+44/−28 Ma (Parrish 1989, personal communication 1992), is the dominant plutonic rock type of the Superior Province basement (Fig. 2; St-Onge *et al.* 1992). It generally displays a well developed tectonic foliation (Figs. 5a & b), commonly defined by elongate quartz grains and polycrystalline feldspar aggregates (Fig. 5c) but also by inequidimensional, high-grade metamorphic minerals (e.g. orthopyroxene, clinopyroxene, hornblende). Metamorphic conditions, determined from two-pyroxene±garnet assemblages in diorite, cluster at approximately 800±50°C and 3.5±0.5 kbars (St-Onge *et al.* 1993). A minimum age for the granulite-facies metamorphism is 2740 Ma based on U–Pb geochronology of metamorphic zircons (R. Parrish personal communication 1992), whereas a maximum age is given by the age of the tonalites.

The tonalite is typically interlayered with granite and granodiorite veins (Figs. 5a & b). These veins are variably deformed (Figs. 5b & d), locally contain granulite-facies mineral assemblages and uncommonly show a small obliquity (<10°) to the regional foliation that is independent of deformation state. Granitic plutons intrude the tonalites and are generally characterized by granulite-facies mineral assemblages and a variably-developed tectonic foliation (Fig. 5e). It should be noted that the foliation in the granitic plutons and veins, where present, is not defined by the dimensional alignment of undeformed primary igneous phases such as plagioclase (as would be the case for magmatic foliations; see Paterson *et al.* 1989), but by dimensional alignment of quartz and K-feldspar that have undergone intracrystalline strain (Fig. 5e). Tonalite and granitic plutons commonly contain centimetre-scale to kilometre-scale xenoliths of tonalite, quartz diorite, amphibolite and pyroxenite (St-Onge *et al.* 1992; Fig. 5f). Both foliated and massive granitic plutons trend E–W (Fig. 2) and thus appear to have been emplaced roughly parallel to the E-trend of the tectonic foliation in the tonalites.

The presence of sub-parallel tectonic foliations defined by granulite-facies mineral assemblages in the tonalites and at least some of the granites and granitic veins indicates that these foliations formed during a single Archean tectonic event at granulite-facies conditions. A magmatic origin for the foliation is ruled out because of the relative constancy of its orientation through rocks of substantially different intrusive ages (e.g. granites and tonalites) and because it is defined by granulite-facies mineral assemblages in a wide variety of rock types (see Paterson *et al.* 1989 for criteria for magmatic vs tectonic foliations). Observations that the granitic veins and plutons locally cross-cut the tonalite foliation further suggest granitic magmatism, granulite-facies metamorphism and deformation were broadly coeval (Lucas & St-Onge 1992).

The relative strain levels associated with the foliation-forming event have been assessed by integrating microstructural observations (Figs. 5c, d & e) with field observations on the regional distribution and intensity of foliations (Fig. 5b) and the aspect ratios of relatively competent xenoliths (gabbros, pyroxenites; Fig. 5f; cf. Passchier *et al.* 1990, p. 54). Of particular importance is the observation that both host rock and vein foliations are tectonic and defined by inequidimensional metamorphic phases (e.g. orthopyroxene, hornblende) and/or plastically-deformed quartz and feldspar (Figs. 5c, d & e). This is despite the fact that the rocks have undergone static annealing, which appears to have principally affected quartz (e.g. Figs. 5c & e). Recognizing that such an assessment has potential pitfalls given the diversity of plutonic rock compositions and intrusive body geometries (see Paterson & Tobisch 1988), we nonetheless suggest that the host tonalites and quartz diorites accumulated more solid-state deformation during this tectonic event than the granitic veins in general (e.g. Fig. 5b). Given the apparent contrast in strain between some of the veins and their host rocks, it seems difficult to argue that all of the veins have been transposed into parallelism with the host rock foliation/layering. From this it follows that at least *some* of the granitic veins were emplaced sub-parallel to host rock foliation/layering during deformation.

Paleoproterozoic Narsajuaq arc

The Narsajuaq arc is represented by a broad range of plutonic rock compositions (Fig. 3; St-Onge *et al.* 1992). The oldest unit is a well-layered sequence of tonalite and diorite, dominated by tonalite (70–80%) and layered on a centimetre to decametre scale (Figs. 6a & b). Preserved cross-cutting relationships indicate that tonalite commonly intrudes diorite, although the converse is observed locally. A tonalite layer from this unit has a U–Pb zircon age of 1863±2 Ma whereas an age of 1844±12 Ma has been obtained for a quartz-diorite layer from near the dated tonalite (St-Onge *et al.* 1992). Amphibolite, pyroxenite and peridotite occur within the layered sequence as both discrete bands, presumably representing veins (sills, dykes) and/or deformed xenoliths, and more equant xenoliths. However, these are in general not cognate xenoliths or microgranitoid enclaves because in lower strain areas they can be seen to have angular, blocky shapes (see Vernon *et al.* 1988, Vernon & Paterson 1993).

All rock types in this layered sequence are intruded by granitic veins (Figs. 6b–e). A granodiorite vein, which was sampled near the 1863±2 Ma tonalite outcrop, has been dated at 1861±2 Ma (St-Onge *et al.* 1992), indicating contemporaneous emplacement of tonalite and granodiorite sheets. A U–Pb zircon age of 1848±5 Ma was obtained for a monzogranite vein from the same outcrop as the 1861±2 Ma granodiorite vein and the 1844±12 Ma quartz-diorite layer (St-Onge *et al.* 1992,

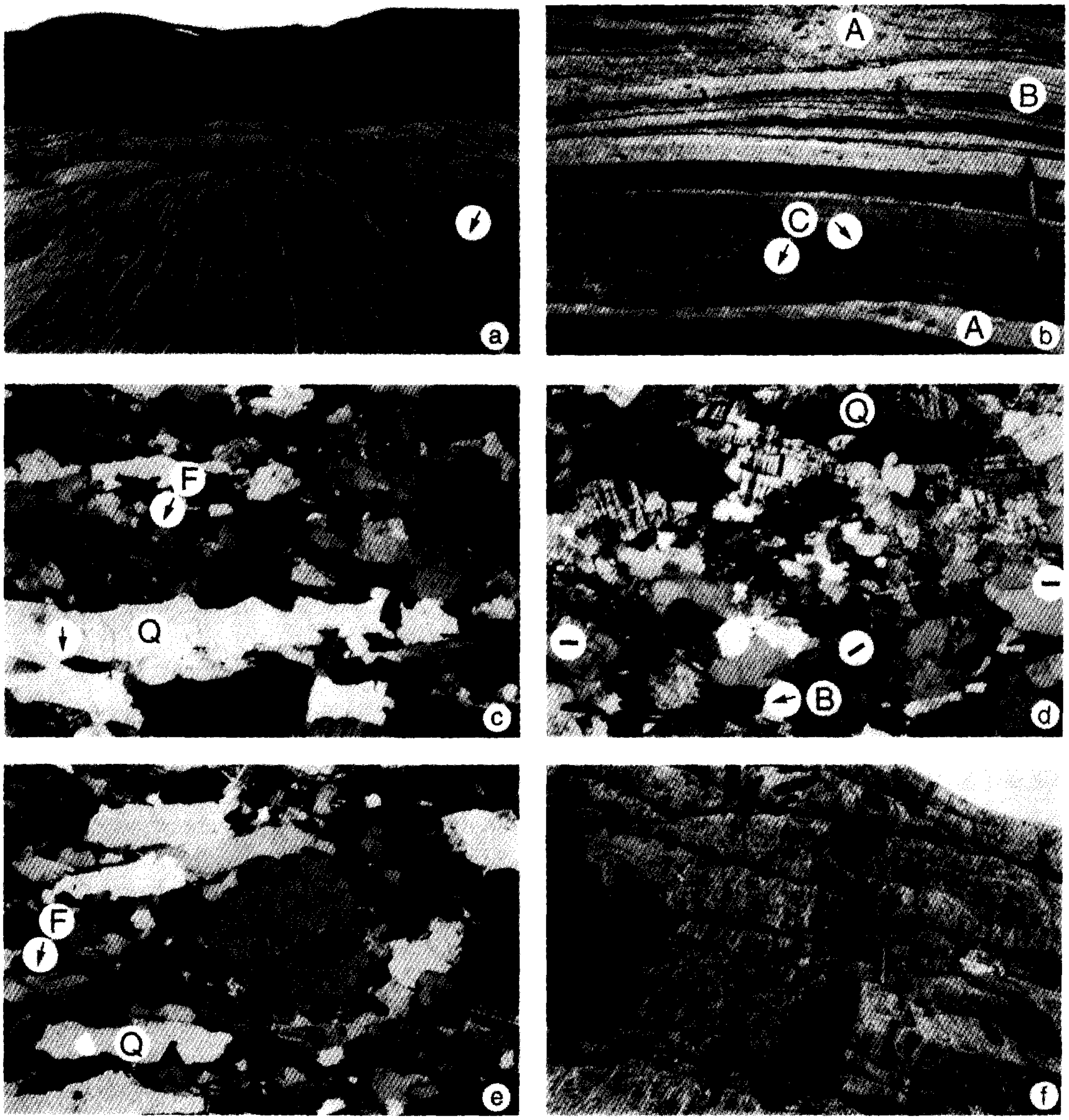

Fig. 5. (a) Archean biotite tonalite with foliation-sub-parallel monzogranite veins. The relatively uniform spacing of the veins and their apparent lateral continuity effectively define the gneissic layering in the outcrop. Hammer (arrowed) is 34 cm long. Veins vary in orientation from concordant with foliation to slightly (<10°) discordant, apparently independent of their bulk deformation state. (b) Sequence of monzogranite veins (light) lying sub-parallel to the foliation in Archean biotite tonalite. The veins range in mesoscopic deformation state from unfoliated (A), to variably foliated (B), to well foliated and effectively indistinguishable in terms of the degree of foliation development from the tonalite. 'C' indicates veinlets probably generated by *in situ* partial melting. Pen is 15 cm long. (c) Photomicrograph illustrating the tectonic foliation in an Archean tonalite defined by the dimensional alignment of quartz (Q) and biotite grains and polycrystalline feldspar aggregates (F). Note evidence for post-tectonic grain growth in quartz (arrow). Width of field of view is 12 mm. (d) Photomicrograph illustrating the tectonic foliation in an Archean granitic vein defined by elongate quartz grains (Q; top half of photo). Vein/host rock contact is marked on photo, and parallels tectonic foliation in host tonalite defined by dimensional alignment of biotite (B) and quartz grains. Width of field of view is 12 mm. (e) Photomicrograph illustrating the tectonic foliation in a sample from an Archean monzogranite pluton. Foliation is defined by the dimensional alignment of quartz (Q) and biotite grains and polycrystalline feldspar aggregates (F). Note evidence for post-tectonic grain growth in quartz. Width of field of view is 12 mm. (f) Xenoliths of foliated and massive quartz diorite, gabbro and pyroxenite in a tonalite pluton from the Archean Superior Province basement. The variable shape, pre-existing foliation and compositional range of the xenoliths indicates that they are most likely country rock fragments as opposed to microgranitoid enclaves (see Vernon *et al.* 1988, Vernon & Paterson 1993). Hammer is 45 cm long.

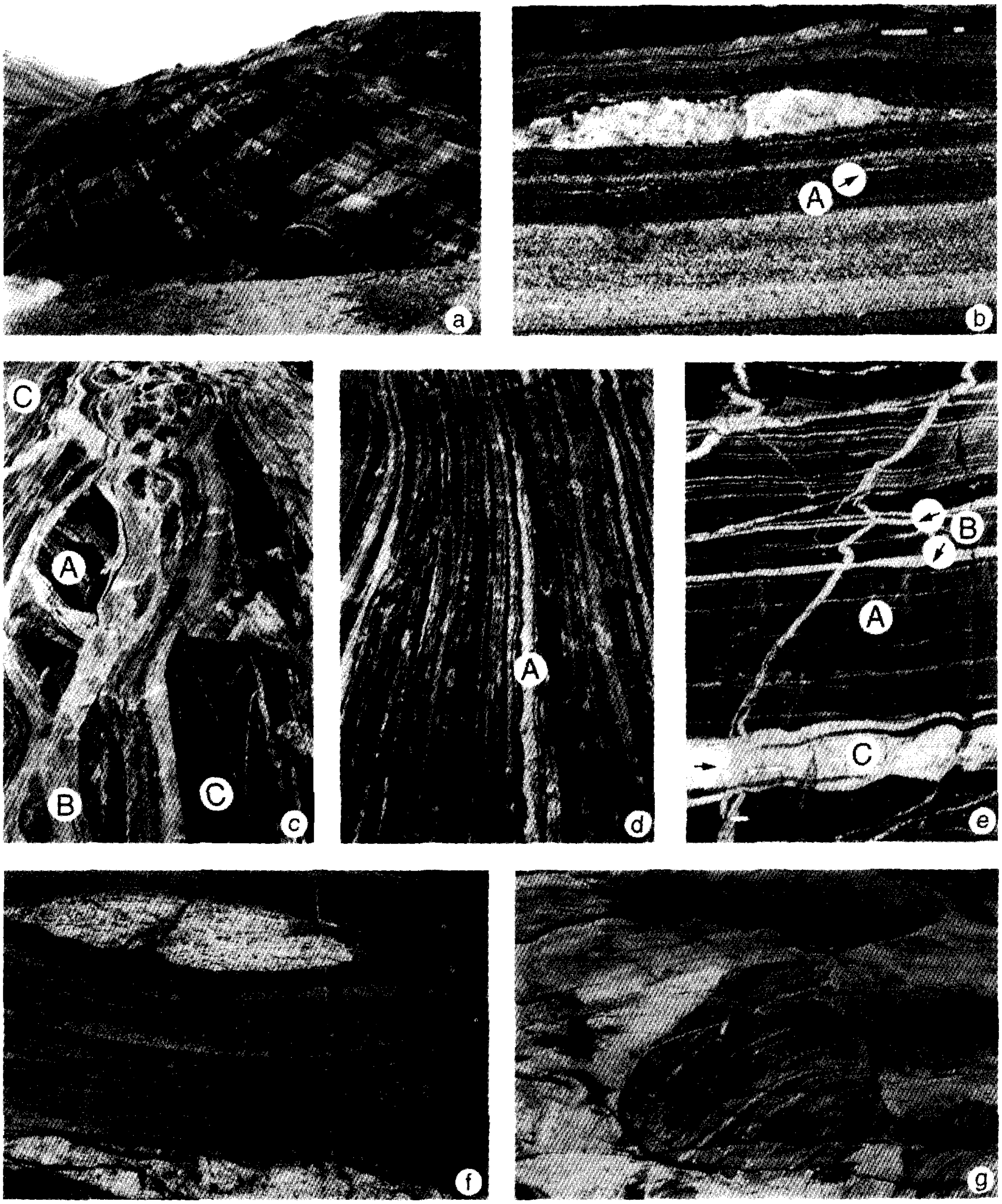

Fig. 6. (a) Layered tonalite–quartz-diorite unit of the Narsajuaq arc (*ca.* 1863 Ma). Granulite-facies foliation is parallel to the compositional layering. Person for scale. (b) Equigranular biotite monzogranite vein lying parallel to foliation and compositional layering in layered tonalite–quartz-diorite unit of the Narsajuaq arc. Monzogranite vein (top) is mesoscopically undeformed in its core but acquires a foliation parallel to that in the enveloping rock as it narrows at either end. Biotite selvedges at the margins of the vein may reflect local reaction between the vein and the wall rock (see Olsen 1983). 'A' indicates veinlets probably generated by *in situ* partial melting. Pen is 15 cm long. (c) Anastomosing, deformed biotite ± hornblende monzogranite veins (light), probably similar in age to the younger (*ca.* 1836–1830 Ma) plutons that cut the gneissic layering in the older tonalite–quartz-diorite unit of the Narsajuaq arc. The veins include rotated blocks of the country rock (A) but are themselves foliated (B) parallel to the regional high-grade fabric (C). Hammer is 45 cm long. (d) Higher strain zone from the same outcrop as Fig. 6(c). The well layered gneiss comprises quartz diorite, tonalite and monzogranite layers. Veins vary in orientation from concordant with foliation (e.g. A) to slightly (<10°) discordant. Hammer is 45 cm long. (e) Multiple ages of granitic veins emplaced in foliated quartz diorite of the older layered tonalite–quartz-diorite unit in Narsajuaq arc. Diagonal, cross-cutting veins (A) post-date emplacement of some narrow, foliation-parallel veins (e.g. B) but are themselves cut by other deformed, foliation-parallel veins (C). Pen (arrowed) is 15 cm long. (f) Deformed xenoliths of tonalite (light) and diorite (dark) in a younger tonalite pluton (1830±2 Ma) of the Narsajuaq arc. The compositional and textural variability of the xenoliths indicates that they are most likely country rock fragments as opposed to microgranitoid enclaves (see Vernon *et al.* 1988, Vernon & Paterson 1993). Hammer is 34 cm long. (g) Large, blocky xenolith of layered tonalite–diorite engulfed within a late monzogranite pluton (*ca.* 1800 Ma; R. Parrish personal communication 1993) of the Narsajuaq arc (north of 'A', Fig. 3). The monzogranite itself contains only a very weakly and heterogeneously developed tectonic foliation. Hammer is 45 cm long.

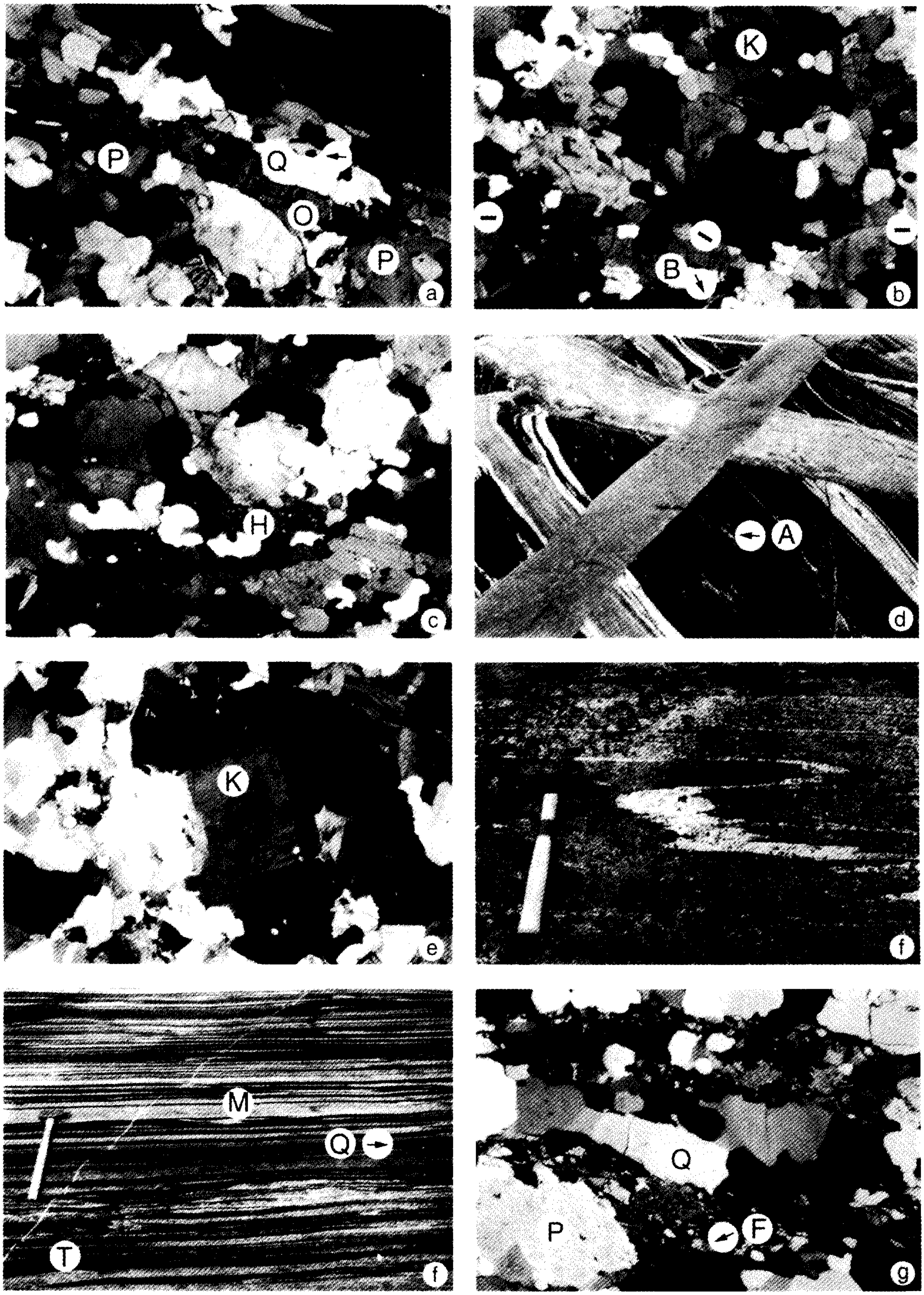

Fig. 7. *Caption overleaf.*

Fig. 7. (a) Photomicrograph illustrating the tectonic foliation in a Narsajuaq arc tonalite defined by the dimensional alignment of orthopyroxene (O), plagioclase (P) and quartz (Q). The presence of orthopyroxene defining the foliation indicates that it formed at granulite-facies conditions. Note that annealed quartz (Q) has isolated elongate feldspar grains (arrow). Width of field of view is 5 mm. (b) Photomicrograph illustrating the contact (marked on photo) between a foliated granodiorite vein (bottom of photo) and an unfoliated monzogranite vein. The granodiorite vein has a U–Pb zircon age of 1861 ± 2 Ma, and its foliation is defined by the dimensional alignment of biotite (B). Note the subhedral K-feldspar grain (K) in the monzogranite vein. Width of field of view is 12 mm. (c) Photomicrograph illustrating the tectonic foliation in a quartz-diorite vein dated at 1844 ± 12 Ma from the Narsajuaq arc. Foliation is principally defined by the dimensional alignment of hornblende (H) in both the quartz diorite and the coarser grained segregation (top of photo). Width of field of view is 12 mm. (d) Granitic veins cutting the layered tonalite–quartz-diorite unit in Narsajuaq arc. At least five generations of veins are present, and vary in deformation state from well foliated (layer-parallel veins) to undeformed (vein cutting diagonally from bottom left to top right). 'A' indicates veinlets probably generated by *in situ* partial melting. Pen is 15 cm long. (e) Photomicrograph of a layer-parallel monzogranite vein with no discernible foliation from the Narsajuaq arc. Note the subhedral, zoned K-feldspar grain (K). The vein is from the same outcrop as Figs. 7b & c, and has a U–Pb zircon age of 1848 ± 5 Ma (R. Parrish personal communication 1992). Width of field of view is 12 mm. (f) Folded granitic vein in tonalite from the Narsajuaq arc, showing evidence for boudinage on long limbs (subparallel to foliation) and tight folding ('thickening') on short limbs. Hammer is 34 cm long. (g) High strain fabric (compositional layering) in Narsajuaq arc rocks generated during deformation adjacent to the sole fault on which the arc overrode the Superior Province basement and its continental margin deposits during arc–continent collision at *ca.* 1.8 Ga (Lucas & St-Onge 1992). Compositional layering defined by tonalite (T), quartz diorite (Q) and monzogranite (M) bands. Photograph taken along west shore of Deception Bay (Fig. 2). Hammer is 45 cm long. (h) Photomicrograph of relatively high strain (mylonitic) foliation in monzogranite from the Archean Superior Province basement. Note development of elongate, polycrystalline quartz (Q) and feldspar (F) ribbons. Fine grained feldspar in ribbons appears to be derived from strain-induced recrystallization of feldspar porphyroclasts (P). Width of field of view is 5 mm.

R. Parrish personal communication 1992). A monzogranite vein with a strongly-developed, granulite-facies tectonic foliation has yielded a U–Pb zircon age of 1835±1 Ma (Parrish 1989). Discrete, kilometre-scale plutons of quartz diorite, quartz monzonite–monzogranite (1836±0.5 Ma, 1834±0.6 Ma) and tonalite (1830±2 Ma) intrude the older layered rocks of Narsajuaq arc (Fig. 3; Parrish 1989, St-Onge *et al.* 1992). The plutons vary in diameter from <1 km to tens of km, and although they generally trend parallel to the regional tectonic fabric, they locally have irregular shapes that cross-cut both outcrop- and regional-scale layering (e.g. 'B', Fig. 3). However, outcrop-scale granitic veins, some of which may be related to these younger (1836–1830 Ma) plutons, are generally sub-parallel to foliation/layering in the host rocks (Figs. 6b–d). The plutons and veins locally include abundant xenoliths (Figs. 6c & f) derived from the country rocks.

A penetrative tectonic foliation that parallels the outcrop-scale compositional layering characterizes most of the plutonic rocks in the Narsajuaq arc (Figs. 6b, d & e). The generally steep, E-trending foliations are defined by amphibolite to granulite-facies metamorphic minerals (e.g. hornblende, biotite, orthopyroxene) in the older layered sequence, the younger plutons and the granitic veins (Figs. 7a–c). The presence of metamorphic orthopyroxene in virtually all rock types (e.g. Fig. 7a) indicates that the regional foliation developed during granulite-facies metamorphism (Lucas & St-Onge 1992). Preliminary estimates of the metamorphic conditions during this event, based on thermochemical modeling of mineral compositions, are 800±50°C and 7±1 kbar (Monday 1994).

Cessation of regional penetrative deformation within the Narsajuaq arc during the waning stages of arc magmatism and granulite-facies metamorphism is indicated by variations between granitic vein orientation and internal strain relative to the host rocks (Figs. 6e and 7d). A weakly foliated, orthopyroxene-bearing tonalite vein that cross-cuts the layering in 1830 ± 2 Ma tonalite gneisses has a U–Pb zircon age of 1826 ± 1 Ma (Parrish 1989, St-Onge *et al.* 1992). The cross-cutting veins shown in Fig. 7(d) are inferred to be late to post-tectonic with respect to the foliation/layering in the gneisses, in contrast to variably foliated, layer-parallel veins that are inferred to be syn-tectonic. Some of the undeformed, layer-parallel and cross-cutting granitic veins may be related to either collisional granites (1803–1800 Ma; R. Parrish personal communication 1992, 1993) and/or post-collisional intrusive rocks, including pegmatitic syenogranite dykes (1759 ± 1 Ma; Parrish 1989) and granite plutons (1742 ± 2 Ma; Dunphy *et al.* 1991). The collisional granites contain large screens of country rock (e.g. 'A', Fig. 3) that are generally at angles of <15° to the strike of regional foliation/layering, suggesting that the plutons may have been emplaced as layer-parallel sheets, possibly at mid-crustal levels (St-Onge *et al.* 1992). Significant rotation of these large plutons, thereby producing an apparent sheet-like geometry, can be ruled out because of the low bulk strain both within and at the margins of the intrusions (Fig. 6g).

The finite strain associated with foliation/layering in the Narsajuaq arc varies in general with the age of the rock (e.g. 1863 vs 1826 Ma tonalites), in addition to obvious considerations such as competency and proximity to discrete structures such as shear zones. Finite strains were qualitatively evaluated in the field (cf. Passchier *et al.* 1990, p. 54) based on the relative intensity of tectonic foliation development and the aspect ratios of relatively competent xenoliths (tonalite, diorite, gabbro, pyroxenite; Figs. 6f & g), and were verified with microstructural studies (Figs. 7a–c). The microstructural studies indicate that the mesoscopically more intense foliations display a greater degree of syn-tectonic recrystallization and grain size reduction relative to mesoscopically more weakly foliated rocks (compare Figs. 7a, b & e). The older layered tonalite–diorite sequence contains a tectonic foliation throughout the area (Fig. 7a), whereas the younger, discrete plutons and granitic veins are variably foliated but in general are less deformed than the layered sequence (Figs. 7b & e). Solid-state deformation of the younger intrusive rocks is heterogeneous, as illustrated by shape variations in strain markers such as granitic veins (compare Figs. 6c & d) and host rock xenoliths.

In order to understand the relations between magmatism, deformation and the development of compositional layering in the Narsajuaq arc, it is important to establish whether or not rocks in the older layered sequence (1863–1844 Ma) were emplaced syn-tectonically. Two principal observations support a syn-tectonic interpretation for the construction of the layered sequence. First, the older intrusive sequence had to have been deformed prior to emplacement of the younger plutons (1836–1830 Ma) because the younger plutons locally cross-cut layering in the older sequence. Second, within the layered sequence, there are contrasts in the relative state of strain between individual layers that cannot be solely attributed to competency contrasts, but instead suggest that at least some of the less deformed layers were emplaced as foliation-parallel sheets. Deformation and metamorphism in the Narsajuaq arc thus appear to have been coeval with the bulk of the 1863–1830 Ma magmatism.

DEVELOPMENT OF THE COMPOSITIONAL LAYERING

The principal observation from both the Archean Superior Province and the Paleoproterozoic Narsajuaq arc is that syn-tectonic granitic veins and sheets, independent of their relative state of strain, in general parallel the foliation/layering of their host rocks (Figs. 5d & 7b). We suggest that this vein–host rock relationship was largely produced by syn-tectonic layer-parallel emplacement of the granitic bodies as opposed to post-emplacement transposition of oblique veins. However,

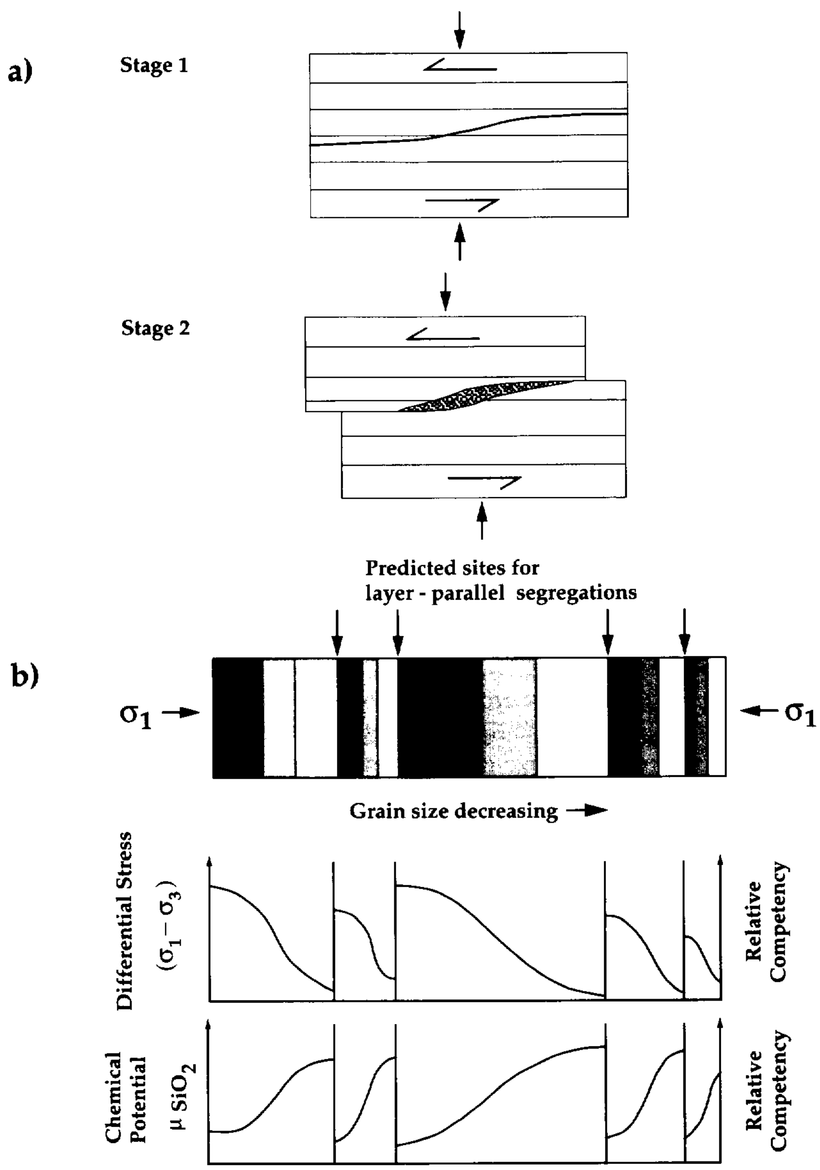

Fig. 8. Schematic models illustrating possible processes through which foliation/layer-parallel veins may develop: (a) vein opening at a dilational jog during bulk shear along foliation/layering [Sawyer & Robin 1986, Hutton & Reavy 1992, Sawyer 1994, see also Tikoff & Teyssier (1992) for '*P*-shear tensional bridge' model]; (b) layer-parallel vein segregation sites (small arrows) at positions of minimum σ_3 and maximum chemical potential of silica (μ_{SiO_2}) in graded greywacke–pelite beds (after Robin 1979, Sawyer & Robin 1986); (c) progressive bulk shear deformation of extension veins (for a discussion of folded veins in shear zones, see Hudleston 1989, Passchier *et al.* 1990, Hanmer & Passchier 1991, and references therein); (d) vein emplacement along extension fractures sub-parallel to layering and at a high angle to regional σ_1, due to a combination of a pre-existing anisotropy and high melt pressures (after Gratier 1987, Wickham 1987, Holliger & Levander 1994a). As illustrated with the schematic Mohr circle construction and block diagram, T_1 refers to the tensile strength of the rock perpendicular to foliation (i.e. parallel to σ_1) whereas T_3 refers to tensile strength parallel to foliation (i.e. parallel to σ_3).

we do recognize that oblique syn-tectonic veins (e.g. Figs. 6e & 7f), as well as oblique and layer-parallel post-tectonic veins (e.g. Fig. 7e), also characterize the gneisses of the Superior Province and the Narsajuaq arc and do not attempt to explain *all* veins in the following model. We also acknowledge that there may be, or may have been, magmatic foliations in some of the plutonic rocks, but believe for the reasons given earlier that the foliations investigated in this study are tectonic in origin.

Foliation/layer-parallel veins may develop through several possible processes: (1) vein opening at a dilational jog during bulk shear along foliation/layering [Fig. 8a; Sawyer & Robin 1986, Hutton & Reavy 1992, Sawyer 1994, see also Tikoff & Teyssier (1992) for the '*P*-shear tensional bridge' model]; (2) layer-parallel vein segregation at positions of minimum values of the least principal stress (σ_3) and maximum chemical potential for the vein species in rheologically stratified rocks (Fig. 8b; Robin 1979, Sawyer & Robin 1986); (3) bulk shear deformation of oblique extension veins, possibly accommodated by localized shear along vein walls or vein deformation (Fig. 8c; Myers 1978, Hudleston 1989, Passchier *et al.* 1990, Hanmer & Passchier 1991, Kröner *et al.* 1994); and (4) vein emplacement along extension fractures sub-parallel to layering (Fig. 8(d); Gratier 1987, Wickham 1987, Holliger & Levander 1994a; see also McCarthy & Thompson 1988, Parsons *et al.* 1992). In the scenario illustrated in Fig. 8(d), veins can be

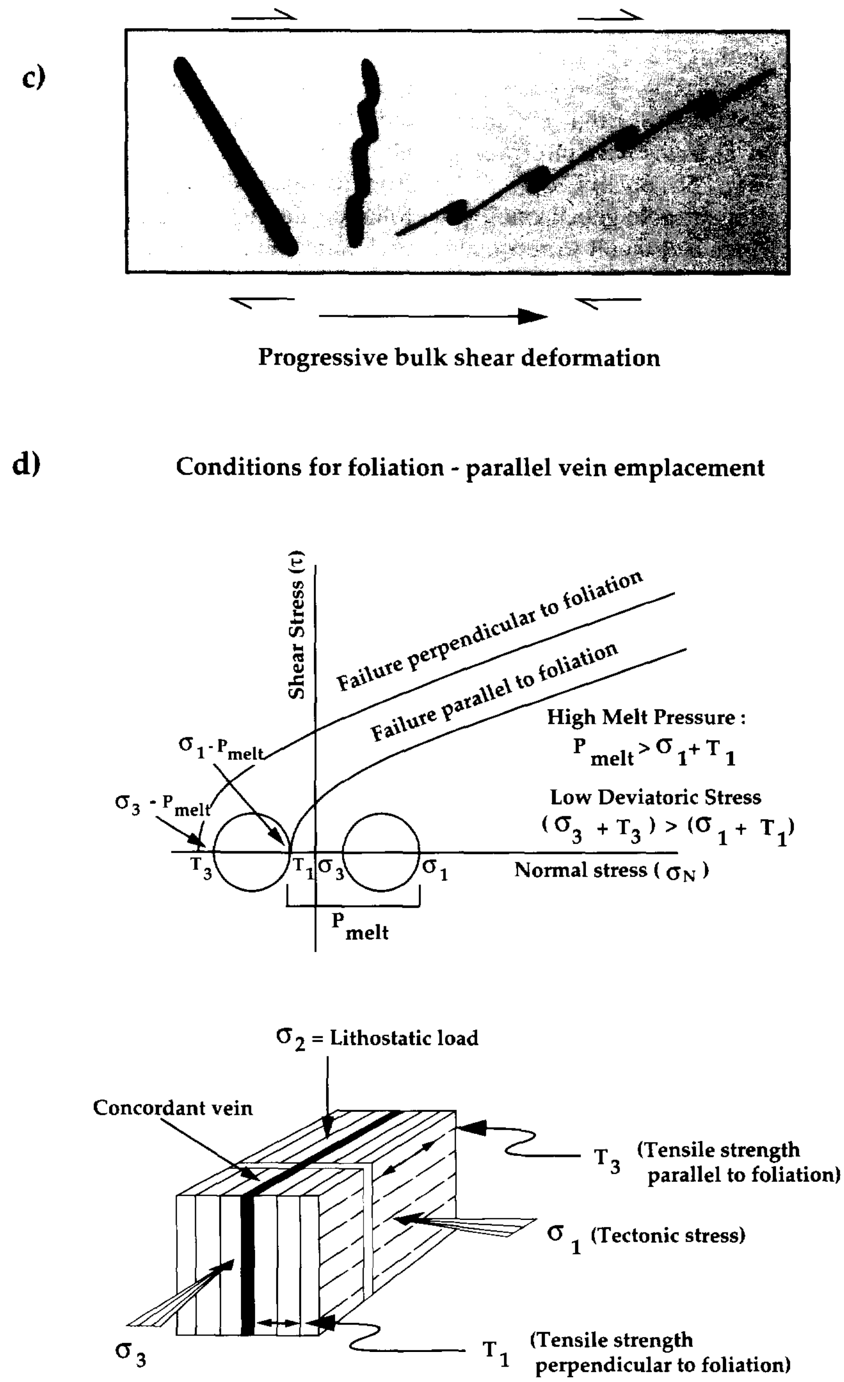

Fig. 8. Continued.

emplaced at high angles to σ_1 (maximum principal stress) when the local deviatoric stress ($\sigma_1 - \sigma_3$) is less than the difference in tensile strength of the rock parallel and perpendicular to foliation (i.e. T_3-T_1>$\sigma_1 - \sigma_3$), and the melt pressure (P_{Melt}) is sufficiently high that it exceeds the sum of σ_1 and the tensile strength of the rock parallel to σ_1 (i.e. P_{Melt}>σ_1+T_1).

The granitic veins in both the Superior Province and the Narsajuaq arc are not considered to be derived from *in situ* partial melting (migmatization) during their respective granulite-facies deformation events for several reasons. First, some of the veins have intrusive relations such as slightly discordant contacts (<10°, Figs. 6c–e) and wallrock xenoliths (Fig. 6c) suggesting an igneous injection mode of emplacement. Second, the veins have a range of compositions (e.g. quartz diorite–syenogranite) that varies independent of host rock composition and bears no systematic relation to predicted minimum melt compositions (see Olsen 1983, Johannes 1988). However, the presence of some mafic (biotite-rich) selvedges at vein margins indicates that the injected melts reacted to some degree with the host rock (Fig. 6b; see Olsen 1983). Third, the metamorphic temperatures associated with the syn-tectonic magmatism (800±50°C) are such that a maximum of only 15–20% melting would be expected for biotite-bearing quartzofeldspathic gneisses (e.g. Waters 1988) and less for hornblende-bearing units, which is not consistent with the volume of melt (>30%, Wickham 1987; but see Sawyer 1994) needed to segregate the veins that are laterally continuous for tens of metres (Figs. 5a, 6a & c). However, we do recognize centimetric granitoid veinlets that are quite distinct from the above veins, and which we interpret to result from partial melting (Figs. 5b, 6b and 7d).

Finally, the veins show a remarkable geochronological and isotopic diversity, which is difficult to reconcile with an origin by any process other than intrusion of externally-derived melts. One outcrop of the Narsajuaq arc layered tonalite–diorite sequence, studied in detail, has layer-parallel veins that yield U–Pb zircon ages of 1861±2 Ma (granodiorite; Fig. 7b), 1848±5 Ma (monzogranite; Fig. 7e) and 1844±12 Ma (quartz diorite; Fig. 7c). The veins were emplaced into tonalite, for which an age of 1863±2 Ma (St-Onge *et al.* 1992; Parrish personal communication 1992) has been obtained from a nearby outcrop. The tonalite and quartz diorite vein have ε_{Nd} values (at their crystallization ages) of +3 to +4 whereas the monzogranite vein yielded an ε_{Nd} value of −10 (J. Dunphy personal communication 1992). Other monzogranite veins from the Narsajuaq arc have ε_{Nd} values (at the inferred crystallization age of 1845 Ma) of −1 and −5, again indicating the significant heterogeneity in Nd-isotopic compositions of veins and adjacent rocks. These results require an external origin for such veins relative to their host rocks. However, we do not have any constraints on the relative transport distances for the vein melts except to say that at least some are probably related to nearby plutons, thus indicating minimum transport distances of hundreds of metres to kilometres.

Stress-induced diffusive mass transfer (Fig. 8b) is ruled-out as a major granitic vein-forming process because the vein compositions and isotopic character appears in general to bear no systematic relation to wallrock composition and isotopic character, as discussed above. The observation that the majority of the syn-tectonic veins lie sub-parallel to foliation/layering, independent of their state of strain (e.g. Figs. 5b, 6b, c & e), implies that they could not have rotated substantially following intrusion (e.g. Fig. 8c). Multiple episodes of syn-tectonic vein emplacement are indicated by cross-cutting relations in Fig. 6(e): oblique (cross-cutting) veins, themselves deformed ('A'), post-date emplacement of some narrow, foliation-parallel veins ('B') but are themselves cut by other foliation-parallel veins ('C') that are also deformed. That the oblique veins are cut by foliation-parallel veins, themselves deformed, precludes a rotational (transpositional) explanation for the foliation-parallel orientation of at least the cross-cutting veins, and by analogy, some or all of the older foliation-parallel veins. Not only are the bulk of the concordant veins not folded and/or boudined (e.g. Fig. 7f), we have not been able to find any evidence of wall rock deformation that may have been associated with vein rotation, such as folds (cf. Hudleston 1989) or fine-scale shear zones in or adjacent to vein walls (Figs. 5d and 7b). Finally, the presence of foliated wall rock xenoliths that do not show any regular or predictable rotation pattern or that are parallel to both veins and layering suggests that the veins could not have been significantly rotated following emplacement. Outcrop observations on the relative state of strain of both veins and host-rock are corroborated by thin section observations. As described above, the microstructures in layer-parallel veins vary from no discernible foliation (Figs. 7b & e) to a tectonic foliation defined by grain shape (Figs. 5d and 7c) to a mylonitic foliation in which quartz and feldspar show optical evidence for crystal-plastic strain (e.g. optical subgrains, strain-induced recrystallization; Figs. 7g & h).

The field and thin section observations and isotopic data support a model of syn-tectonic emplacement of externally-derived granitic melts in extension fractures parallel to the foliation/layering (Fig. 8d). This model does not require the transposition of all veins (cf. Myers 1978, Kröner *et al.* 1994), and therefore does not presuppose the amount of strain inherent in the banded gneisses of the Paleoproterozoic Narsajuaq arc and Archean Superior Province. However, it is difficult to discern whether layer-parallel veining occurred (1) at high angles to the regional shortening direction in the presence of melt pressures exceeding lithostatic pressure (Fig. 8d), or (2) along transtensional steps associated with shear bands (Fig. 8a). Observations of both strictly concordant (e.g. Fig. 6b) and slightly discordant (e.g. Figs. 6c & d) syn-tectonic veins may indicate that both of these mechanisms were operative. The low deviatoric stresses required by the model shown in Fig. 8(d) would be expected for high-temperature (~800°C), lower crustal tectonic regimes (see Etheridge 1983, Carter & Tsenn 1987, Ranalli & Murphy 1987, Ord & Hobbs 1989) such as the Superior Province basement at *ca.* 2.9–2.7 Ga or the Narsajuaq arc at 1.86–1.83 Ga. It is noted that alternative, although not entirely independent, models have been discussed for sill and dyke emplacement at high angles to the regional maximum principal stress (σ_1) direction, principally to explain the development of horizontal layering during lithospheric extension (McCarthy & Thompson 1988, Parsons *et al.* 1992). These models posit that differences in the deviatoric stress supported by layers of contrasting rheology, coupled with high magma pressures, can effectively result in principal stress interchanges in the weaker layers that enable extension fracturing to occur perpendicular to the regional σ_1 direction.

IMPLICATIONS

If the granitic veins were emplaced in extension fractures parallel to host-rock layering, as suggested by our observations, this process has substantial importance for the development and mechanical behaviour of the middle and lower crust. We examine three principal implications. First, syn-tectonic intrusion of layer-parallel granitic veins can generate a regional compositional layering without requiring high transpositional strains. A corollary of this is that if the middle/lower crust is undergoing heterogeneous or localized deformation (i.e. along a ductile thrust zone; see Davidson 1984, Hanmer 1988), then the deformation zones themselves may become loci for granitic intrusion, again as layer-parallel veins and sheets (Hanmer & McEachern 1992). Careful documentation of outcrop-scale structures within 'gneiss' belts is required to distinguish tectonite or 'straight gneiss' (e.g. Myers 1978, Davidson 1984, Hanmer 1988, Kröner *et al.* 1994) from layered or gneissic rocks that have not experienced the same intensity of deformation and thus may have a different tectonic significance.

Second, syn-tectonic veining may focus deformation into transiently weak corridors (cf. Hollister & Crawford 1986) by partitioning of deformation into partially solidified granitic magmas. However, we have not observed any unequivocal evidence of magmatic flow structures (e.g. foliations, Paterson *et al.* 1989, Vernon & Paterson 1993) in the veins or plutons, due at least in part to overprinting solid-state deformation.

Third, the metre- to kilometre-scale of the granitic veins and sheet-like intrusions may contribute to the overall seismic reflectivity of the middle and lower crust. Seismic imaging of the deep crust has often shown it to be highly reflective (Green *et al.* 1988, 1990, Clowes *et al.* 1992, Lucas *et al.* 1993). This characteristic reflectivity is generally ascribed to compositionally layered rock produced by (1) ductile strain and/or (2) the intrusion of melts, generally thought to be of mafic composition (Hurich *et al.* 1985, Goodwin & Thompson 1988, McCarthy & Thompson 1988, Green *et al.* 1988, 1990, Warner 1990a, Holbrook *et al.* 1991, Milkereit *et al.* 1991, Litak & Hauser 1992). Two critical implications of synthetic seismograms for this study are that the reflectivity is enhanced by layering at all scales (i.e. below the minimum wavelength determined by source frequency), and that the seismic response includes many edge effects and both constructive and destructive interference from outcrop-scale layering and from anisotropy within layers (Christensen 1989, Fountain *et al.* 1990, Levander & Holliger 1992, Szymanski & Christensen 1993, Holliger & Levander 1994b). Furthermore, granitic veins/sheets can provide significant impedance contrast for reflections if the host rock has a sufficiently mafic composition (e.g. amphibolite to granulite-grade diorite, anorthosite, amphibolite; cf. Fountain *et al.* 1990, Warner 1990b, Milkereit *et al.* 1991). Alternatively, the veins/sheets could have a mafic composition and be emplaced in more felsic rocks to yield an equivalent impedance contrast (cf. McCarthy & Thompson 1988, Warner 1990a, Holbrook *et al.* 1991, Litak & Hauser 1992). The mid-crustal sections of the ancient magmatic arcs that are exposed in the Ungava orogen provide firm evidence that syn-tectonic, layer-parallel vein/sheet emplacement is an important process for forming deep crustal compositional layering, which thus may contribute to deep crustal reflectivity.

CONCLUSIONS

(1) Regional compositional layering was developed during episodes of granitic magmatism, ductile transpressional deformation and high grade metamorphism in both the Archean Superior Province basement and the Paleoproterozoic Narsajuaq arc.

(2) Syn-tectonic granite veins appear to have been intruded sub-parallel to existing layering (presumably at high angles to the regional shortening direction), thus contributing to the development of compositional layering independently of transpositional deformation.

(3) Compositional layering is enhanced by post-veining deformation, and syn-tectonic veining may focus deformation into transiently weak corridors (see Hollister & Crawford 1986).

(4) Compositional layering in orthogneiss complexes, generated and/or enhanced by emplacement of granitic (or mafic) veins and sheets at low angles to a pre-existing anisotropy, may contribute to the overall seismic reflectivity of the middle and lower crust.

Acknowledgements—The results presented in this paper stem from a regional mapping project undertaken by the Geological Survey of Canada (GSC) in northern Québec (1989–1991). The Polar Continental Shelf Project is acknowledged for generous helicopter support during the full course of the project. Field assistants Kathy Bethune, Janet Dunphy, John Ketchum, Patrick Monday, Emmanuel Saydeh and Robert Thivierge are thanked for their careful mapping. Brian Cullen is gratefully acknowledged for his assistance with the figures. Bernd Milkereit, Alan Green, Simon Hanmer, Jack Henderson and Paul Williams are gratefully acknowledged for their helpful comments on the manuscript, as are Richard Norris, David Shelley and an anonymous reviewer for their constructive journal reviews. Geological Survey of Canada Contribution Number 24793.

REFERENCES

Bégin, N. J. 1992. Contrasting mineral isograd sequences in metabasites of the Cape Smith Belt, northern Québec, Canada: three new bathograds for mafic rocks. *J. metamorph. Geol.* **10**, 685–704.

Carter, N. L. & Tsenn, M. C. 1987. Flow properties of continental lithosphere. *Tectonophysics* **136**, 27–63.

Clemens, J. D. & Mower, C. K. 1992. Granitic magma transport by fracture propagation. *Tectonophysics* **204**, 339–360.

Christensen, N. I. 1989. Reflectivity and seismic properties of the deep continental crust. *J. geophys. Res.* **94**, 17,793–17,804.

Clowes, R. M., Cook, F. A., Green, A. G., Keen, C. E., Ludden, J. N., Percival, J. A., Quinlan, G. M. & West, G. F. 1992. Lithoprobe: new perspectives on crustal evolution. *Can. J. Earth Sci.* **29**, 1813–1864.

Davidson, A. 1984. Identification of ductile shear zones in the southwestern Grenville Province of the Canadian Shield. In: *Precambrian Tectonics Illustrated* (edited by Kroner, A. & Greiling, R.). E. Schweizeart'sche Verlags buchhandlung, Stuttgart, 207–235.

Dunphy, J. M. & Ludden, J. N. 1994. Geochemical and isotopic characteristics of an Early Proterozoic accreted magmatic arc (Narsajuaq terrane), northern Quebec, Canada. In: *Abstracts with Programs. International Conference on Geochronology, Cosmochronology and Isotope Geology* **8**. USGS Circular 1107, 88.

Dunphy, J. M., Parish, R. & Ludden, J. 1991. Stitching together the Ungava orogeny: the geochemistry and age relationships of the Iac Duquet pluton. *Geol. Ass. Canada Prog. w. Abs.* **16**, A32.

Eskola, P. 1932. On the principles of metamorphic differentiation. *Bull. Comm. géol. Finl.* **97**, 68–77.

Etheridge, M. A. 1983. Differential stress magnitudes during regional deformation and metamorphism. Upper bound imposed by tensile fracturing. *Geology* **11**, 231–234.

Francis, D. M., Ludden, J. & Hynes, A. J. 1983. Magma evolution in a Proterozoic rifting environment. *J. Petrol.* **24**, 556–582.

Fountain, D. M. & Salisbury, M. H. 1981. Exposed cross-sections through the continental crust: implications for crustal structure, petrology and evolution. *Earth Planet. Sci. Lett.* **56**, 263–277.

Fountain, D. M., Salisbury, M. H. & Percival, J. A. 1990. Seismic structure of the continental crust based on rock velocity measurements from the Kapuskasing uplift. *J. geophys. Res.* **95**, 1167–1186.

Goodwin, E. B. & Thompson, G. A. 1988. The seismically reflective crust beneath highly extended terranes: Evidence for its origin in extension. *Bull. geol. Soc. Am.* **100**, 1616–1626.

Gratier, J. P. 1987. Pressure solution-deposition creep and associated tectonic differentiation in sedimentary rocks. In: *Deformation of Sediments and Sedimentary Rocks* (edited by Jones, M. E. & Preston, R. M. F.). *Spec. Publ. geol. Soc. Lond.* **29**, 25–38.

Green, A. G., Milkereit, B., Davidson, A., Spencer, C., Hutchinson, D. R., Cannon, W. F., Lee, M. W., Agena, W. F., Behrendt, J. C. & Hinze, W. J. 1988. Crustal structure of the Grenville front and adjacent terranes. *Geology* **16**, 788–792.

Green, A., Milkereit, B., Percival, J., Davidson, A., Parrish, R., Cook, F., Geis, W., Cannon, W., Hutchinson, D., West, G. & Clowes, R. 1990. Origin of deep crustal reflections: seismic profiling across high-grade metamorphic terranes in Canada. In: *Seismic Probing of Continents and Their Margins* (edited by Leven, J. H., Finlayson, D. M., Wright, C., Dooley, J. C. & Kennett, B. L. N.). *Tectonophysics* **173**, 627–638.

Hanmer, S. 1988. Ductile thrusting at mid-crustal level, southwestern Grenville Province. *Can. J. Earth Sci.* **25**, 1049–1059.

Hanmer, S. & Passchier, C. 1991. Shear-sense indicators: A review. *Geol. Surv. Pap. Can.* **90**, 17.

Hanmer, S. & McEachern, S. 1992. Kinematical and rheological evolution of a crustal-scale ductile thrust zone, Central Metasedimentary Belt, Grenville orogen, Ontario. *Can. J. Earth Sci.* **29**, 1779–1790.

Hoffman, P. F. 1989. Precambrian geology and tectonic history of North America. In: *The Geology of North America—An Overview* (edited by Bally, A. W. & Palmer, A. R.). *Geol. Soc. Am.*, Boulder Co., 447–512.

Holbrook, W. S., Catchings, R. D. & Jarchow, C. M. 1991. Origin of deep crustal reflections: Implications of coincident seismic refraction and reflection data in Nevada. *Geology* **19**, 175–179.

Hollister, L. S. & Crawford, M. L. 1986. Melt-enhanced deformation: A major tectonic process. *Geology* **14**, 558–561.

Holliger, K. & Levander, A. 1994a. Lower crustal reflectivity modeled by rheological controls on mafic intrusions. *Geology* **22**, 367–370.

Holliger, K. & Levander, A. R. 1994b. Structure and seismic response of extended continental crust: Stochastic analysis of the Strona-Ceneri and Ivrea zones, Italy. *Geology* **22**, 79–82.

Hudleston, P. J. 1989. The association of folds and veins in shear zones. *J. Struct. Geol.* **11**, 949–957.

Hurich, C. A., Smithson, S. B., Fountain, D. M. & Humphreys, M. C. 1985. Seismic evidence of mylonite reflectivity and deep structure of the Kettle Dome metamorphic core complex, Washington. *Geology* **13**, 577–580.

Hutton, D. H. W. & Reavy, R. J. 1992. Strike-slip tectonics and granite petrogenesis. *Tectonics* **11**, 960–967.

Johannes, W. 1988. On the origin of layered migmatites. In: *Migmatites, Melting and Metamorphism* (edited by Atherton, M. P. & Gribble, C. D.). Shiva Geology Series, Cheshire, 234–248.

Johannes, W. & Gupta, L. N. 1982. Origin and evolution of a migmatite. *Contrib. Mineral. Petrol.* **79**, 114–123.

Jordan, P. G. 1987. The deformational behaviour of bimineralic limestone–halite aggregates. *Tectonophysics* **135**, 185–197.

Kröner, A., Kehelpannala, K. V. W. & Kriegsman, L. M. 1994. Origin of compositional layering and mechanism of crustal thickening in the high-grade gneiss terrain of Sri Lanka. *Precamb. Res.* **66**, 21–37.

Levander, A. R. & Holliger, K. 1992. Small-scale heterogeneity and large-scale velocity structure of the continental crust. *J. geophys. Res.* **97**, 8787–8804.

Litak, R. K. & Hauser, E. C. 1992. The Bagdad Reflection Sequence as tabular mafic intrusions: evidence from seismic modeling of mapped exposures. *Bull. geol. Soc. Am.* **104**, 1315–1325.

Lucas, S. B. 1989. Structural evolution of the Cape Smith Thrust Belt and the role of out-of-sequence faulting in the thickening of mountain belt. *Tectonics* **8**, 655–676.

Lucas, S. B. 1990. Relations between thrust belt evolution, grain-scale deformation, and metamorphic processes: Cape Smith Belt, northern Canada. *Tectonophysics* **178**, 151–182.

Lucas, S. B. & Byrne, T. 1992. Footwall involvement during arc–continent collision, Ungava orogen, northern Canada. *J. geol. Soc. Lond.* **149**, 237–248.

Lucas, S. B., Green, A., Hajnal, Z., White, D., Lewry, J., Ashton, K., Weber, W. & Clowes, R. 1993. Deep seismic profile across a Proterozoic collision zone: Surprises at depth. *Nature* **363**, 339–342.

Lucas, S. B. & St-Onge, M. R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Québec. Part 2: Structural and metamorphic history. *Can. J. Earth Sci.* **29**, 765–782.

Lucas, S. B., St-Onge, M. R., Parrish, R. R. & Dunphy, J. M. 1992. Long-lived continent–ocean interaction in the Early Proterozoic Ungava orogen, northern Québec, Canada. *Geology* **20**, 113–116.

Machado, N., David, J., Scott, D. J., Lamothe, D., Philippe, S. & Gariepy, C. 1993. U–Pb geochronology of the western Cape Smith Belt, Canada: new insights on the age of initial rifting and arc magmatism. *Precamb.* **63**, 211–223.

McCarthy, J. & Thompson, G. A. 1988. Seismic imaging of extended crust with emphasis on the western United States. *Bull. geol. Soc. Am.* **100**, 1361–1374.

Milkereit, B., Percival, J. A., White, D., Green, A. G. & Salisbury, M. H. 1991. Seismic reflectors in high-grade metamorphic rocks of the Kapuskasing uplift: Results of preliminary drill site surveys. In: *Continental Lithosphere: Deep Seismic Reflections. Am. geophys. U. Geodynamics Series* **22**, 39–45.

Misch, P. 1949. Metasomatic granitization of batholithic dimensions: Pt. 1, Synkinematic granitization in Nanga Parbat area, northwest Himalayas. *Am. J. Sci.* **247**, 209–245.

Misch, P. 1968. Plagioclase compositions and non-anatectic origin of migmatitic gneisses in Northern Cascade Mountains of Washington State. *Contrib. Mineral. Petrol.* **17**, 1–70.

Monday, P. 1994. Narsajuaq terrane, Ungava Orogen, Canada: granulite-facies metasedimentary and metaigneous rocks and tectonic implications. Unpublished M.Sc. thesis, Queens University, Kingston, Ontario, Canada.

Myers, J. S. 1978. Formation of banded gneisses by deformation of igneous rocks. *Precamb. Res.* **6**, 43–64.

Olsen, S. N. 1983. A quantitative approach to local mass balance in migmatites. In: *Migmatites, Melting and Metamorphism* (edited by Atherton, M. P. & Gribble, C. D.). Shiva Geology Series, Cheshire, 201–233.

Ord, A. & Hobbs, B. E. 1989. The strength of the continental crust, detachment zones and the development of plastic instabilities. *Tectonophysics* **158**, 269–289.

Parrish, R. R. 1989. U–Pb geochronology of the Cape Smith Belt and Sugluk block, northern Québec. *Geoscience Can.* **16**, 126–130.

Parsons, T., Sleep, N. H. & Thompson, G. A. 1992. Host rock rheology controls on the emplacement of tabular intrusions: Implications for underplating of extending crust. *Tectonics* **11**, 1348–1356.

Passchier, C. W., Myers, J. S. & Kroner, A. 1990. *Field Geology of High-Grade Gneiss Terrains*. Springer-Verlag, Berlin.

Paterson, S. R. & Tobisch, O. T. 1988. Using pluton ages to date regional deformation: Problems with commonly used criteria. *Geology* **16**, 1108–1111.

Paterson, S. R., Vernon, R. H. & Tobisch, O. T. 1989. A review of criteria for the identification of magmatic and tectonic foliations in granitoids. *J. Struct. Geol.* **11**, 349–363.

Ramberg, H. 1952. *The Origin of Metamorphic and Metasomatic Rocks*. Chicago University Press, Chicago.

Ranalli, G. & Murphy, D. C. 1987. Rheological stratification of the lithosphere. *Tectonophysics* **132**, 281–295.

Robin, P.-Y. F. 1979. Theory of metamorphic segregation and related processes. *Geochim. Cosmochim. Acta* **43**, 1587–1600.

Sawyer, E. W. 1994. Melt segregation in the continental crust. *Geology* **22**, 1019–1022.

Sawyer, E. W. & Robin, P.-Y. F. 1986. The subsolidus segregation of layer-parallel quartz–feldspar veins in greenschist to upper amphibolite facies metasediments. *J. metamorph. Geol.* **4**, 237–260.

Shelley, D. 1974. Mechanical production of metamorphic banding, a critical appraisal. *Geol. Mag.* **3**, 287–292.

Scott, D. J., St-Onge, M. R., Lucas, S. B. & Helmstaedt, H. 1991. Geology and chemistry of the early Proterozoic Purtuniq ophiolite, Cape Smith Belt, northern Quebec, Canada. In: *Ophiolite Genesis and Evolution of The Oceanic Lithosphere* (edited by Peters, Tj.). Kluwer Academic Publishers, Amsterdam, 825–857.

Smithson, S. B., Johnson, R. A. & Hurich, C. A. 1986. Crustal reflections and crustal structure. In: *Reflection Seismology: The Continental Crust* (edited by Barazangi, M. & Brown, L. D.). *Am. geophys. U. Geodynamics Series* **14**, 21–32.

St-Onge, M. R. & Lucas, S. B. 1990. Evolution of the Cape Smith

Belt: Early Proterozoic continental underthrusting, ophiolite obduction and thick-skinned folding. In: *The Early Proterozoic Trans-Hudson Orogen of North America* (edited by Lewry, J. F. & Stauffer, M. R.). *Spec. Pap. geol. Assoc. Can.* **37**, 313–351.

St-Onge, M. R. & Lucas, S. B. 1991. Evolution of regional metamorphism in the Cape Smith Thrust Belt (northern Québec, Canada): interaction of tectonic and thermal processes. *J. metamorph. Geol.* **9**, 515–534.

St-Onge, M. R., Lucas, S. B. & Ijewliw, O. 1993. Thermotectonic evolution of basement gneisses and imbricated cover units during Paleoproterozoic arc–continent collision, Ungava orogen, Quebec. *Abs. w. Progs. geol. Soc. Am.*, Annual Meeting **25**, A–425.

St-Onge, M. R., Lucas, S. B. & Parrish, R. 1992. Terrane accretion in the internal zone of the Ungava orogen, northern Québec. Part 1: Tectonostratigraphic assemblages and their tectonic implications. *Can. J. Earth Sci.* **29**, 746–764.

St-Onge, M. R., Lucas, S. B. & Scott, D. J. In press. The Ungava Orogen and the Cape Smith Thrust Belt. In: *Tectonic Evolution of Greenstone Belts* (edited by de Wit, M. J. & Ashwal, L. D.). *Oxford Monographs on Geology and Geophysics Series*, Oxford University Press.

Szymanski, D. L. & Christensen, N. I. 1993. The origin of reflections beneath the Blue Ridge-Piedmont allochthon: A view through the Grandfather Mountain window. *Tectonics* **12**, 265–278.

Talbot, J. L. & Hobbs, B. E. 1968. The relationship of metamorphic differentiation to other structural features at three localities. *J. Geol.* **76**, 581–587.

Tikoff, B. & Teyssier, C. 1992. Crustal-scale, an echelon '*P*-shear' tensional bridges: A possible solution to the batholithic room problem. *Geology* **20**, 927–930.

Turner, F. J. 1941. The development of pseudo-stratification by metamorphic differentiation in the schists of Otago, New Zealand. *Am. J. Sci.* **239**, 1–16.

Vernon, R. H., Etheridge, M. A. & Wall, V. J. 1988. Shape and microstructure of microgranitoid enclaves: indicators of magma mingling and flow. *Lithos* **22**, 1–11.

Vernon, R. H. & Paterson, S. R. 1993. The Ardara pluton, Ireland: deflating an expanded intrusion. *Lithos* **31**, 17–32.

Warner, M. 1990a. Absolute reflection coefficients from deep seismic reflections. In: *Seismic Probing of Continents and Their Margins* (edited by Leven, J. H., Finlayson, D. M., Wright, C., Dooley, J. C. & Kennett, B. L. N.). *Tectonophysics* **173**, 15–23.

Warner, M. 1990b. Basalts, water, or shear zones in the lower continental crust? In: *Seismic Probing of Continents and Their Margins* (edited by Leven, J. H., Finlayson, D. M., Wright, C., Dooley, J. C. & Kennett, B. L. N.). *Tectonophysics* **173**, 163–174.

Waters, D. J. 1988. Partial melting and the formation of granulite facies assemblages in Namaqualand, South Africa. *J. metamorph. Geol.* **6**, 387–404.

Wickham, S. M. 1987. The segregation and emplacement of granitic magmas. *J. geol. Soc. Lond.* **144**, 281–297.

Williams, P. F. 1972. Development of metamorphic layering and cleavage in low-grade metamorphic rocks at Bermagui, Australia. *Am. J. Sci.* **272**, 1–47.

Williams, P. F. 1990. Differentiated layering in metamorphic rocks. *Earth Sci. Rev.* **29**, 267–281.

Yardley, B. W. D. 1978. Genesis of the Skagit Gneiss migmatites, Washington, and the distinction between possible mechanisms of migmatization. *Bull. geol. Soc. Am.* **89**, 941–951.

Zingg, A., Handy, M. R., Hunziker, J. C. & Schmid, S. M. 1990. Tectonometamorphic history of the Ivrea Zone and its relationship to the crustal evolution of the Southern Alps. In: *The Nature of the Lower Continental Crust* (edited by Fountain, D. M. & Boriani, A.). *Tectonophysics* **182**, 160–192.