PERGAMON

Journal of Structural Geology 23 (2001) 903–921

JOURNAL OF STRUCTURAL GEOLOGY

www.elsevier.nl/locate/jstrugeo

Formation of low-temperature mylonites and phyllonites by alkali-metasomatic weakening of felsic volcanic rocks during progressive, subduction-related deformation

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Received 5 April 1999; accepted 2 May 2000

Abstract

Dacitic to rhyolitic volcanic rocks of the Spruce Lake nappe experienced two phases of alkali-metasomatism as a result of fluids channelling along shear zones. The shear zones formed during a progressive, thrust-related deformation associated with underplating and incorporation of the volcanic rocks into the Brunswick subduction complex of northern New Brunswick. The fluids mainly represent chemically and isotopically modified seawater released by dewatering of the associated underthrusted shaly sedimentary rocks. Both phases of metasomatism weakened the felsic rocks, leading to strain localisation. Albitisation of felsic volcanic rocks as a result of Na-metasomatism during underthrusting facilitated formation of mylonites near peak high-pressure metamorphism (330–370°C, 600–800 MPa). The mylonites are preferentially preserved in the roof-thrust shear zone of the Spruce Lake nappe. Core-mantle structures, bulging and crystallographically preferred orientations indicate that albite behaved more ductilely than K-feldspar. The ductility of albite at these low temperatures is interpreted as a function of abundant intragranular fluids. Phengite-rich phyllonites formed after peak high–pressure metamorphism during uplift by out-of-sequence thrusting. These phyllonites are generally characterised by a slight gain in K and loss of Na and are best developed in the basal shear zones of the Spruce Lake nappe.

1. Introduction

Structural underplating occurs when subducting rocks bypass the toe of the wedge and are transferred from the down-going plate into the overriding subduction complex. This type of underthrusting and accretion has been observed seismically (e.g. Bernstein-Taylor et al., 1992); however, the mechanisms by which underplating is accommodated structurally and how the associated strain becomes localised are less well known (cf. Moore, 1989).

In this paper, we discuss the structures and deformation-induced alteration associated with underplating and the subsequent uplift of dacitic to rhyolitic volcanic rocks that, together with associated basalts and shales, were incorporated into the Late Ordovician to the Late Silurian (c. 445–425 Ma) Brunswick subduction complex (van Staal, 1994) in northern New Brunswick (Fig. 1). The subduction-related deformation (D_1) resulted in a complex stack of ductile thrusts, which were overprinted by a polyphase deformation history (D_2–D_4) during transpression and collapse (de Roo and van Staal, 1994). Subduction-related structures (D_1) and the products of the associated high-pressure–low-temperature metamorphism (M_1) are well preserved, despite the complex structural overprint. Preservation is mainly caused by continuous refrigeration (Ernst, 1988) of the Brunswick complex during uplift and exhumation of the underplated rocks and the heterogeneity of the subsequent deformations (van Staal, 1994; van Staal and de Roo, 1995).

The ambient low temperature of metamorphism in the Brunswick complex promoted strain localisation. Strain localisation is particularly obvious in D_1 shear zones developed in the volcanic rocks where unfoliated low-strain pods locally occur adjacent to high strain tectonites (van Staal, 1994, fig. 4a,b). Shear zones commonly are important channels for upward moving fluids derived by devolitalisation processes in subduction complexes (e.g. Bebout and Barton, 1989; Selverstone et al., 1991), particularly since steady-state subduction and underplating of hydrous rocks (e.g. water-bearing sedimentary rocks) provides a continuous source of fluid to the deforming rock pile above it. Fluid infiltration may lead to replacement reactions facilitated by ionic transfer along chemical potential gradients, which may ultimately weaken the rocks sufficiently for subsequent increments of deformation to become further

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PII: S0191-8141(00)00163-2

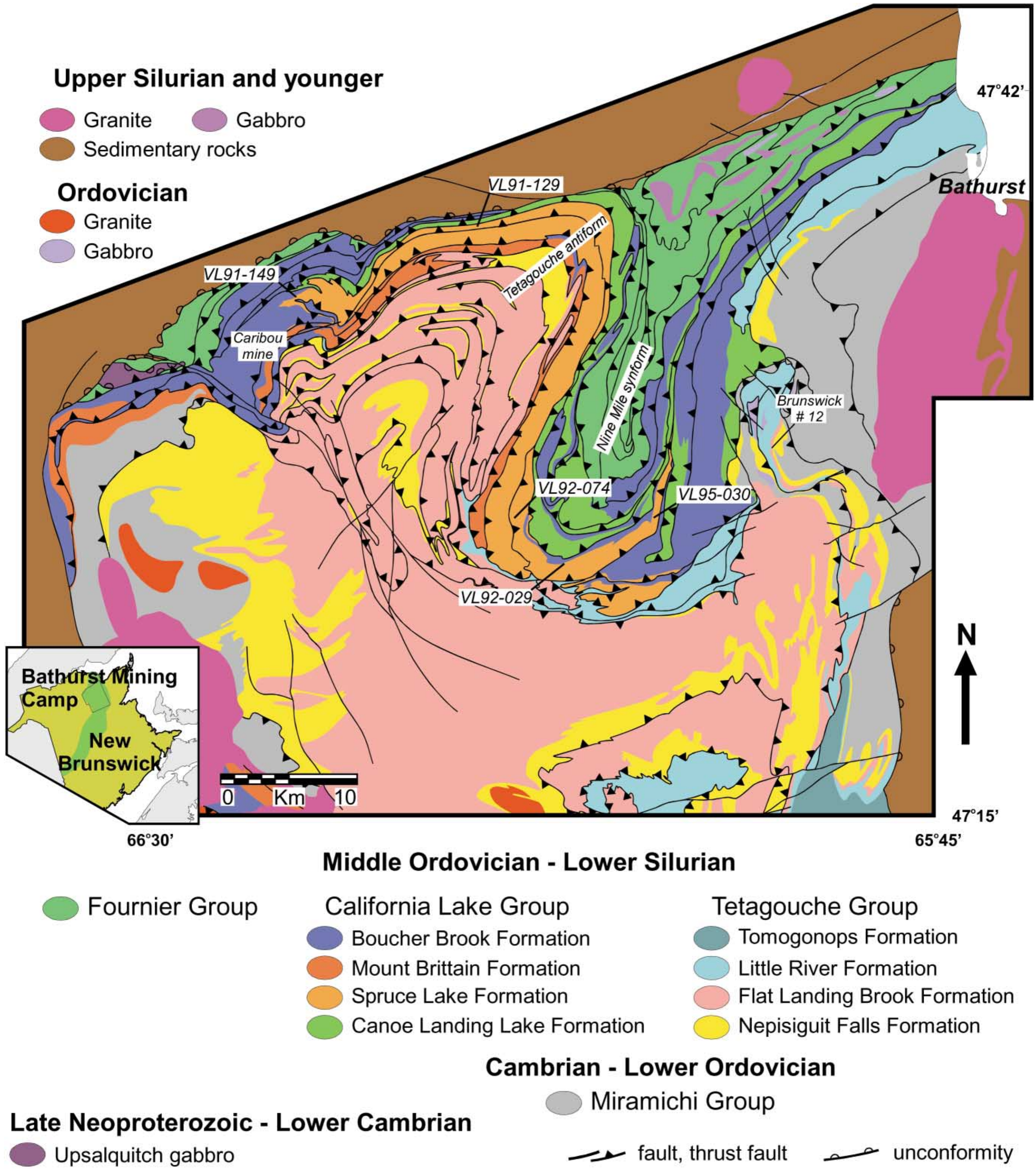

Fig. 1. Simplified geology of part of the Brunswick complex based on recent mapping and the new, high-resolution magnetic and electromagnetic survey of northern New Brunswick (van Staal and Rogers, 2000). The location of samples discussed in Fig. 10 and in the text are included except for VL91-005, VL91-010, VL91-013, VL91-015, VL91-041, VL91-046, VL91-054 and VL91-058, which all come from the drill-core intersecting the hanging wall of the Caribou massive sulphide deposit (see Appendix A).

localised within the altered rocks (i.e. reaction softening; White and Knipe, 1978; Wintsch et al., 1995). Aqueous fluids may also cause hydrolytic weakening of quartz and feldspar (Tullis and Yund, 1980; Tullis et al., 1996) and can also promote strain localisation. We propose that syn-tectonic alkali-metasomatism, particularly the albitisation of feldspar had a marked weakening effect on the felsic volcanic rocks.

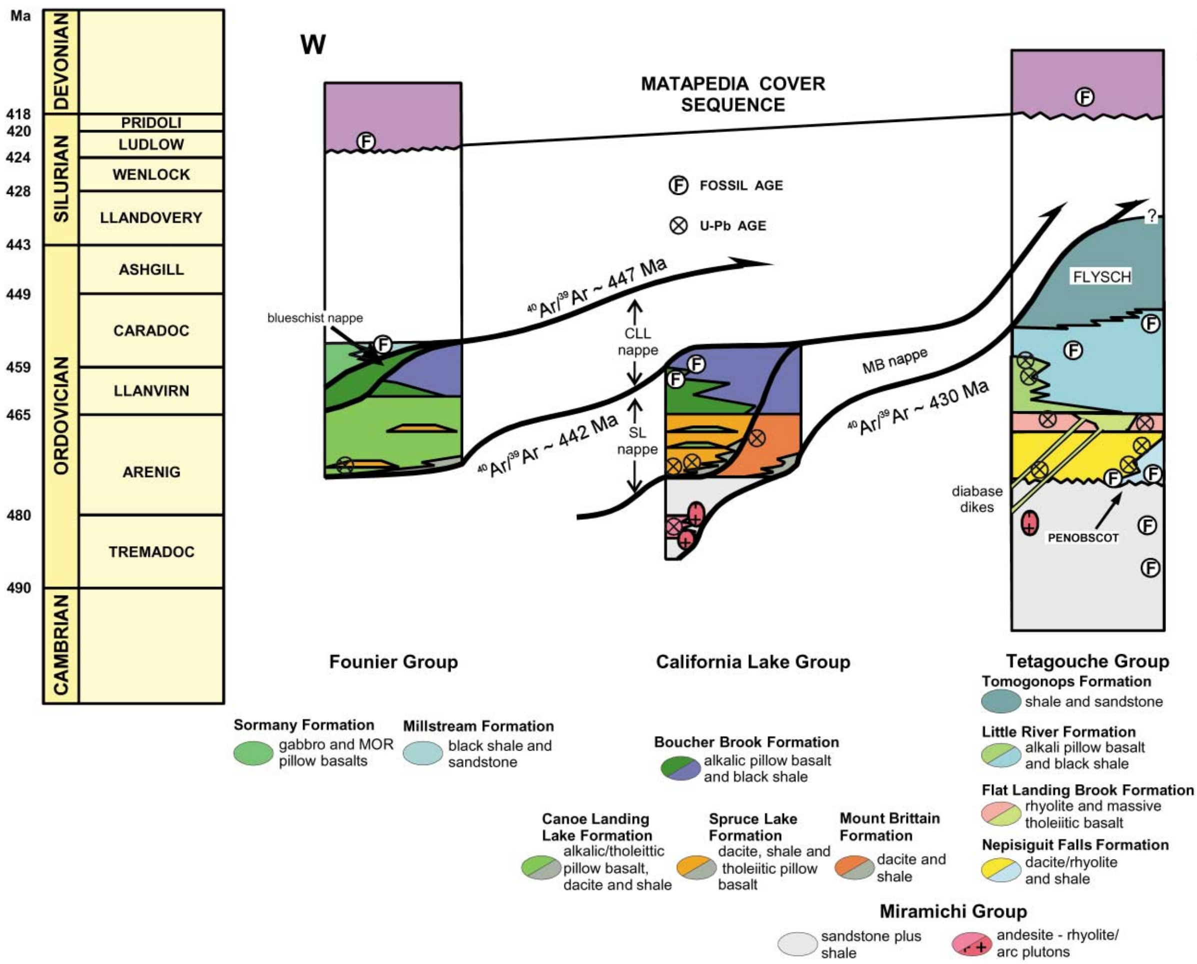

Fig. 2. Schematic tectonostratigraphy of the Ordovician and Silurian rocks of the Brunswick complex (van Staal and Rogers, 2000). The ^{40}Ar/^{39}Ar ages are for D_1 phengites selected from representative tectonites of each nappe.

2. Geological setting

The structures discussed here developed in variably deformed and metamorphosed dacitic to rhyolitic, feldsparphyric lava flows and/or cryptodomes that, together with interlayered shale and tholeiitic pillow basalts form part of the middle to upper Arenig (c. 472–468 Ma) Spruce Lake Formation of the California Lake Group in the New Brunswick Appalachians (van Staal and Rogers, 2000). The rocks of the California Lake Group, together with the volcanic and sedimentary rocks of the Fournier and Tetagouche groups, were incorporated into the southeast-facing Brunswick subduction complex (Fig. 1) by progressive underplating to depths corresponding with the blueschist, pumpellyite–actinolite, or high-pressure greenschist facies beneath an accretionary wedge (van Staal et al., 1990; van Staal, 1994). The underplated volcanic fragments constitute both relatively buoyant, rifted arc edifices and seamounts, which have compositions similar to the volcanic rocks of the present-day Japan Sea (van Staal et al., 1991). During underplating, the rocks were imbricated into numerous thrust-bound nappes, with each major nappe distinguished and generally named after its distinct volcanic unit and/or metamorphic history (Fig. 2).

The Spruce Lake nappe is a coherent, but internally imbricated thrust sheet that emplaced the volcanic rocks of the Arenig Spruce Lake Formation and its stratigraphic cover of the Llanvirn–Caradoc Boucher Brook Formation (Fig. 2) above coeval or younger rocks of the Mount Brittain and Strachens Lake nappes (Figs. 1–4). The Spruce Lake nappe is structurally overlain, depending on location, either by the blueschist nappe or the Canoe Landing Lake nappe.

The Spruce Lake Formation was chosen for this study to eliminate, as much as possible, ambiguities concerning the potential control imposed by inherited zones of weakness, such as pyroclastic flow structures and/or pre-tectonic zones of seafloor hydrothermal alteration on the localisation of the earliest strain increments. The studied felsic volcanic rocks occur stratigraphically above the syngenetic massive sulphide deposits of the Spruce Lake Formation (Figs. 3 and 4; van Staal and Rogers, 2000). Generally, the footwall,

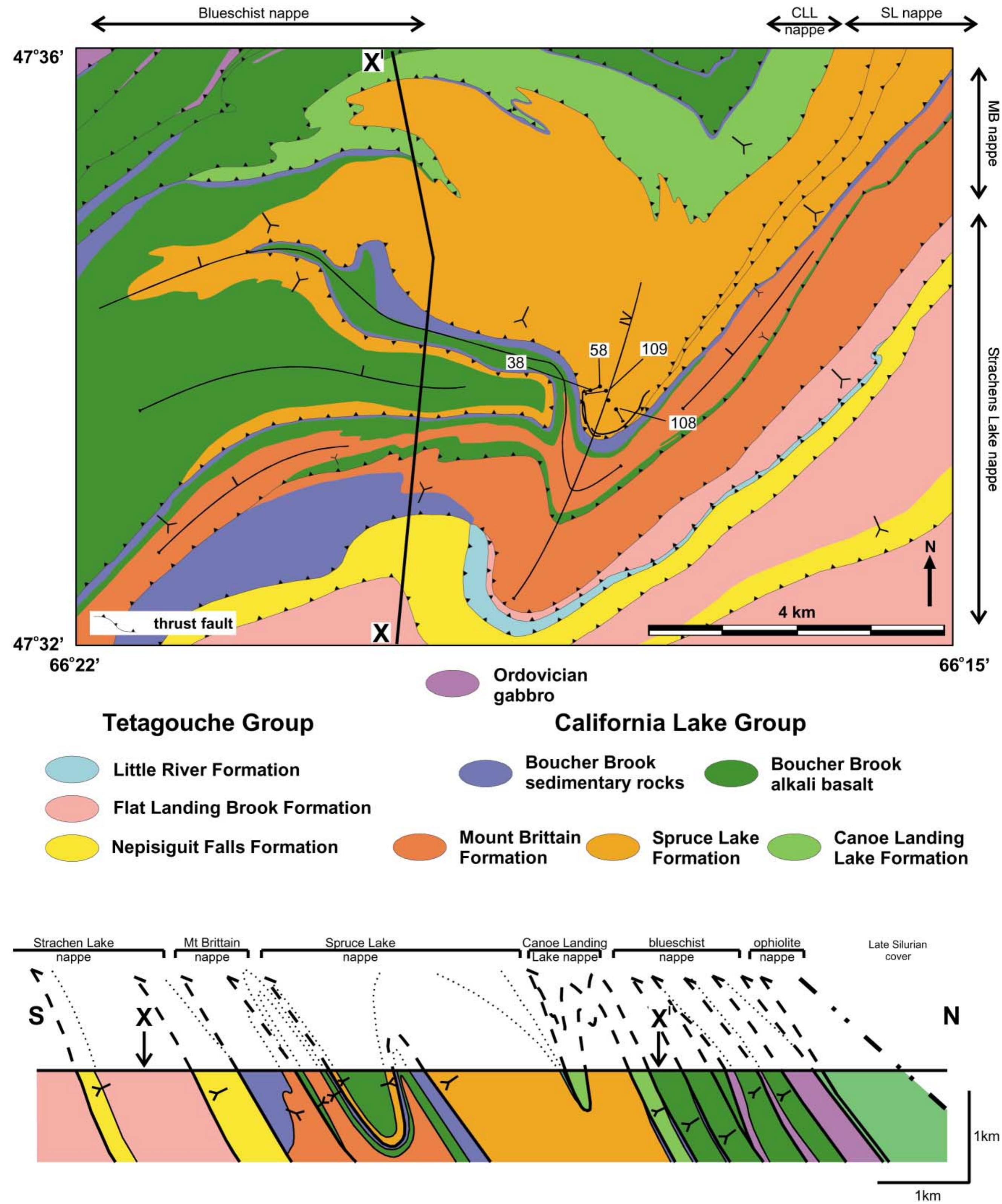

Fig. 3. Simplified geological map of the Spruce Lake and adjacent nappes in the vicinity of the Caribou massive sulphide deposit. Relative ages of the structures are outlined by the Roman numerals on the axial traces. The F_{1b} folds, which are outlined by Roman numeral I on the axial traces, plunge moderately towards the west. More information concerning the structural data (foliations and lineations) of the same map area is given in van Staal (1994). The profile is schematic, but is constructed with the aid of geophysical and all available structural data. Only part of the profile between X and X^1 is pertinent to the map. The Strachens Lake nappe incorporates rocks of the Tetagouche Group. CLL = Canoe Landing Lake nappe; SL = Spruce Lake nappe; MB = Mount Brittain nappe. Surface projections of the drill-holes listed in Appendix A are shown.

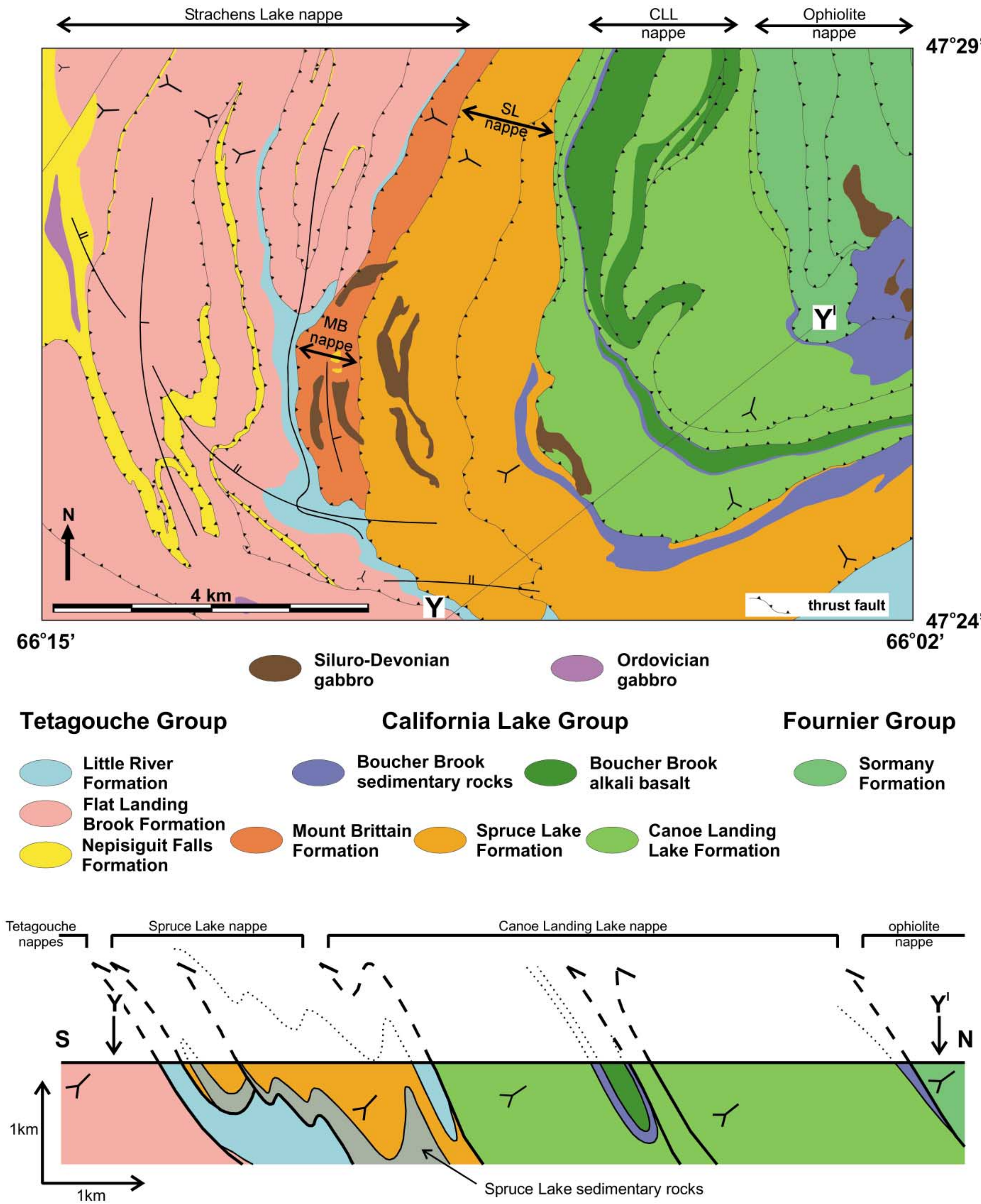

Fig. 4. Simplified geological map of the Spruce Lake and adjacent nappes in the hinge of the F_4 Nine Mile synform. Structural data belonging to this map are given in van Staal (1994) and van Staal and de Roo (1995). Note that there are at least two generations of thrusts. The oldest generation of thrusts are folded by northerly plunging F_{1b} folds and truncated by the younger thrusts, which themselves are folded by at least F_2. Only part of the profile between Y and Y^1 is pertinent to the map. For abbreviations see Fig. 3.

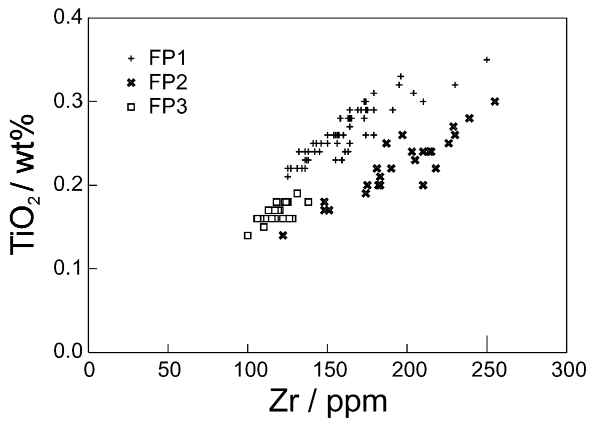

Fig. 5. TiO_2–Zr discrimination plot illustrating the similar but distinct compositions of the various feldspar-phyric felsic volcanic suites of the Spruce Lake Formation (referred to herein as FP1, FP2 and FP3). The trend developed by the samples of each suite predominantly reflects the effects of mass-transfer on the absolute abundances of these elements.

but not the hanging wall, of such deposits shows evidence of synvolcanic hydrothermal alteration (Lentz and van Staal, 1995; Lentz et al., 1997). Moreover, preliminary stable isotope data on deformed basalts in the immediate structural hanging wall of the Spruce Lake nappe (van Staal et al., 1999) provides some constraint on the fluid-rock evolution during subduction and metamorphism of the Spruce Lake volcanic rocks.

The metamorphic M_1 assemblages in the mafic rocks of the Brunswick complex are essentially identical to those in the low-temperature metabasalts of the Sanbagawa subduction complex of Japan (e.g. Toriumi, 1975; Otsuki and Banno, 1990). Metamorphic conditions of the Spruce Lake and bounding nappes varied between c. 600–800 MPa and 330–370°C (van Staal et al., 1990; Currie and van Staal, 1999), consistent with the presence of pumpellyite and stilpnomelane and the absence of biotite in the felsic volcanic tectonites.

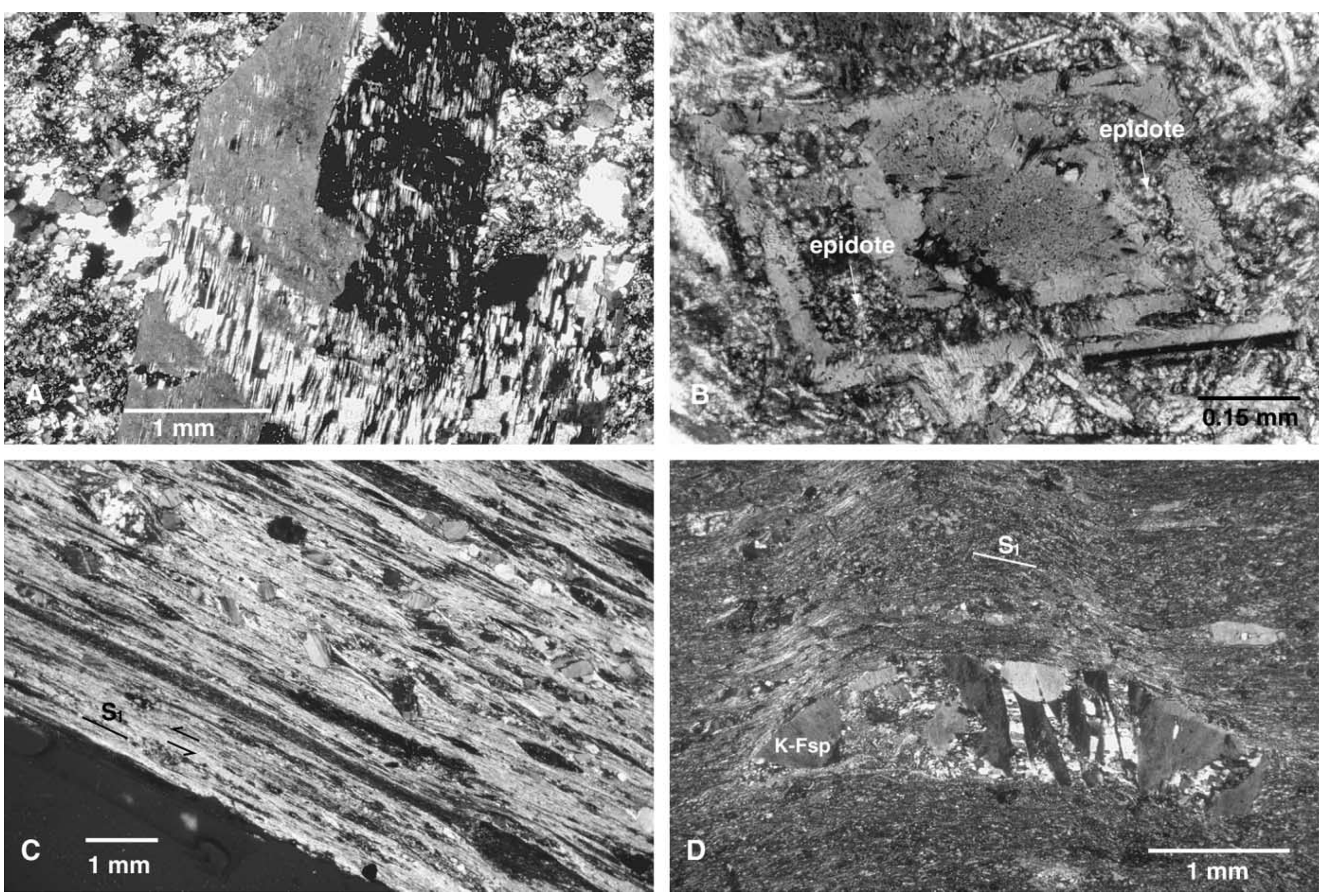

Fig. 6. (a) Vein cutting through low strain Spruce Lake dacite with preserved felsitic groundmass. Note lack of sericitisation, and the presence of chessboard albite in patch-perthite and where the vein cuts the alkali feldspar phenocryst. (b) Nearly euhedral albite pseudomorph after a plagioclase phenocryst in mafic volcanic rock interlayered with the felsic volcanic rocks. Relict zoning is outlined by epidote bands. (c) Spruce Lake phyllonite with a train of rotated feldspar porphyroclasts and asymmetrical tails indicating a reverse sense of motion. (d) Segmented K-feldspar porphyroclasts with quartz fibres sealing intergranular fracture in Spruce Lake felsic volcanic tectonite. Note that cigar-shaped feldspar is microscopically untwinned and apparently modified by solution transfer where it is in contact with S_1 mica seams.

Table 1
Representative compositions of feldspars determined with the electron microprobe from the felsic volcanic rocks of the Spruce Lake, Strachens Lake and Canoe Landing Lake nappes

Location		Spruce Lake nappe								Strachens Lake nappe	Canoe Landing Lake nappe	
Sample		VL91-080	VI91-129	VL91-140	VL92-029	VL92-065	VL92-066	VL92-074	VL92-091	VL96-016	VL96-020	VL96-022
Plagioclase	Ab	95.93	98.67	99.06	99.18	99.10	97.69	97.98	98.69	98.90	98.97	99.34
	An	0.35	1.06	0.31	0.46	0.49	0.49	1.50	0.62	0.40	0.12	0.15
	Or	3.71	0.27	0.64	0.13	0.41	1.62	0.47	0.62	0.55	0.57	0.51
K-feldspar	Ab	3.35		1.61		2.52	2.06		2.07	2.42	2.22	3.46
	An	0.37		0.02		0.00	0.00		97.73	96.73	0.00	0.23
	Or	96.27		98.12		97.12	97.19		0.10	0.00	97.78	96.31
	Ce			0.24		0.37	0.75		0.10	0.86	0.00	0.00

3. Characteristics of the Spruce Lake nappe felsic volcanic rocks

The Spruce Lake Formation contains three suites of feldspar–phyric dacitic to rhyolitic metavolcanic rocks (FP1, FP2 and FP3) that can be distinguished by petrography and chemistry, with each suite having similar, but distinct trace element compositions (Fig. 5; Rogers, 1994; 1995). Where strain is low or moderate, the felsic volcanic rocks of all three suites contain euhedral to subhedral, unbroken crystals up to 2 cm long of alkali–feldspar and albite in a fine-grained felsitic, devitrified or locally spherulitic quartzo–feldspathic groundmass (Fig. 6a). This texture, granophyric intergrowths, primary twins according to the Carlsbad and Baveno laws, typically having irregular, stepped boundaries, and embayments, indicate that the large albite and alkali-feldspar grains are pseudomorphs after phenocrysts (Vernon, 1965; Smith, 1974). Where strain is high, deformation has transformed the large feldspar crystals into porphyroclasts. Porphyroclasts developed from alkali-feldspar crystals are generally partially or completely replaced by chessboard albite.

Electron microprobe analyses of the feldspars sampled throughout the felsic volcanic rocks of the Brunswick complex (Table 1) revealed that all feldspars, irrespective of the amount of strain in the rocks, have albite and K-feldspar compositions inappropriate for dacitic to rhyolitic volcanic protoliths (Carmichael et al., 1974; Brown and Parsons, 1994). Detailed X-ray diffraction (Juras, 1981; Nelson, 1983) and U-stage optical studies (van Staal, 1985) carried out on coeval felsic volcanic rocks near Brunswick No. 12 mine (Fig. 1) have also shown that the alkali-feldspar crystals are generally pseudomorphed by microcline. However, mixtures of triclinic and monoclinic, and domains of optically untwinned, intermediate perthitic microcline, orthoclase, or both ($2V_\alpha = 60–74°$), displaying a 'tweed'-like texture, are locally preserved. Preferential growth of pumpellyite, clinozoisite or epidote in albite pseudomorphs after plagioclase phenocrysts in both mafic and felsic volcanic rocks (Fig. 6b), suggests that the original plagioclase phenocrysts contained more calcium prior to metamorphism. Considering the dacitic to rhyolitic composition of the Spruce Lake Formation rocks, the plagioclase was probably originally oligoclase (Carmichael et al., 1974; Brown and Parsons, 1994).

4. Evidence for a progressive D_1 deformation

Structural and stratigraphic evidence for the existence of a nappe stack has been presented previously (e.g. van Staal, 1994; van Staal and Rogers, 2000). Structural cut-offs in the footwall of repeated sections with older rocks structurally emplaced above younger rocks are the primary evidence (Figs. 3 and 4). The thrusts are represented by D_1 ductile shear zones, which are characterised by a very well

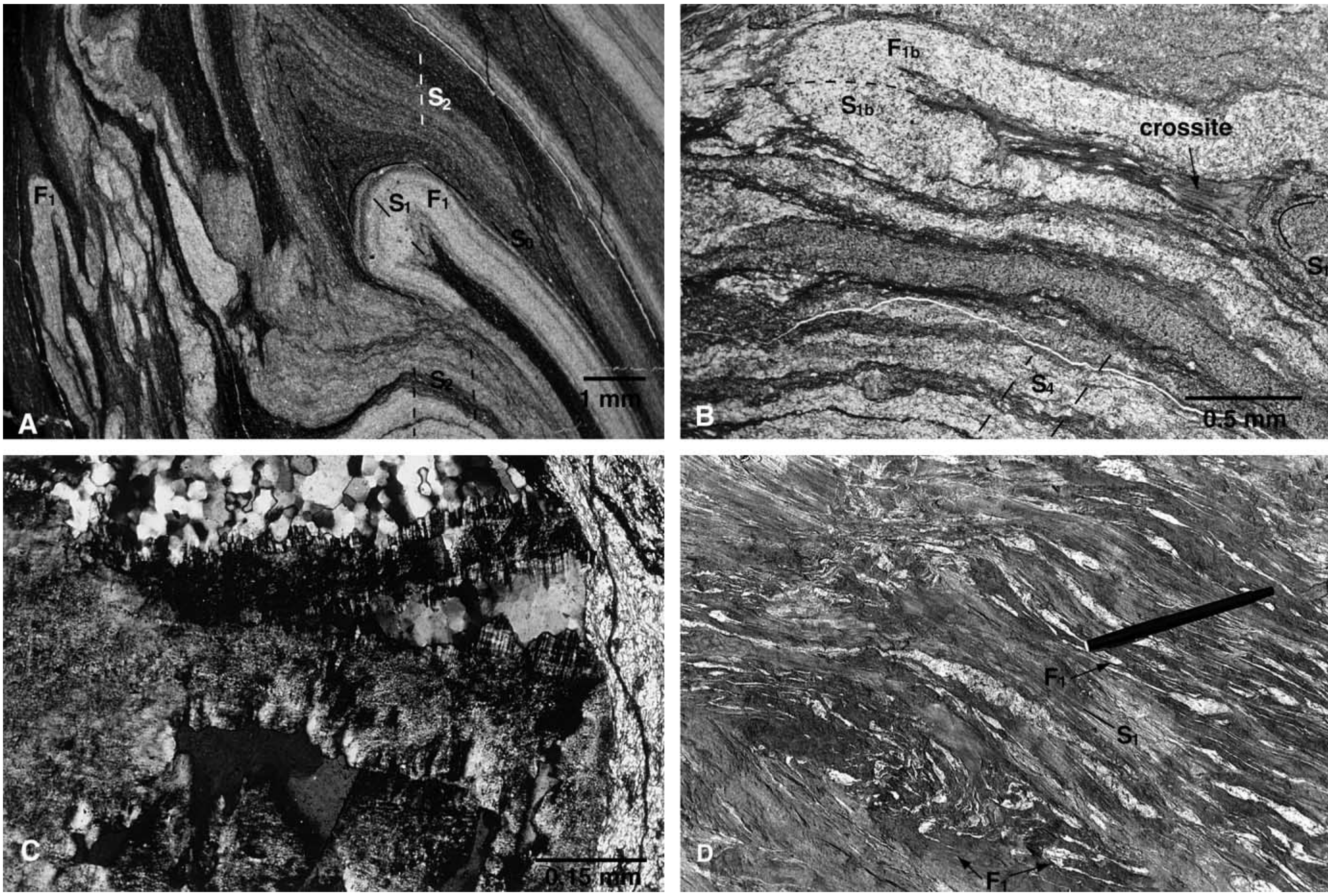

Fig. 7. (a) Refolding of microscopic F_1 fold in the hinge region of a F_2 fold in metasedimentary rocks interlayered with the volcanic rocks. Note that F_1 folds a bedding parallel foliation (S_0) and also has a new crenulation cleavage in the hinge region (S_1), together defining a composite foliation. The bedding foliation is tectonic and not a metamorphically enhanced sedimentary/diagenetic structure because it can be traced without interruption into deformed volcanic flows and associated subvolcanic intrusives. (b) Phyllonite at the base of the blueschist nappe with a F_{1b} fold in an albite bearing vein and S_1 defined by crossite, chlorite and epidote. Crossite is crenulated in the hinge region of F_{1b} and is accompanied by new growth of crossite parallel to the axial surface of the crenulations (S_{1b}). F_2, F_3 and F_4 overprint S_{1b} and F_{1b}. (c) Maximum microcline overgrowth ($2V_\alpha = 88 \pm 2°$; $X \wedge (010) = \pm 18°$) on segmented intermediate ordered K-feldspar ($2V_\alpha = 74 \pm 2°$). (d) High strain D_1 phyllonite with completely dismembered vein complex in the basal shear zone of the Spruce Lake nappe. Note numerous rootless intrafolial F_1 folds of quartz-albite–K-feldspar veins.

developed L/S fabric (Fig. 6c and d). Kinematic indicators in the D_1 shear zones are consistent with a reverse sense of shear towards the east or southeast (Fig. 6c). Where there is a nearly continuous three-dimensional outcrop, such as in the mines, large sheath folds with complicated geometries occur within the wider shear zones (e.g. van Staal and Williams, 1986; de Roo et al., 1991; Park, 1996). Both the shear zones and the enclosed sheath folds were subsequently refolded by upright to steeply inclined folds with dominantly shallow to moderate plunges (Figs. 3 and 4); hence, the D_1 shear zones that accommodated the thrust translations probably, at least in part, had shallow to moderate dips prior to the refolding.

Removal of the deformation that post-dates underplating and high-pressure metamorphism (for details see van Staal and de Roo, 1995) shows that the rocks in each thrust sheet generally young upwards in a northerly direction (van Staal, 1994), whereas the sedimentary cover of the volcanic rocks becomes progressively younger towards the southeast (Fig. 2). This suggests that underplating progressed from northwest to southeast, and formed a series of southeastward propagating duplexes. This interpretation is supported by $^{40}Ar/^{39}Ar$ analysis of the low-temperature (<400°C) M_1 phengites from the major thrust nappes (Fig. 2), which suggest that D_1 is diachronous and becomes progressively younger in the structurally deeper nappes. Some of the major thrust sheet boundaries also coincide with minor, but consistent jumps in metamorphic grade, however, such that a partially inverted sequence is created with higher-pressure rocks overlying lower-pressure rocks. Hence, some of these major nappe-bounding shear zones represent 'out-of sequence' thrusts, which must have remained active after peak metamorphism of their hanging wall rocks.

Detailed new mapping combined with a recent low altitude geophysical survey (c. 50 m with flight-line spacing of 200 m) of the Brunswick subduction complex (Geological Survey of Canada, Open File 3294, 1996), improved the resolution of critical magnetic (e.g. magnetite-bearing

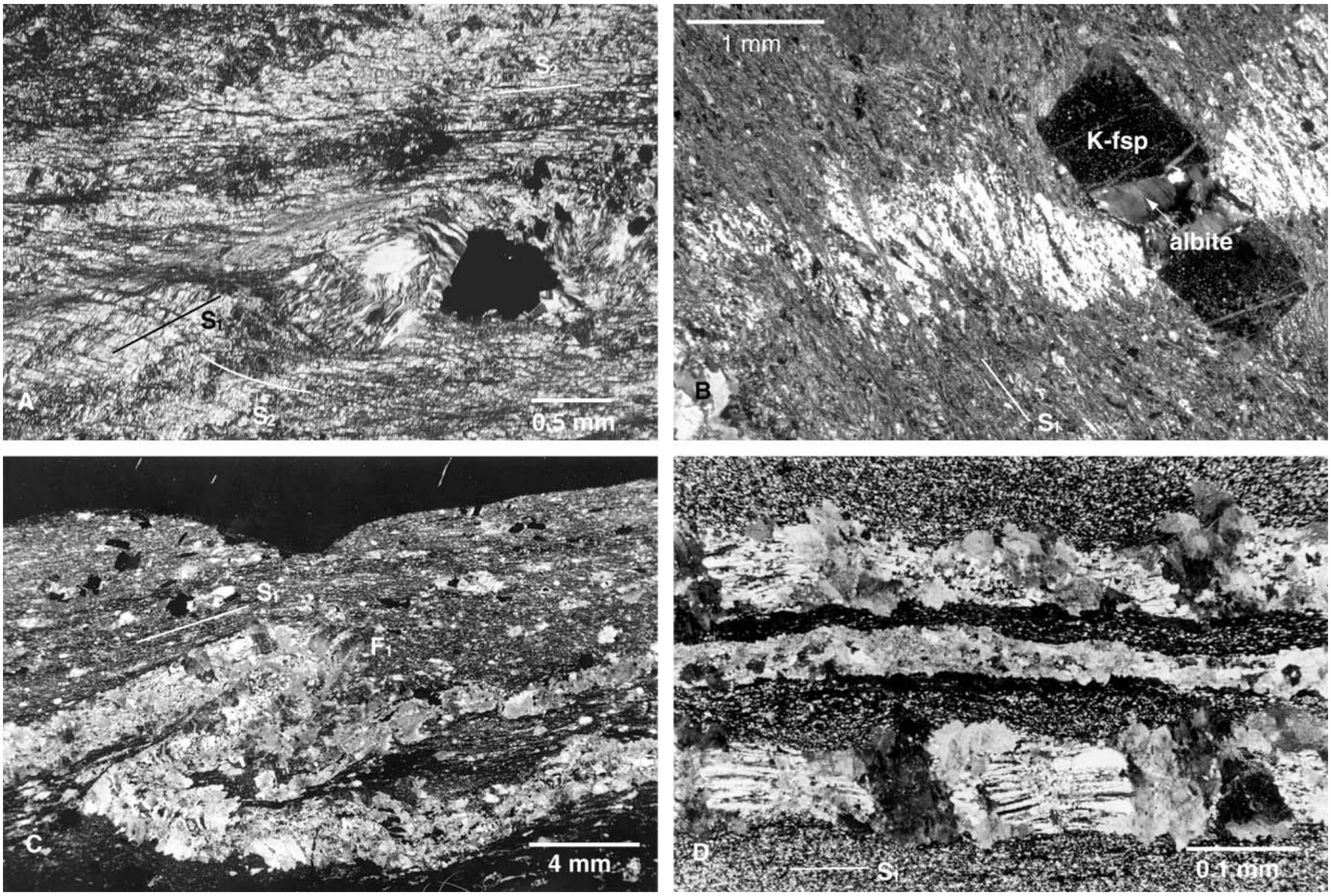

Fig. 8. (a) Strain fringe around pyrite in a phyllonite crenulated by F_2. Note F_2 microfolding of quartz fibres in the strain fringe. (b) F_1 folded, strongly recrystallised quartz-albite vein in a Spruce Lake tectonite. Chessboard albite fibres seal boudinaged K-feldspar phenocryst where intersected by the vein. Note that albite in the vein is strongly recrystallised, except for the albite fibres protected by the phenocryst. (c) F_1 intrafolial fold in an ankerite vein that was first stretched and rotated into parallelism with S_1 before folding since quartz fibres (d) that seal the boudinaged ankerite vein are also folded by F_1 in the hinge of the fold. This sample is a good illustration of the progressive nature of D_1.

basalts) and conductive markers (e.g. graphite and sulphide-bearing shales). This new data supported earlier interpretations that the D_1 thrust-related deformation produced several generations of structures (e.g. van Staal, 1994; Park, 1996). Microstructural studies show that the main foliation (S_1) in all rocks is a composite foliation (Fig. 7b), defined by minerals (e.g. sodic amphibole and phengite) that are characteristic of the subduction-related metamorphism (M_1). S_1 is both axial planar to, and folded by, F_1 folds (Fig. 7a and b). D_1 locally comprises two generations of folds (F_{1a} and F_{1b}) on a mesoscopic scale (van Staal, 1987, fig. 3) whereas, the new mapping suggests several generations of thrusts. Some thrusts were initially folded and then cut by other, younger thrusts (Fig. 4). Both sets of thrusts are folded by or associated with tight to isoclinal, generally s-shaped, upright folds previously interpreted as F_2 structures (de Roo and van Staal, 1994; van Staal and de Roo, 1995, fig. 5). These folds, however, have the same style and associated microstructures as the folds truncated by the youngest generation of thrusts, and an axial plane cleavage defined by minerals typical of M_1. Moreover, large, *z*-asymmetric F_2 folds (Fig. 4) locally refold these structures. We refer to these structures herein as F_{1b} since they refold the earlier, less common intrafolial F_{1a} sheath folds. The upright F_{1b} folds, together with rare sinistral transcurrent shear zones, have been interpreted to have formed in response to continuous sinistral oblique convergence (van Staal, 1994; van Staal and de Roo, 1995).

5. Microstructural evidence for fluid-assisted deformation and fluid infiltration

The ubiquitous presence of a differentiated layering in all tectonites and the large volume of metamorphic veins in the tectonites and shear zones (van Staal and Williams, 1984; de Roo and Williams, 1990; Park, 1996) suggest that solution-transfer creep (Durney, 1972; Williams, 1972) was an important deformation mechanism during D_1. Solution-transfer refers to a deformation process in which the strain in the rock is primarily achieved by mass transport of dissolved matter via a pore and/or grain boundary fluid; the intergranular strain compatibility is maintained by non-uniform grain boundary sliding (Paterson, 1995). The efficiency of such a deformation mechanism depends on the degree of connectivity between the grain boundary fluid films and the fluid present in the pore spaces. Connectivity and also fluid flow may be enhanced significantly by the continuous opening of microfractures (Etheridge et al., 1984; Williams, 1990). An important question is whether the solution-transfer redistributes material merely on a local scale (closed system) within the rocks or whether material is moved in or out of the rocks on a regional scale (open system).

Feldspar and quartz crystals commonly display dissolution structures where they are in contact with the foliation septa (Figs. 6d and 7c). Strain shadows, strain fringes (Fig. 8a), overgrowths and intragranular fractures are potentially local precipitation sites for some of the dissolved material. The strong spatial association of vein complexes, however, which may constitute more than 50% of the surface area of individual outcrops, with the shear zones (Fig. 7d) throughout the Brunswick complex, suggests fluid channelling. Fluid channelling commonly is accompanied by metasomatism caused by fluid-assisted mass transfer (O'Hara, 1988; Selverstone et al., 1991; Dipple and Ferry, 1992; Streit and Cox, 1998), particularly in a subduction zone setting (Bebout and Barton, 1989). Mass balance and isotope studies of relatively mildly strained felsic volcanic tectonites in the Brunswick complex indicate that even the rocks outside the shear zones were deformed in an open system with high fluid-rock ratios (Lentz, 1999).

The vein mineralogy and fillings between the grain segments mainly comprises fibrous quartz and feldspar (Figs. 6d, 7c and 8b). Albite is particularly common in most veins, independent of the composition of the host rock. The presence of elongated M_1 minerals such as phengite, stilpnomelane, Ca- or Na-amphibole, Na-pyroxene, titanite, pumpellyite, and epidote-group (zoisite, clinozoisite, and epidote) (Fig. 8a) with their long axes parallel to L_1 (van Staal et al., 1990, fig. 4) suggests growth during D_1, consistent with the folding of most veins, shadows and fringes, by F_1 or F_2 (van Staal, 1994, fig. 4e; Park, 1996). Moreover, routine microprobe and scanning electron microscope analyses of the minerals from the veins, segment fillings, and foliation planes have not yielded any systematic or significant differences in composition with respect to those in the host rock. The vein mineral assemblage, thus, generally appears to have achieved equilibrium before the end of M_1 with all or part of the surrounding rock. Intense folding and boudinaging of most veins, such that they now lie within the composite S_1 foliation, suggests that veining started during the earliest increments of D_1. Fibrous minerals such as quartz and feldspar are partially or completely transformed into elongated mineral aggregates in such veins (Fig. 8a–d). The highly deformed veins are commonly cut by less deformed veins, which are discordant to S_1 and have partially or completely preserved fibrous textures (Park, 1996), yet are folded by F_2. These late D_1 veins contain minerals typical of the peak-temperature metamorphism (e.g. stilpnomelane; Currie and van Staal, 1999, fig. 2b) or retrograde minerals like vermiculite (van Staal, 1985). The crosscutting relationships, mineralogy, and the variable state of strain suggest that veins formed continuously or at several stages during D_1, consistent with the progressive nature of D_1.

Where the fibrous structure is relatively well preserved, intrafibre cracks (Fig. 9a) suggest a crack-seal growth history (Ramsay, 1980; Cox and Etheridge, 1983; de Roo and Williams, 1990) for at least these veins. A crack-seal mechanism implies transient build-ups of high fluid pressures that facilitate the cyclic tensile failure in an overall ductile shear zone. This, and the complex, long history of veining, suggest some degree of fluid replenishment as a result of fluid infiltration (Etheridge et al., 1984; Fisher, 1996) in the shear zones.

6. Evidence for fluid channelling from oxygen isotope data

Channelled fluid flow is independently indicated by the shifts in the oxygen and hydrogen isotope compositions within D_1 shear zones in the blueschist nappe immediately above the Spruce Lake nappe (Figs. 1 and 3). Relatively weakly-strained, low grade (actinolite–pumpellyite/ greenschist facies) metabasalts closely associated with, and preserved as, pods in the highly strained blueschist nappe have heavy $\delta^{18}O$ values near 10‰ and low δD values of c. −80‰. This indicates interaction with a low-temperature fluid during the earliest stages of subduction or a pre-tectonic, low-temperature seafloor alteration of the basalts (e.g. Taylor and South, 1985). In contrast, the blueschist tectonites show marked oxygen isotope depletion (Fig. 10) with respect to these low-strain metabasalts, indicating extensive re-equilibration with an isotopically homogeneous, higher temperature fluid at high fluid/rock ratios. The low-strain metabasalts show a trend of re-equilibration from low-temperature altered compositions to the field occupied by the blueschists, suggesting progressive isotopic re-equilibration of the basalts during prograde metamorphism and metasomatism accompanying progressive underthrusting. Isotopic equilibration with relatively high-temperature fluids culminated with formation of the blueschist tectonites as a result of high fluid/rock ratios in the shear zones during underplating.

The presence of titanite, pumpellyite and epidote group minerals in rocks and veins of the Spruce Lake nappe and elsewhere indicate that the metamorphic fluid was essentially devoid of CO_2 and dominantly aqueous (van Staal,

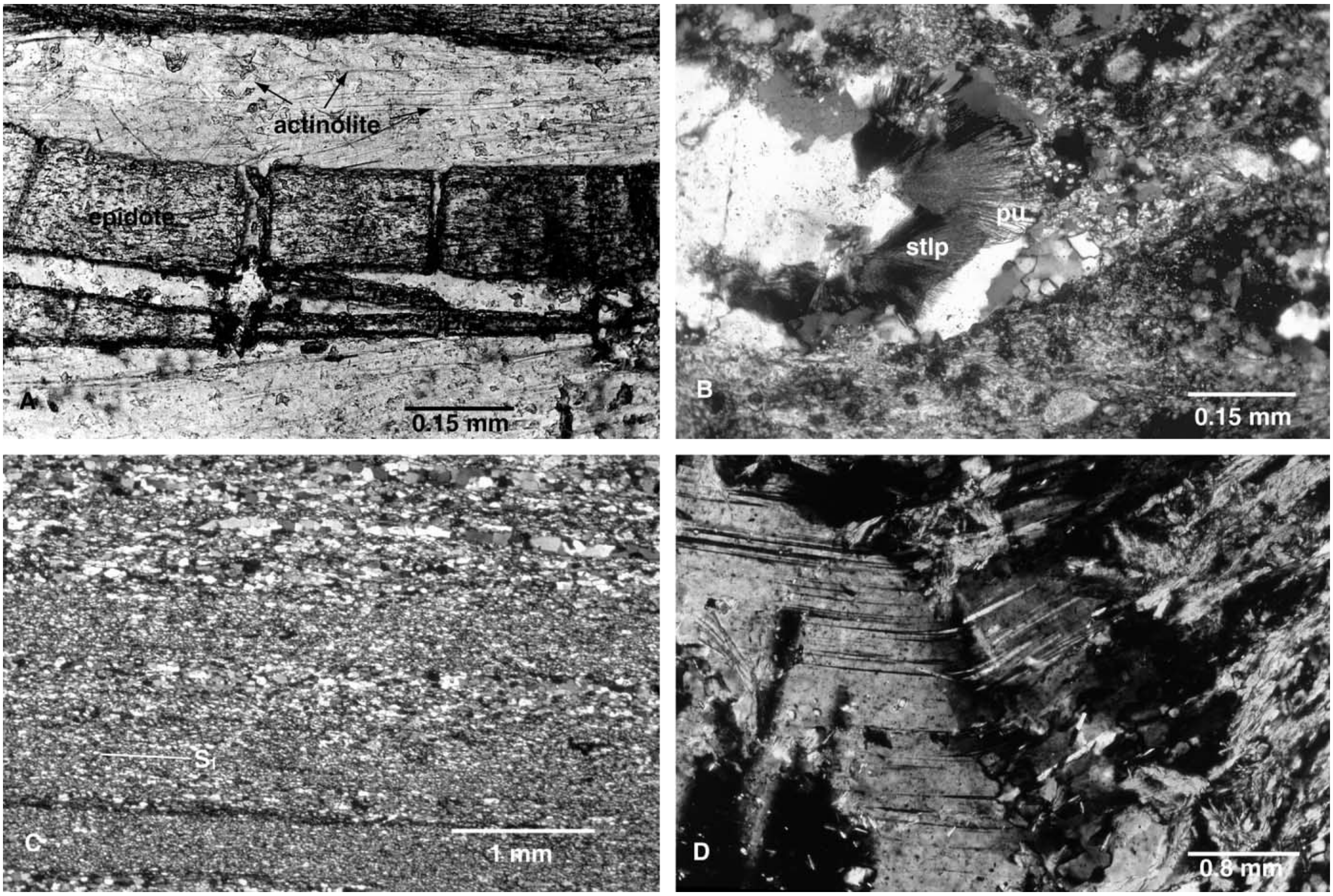

Fig. 9. (a) Close-up of boudinaged epidote fibres in epidote, sodic actinolite, albite, quartz and stilpnomelane vein in metabasalt. With the exception of stilpnomelane, these minerals are also pervasive in the host metabasalt. Quartz and bluish green, sodic actinolite needles seal the cracks. Note dark zones perpendicular to length of epidote crystals that likely represent healed cracks. Stilpnomelane replaced other mafic minerals along the rims of the vein and in a c. 3 cm-wide halo around the vein (van Staal, 1985). Vein is mildly deformed by D_1 and folded by F_2 on outcrop scale, suggesting the fluids responsible for growth of stilpnomelane were introduced relatively late during D_1. (b) Colourless pumpellyite sheafs and needles, partially pseudomorphed by stilpnomelane in deformed and recrystallised vein in Spruce Lake tectonite. (c) Laminated, quartz-albite ultramylonite (VL91-129) developed from dacite in the roof-thrust-related shear zone of the Spruce Lake nappe. Mylonite contains a small amount of fish-shaped phengite. (d) Kinked albite porphyroclast with lenticular deformation twins in high strain tectonite. Note bulging and new grains along part of the kink-band boundary.

1985). The whole-rock isotopic data on basalts discussed above suggest metamorphic fluids with $\delta^{18}O$ of approximately 3‰ at c. 350°C (Currie and van Staal, 1999), based on feldspar-water fractionation factors (O'Neil and Taylor, 1967). High water/rock ratios are implicated by the narrow range of $\delta^{18}O$ shown in Fig. 10 for widely spaced samples. Such channelled fluids largely comprised isotopically evolved seawater, and evidence the major egress of marine pore fluids during subduction and accompanying prograde metamorphism of the associated sedimentary rocks.

7. Feldspar microstructure in the shear zones

Recent structural studies in the Brunswick complex have emphasised the presence of D_1-related quartzo-feldspathic mylonites (Figs. 9c and 11b) and phyllosilicate-rich mylonites (phyllonites, Fig. 6c) in the felsic volcanic rocks (e.g. van Staal, 1987, 1994; de Roo and Williams, 1990; Park, 1996). Where the transition from moderately strained felsic tectonites into mylonitic rocks is exposed in detail, the boudinage and grain-size reduction of feldspar porphyroclasts increases progressively towards the mylonites and phyllonites, culminating in ultramylonites lacking porphyroclasts. Extensions of at least 1500% are recorded by segmented feldspar porphyroclasts (e.g. de Roo and Williams, 1990). Although the extensions displayed by the boudinaged porphyroclasts provide only rough order of magnitude estimates of the finite bulk strain, the relative ordering of the strain in rocks by this method for our mass balance calculations below is probably reliable (Figs.12 and 13).

The low-grade metamorphic conditions experienced ($\geq$600 MPa with $T < 400$°C) by all of the rocks raises the question of how the mylonites formed, particularly the

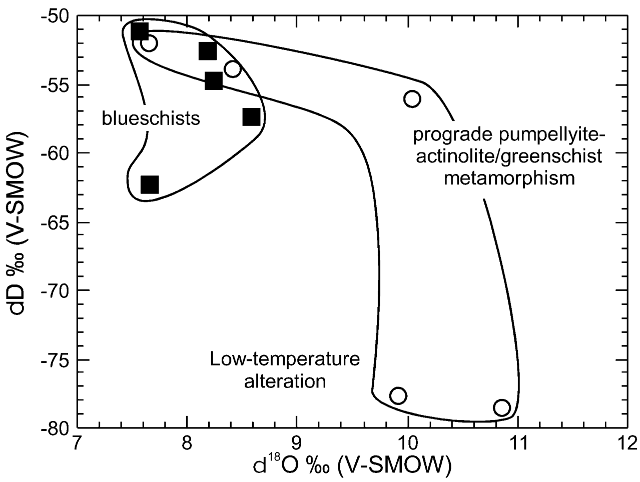

Fig. 10. Plot of whole-rock δD versus $\delta^{18}O$ for metabasalts occurring as relatively weakly strained pods within the blueschist tectonites or immediately adjacent to the blueschist nappe. Data illustrates a marked isotopic shift from metabasalts that preserve the imprint of a low temperature alteration to highly strained blueschists via a prograde, low temperature–high pressure metamorphism and accompanying metasomatism. Metasomatism under increasing fluid/rock ratios of isotopically and chemically modified seawater expelled from the underthrusted sedimentary rocks facilitated formation of the blueschists.

mechanism by which the feldspar porphyroclasts were reduced in grain-size, and why some feldspar grains were replaced by phyllosilicates and others not. This is pertinent, because there is a large consensus that dynamic recrystallisation of feldspar only becomes possible at upper greenschist facies or higher conditions (>450°C; e.g. Voll, 1976; Simpson and Wintsch, 1989). On the other hand, de Roo and Williams (1990) argued that the low-temperature quartzo–feldspathic mylonites that they studied in the Brunswick complex mainly formed by rotation recrystallisation of quartz and feldspar. The intracrystalline deformation of quartz and feldspar ('dry' deformation of de Roo and Williams (1990)) cyclically alternated with stages of 'wet' deformation dominated by solution transfer.

We concur with the observations of de Roo and Williams (1990) that the volcanic tectonites formed by a combination of solution transfer and intracrystalline deformation mechanisms in quartz and feldspar, irrespective of whether they alternated cyclically or not. We are, however, not convinced that subgrain rotation was the dominant process during dynamic recrystallisation. The core-mantle structures they show in feldspar grains could have also formed by strain-induced boundary migration recrystallisation, whereby the new grains form by bulging. Bulging may start to operate at lower temperatures in feldspar than rotation crystallisation (Tullis and Yund, 1991; Fitz Gerald and Stünitz, 1993).

Evidence of intracrystalline deformation, ranging from bent or kinked crystals, deformation bands, undulose extinction, tapering deformation twins (Fig. 9d), and bulging of twin and kink boundaries, occurs in both albite and K-feldspar porphyroclasts, although they are much more prevalent in albite (Figs. 9d and 11a). However, intracrystalline deformation is generally accompanied by microfracturing and grain segmentation, particularly in K-feldspar. Some feldspar grains are transected by numerous microfractures, which may divide the crystal into many new grains that are slightly misoriented with respect to each other; these new grains are connected by narrow bands of recrystallised albite along the fracture planes (Fig. 11a). In our samples, core-mantle structures (White, 1976), with the mantle becoming progressively finer grained and grading into a fine-grained groundmass (mortar structure), is only developed in association with albite porphyroclasts (Fig. 11b), and not K-feldspar. The presence of strong crystallographic preferred orientations (CPO) in bands dominated by fine-grained albite, particularly where the mantles of the albite porphyroclasts grade into the fine-grained groundmass is consistent with dynamic recrystallisation of albite as a strain-accommodating process. Fitz Gerald and Stünitz (1993) studied mylonitic granitoid rocks in the Alps that formed at nearly identical metamorphic conditions to those of the Brunswick complex. Based on detailed optical and transmission electron microscope studies, Fitz Gerald and Stünitz favoured grain-size reduction by nucleation recrystallisation of albite at the expense of plagioclase and K-feldspar, followed by fluid-assisted grain boundary sliding, rather than grain boundary migration or subgrain rotation mechanisms (see also Stünitz and Fitz Gerald (1993)). Despite the similarities between the microstructures in their samples and the Brunswick complex, suggesting that nucleation recrystallisation may have been important in our rocks as well, they did not observe any CPO of albite. The activity of nucleation recrystallisation versus grain boundary bulging remains ambiguous in our rocks, but the potential importance of the former process cannot be ruled out because of the evidence for syn-tectonic albitisation of feldspar (as described below). Regardless of which of these two mechanisms was dominant, we concur with Fitz Gerald and Stünitz (1993), that at low temperature conditions (330–400°C) dynamic recrystallisation of albite can be enhanced by the presence of an aqueous fluid. Thus, syn-tectonic, fluid-assisted growth of albite at the expense of K-feldspar may have weakened the felsic volcanic rocks.

8. Feldspar alteration

8.1. Albitisation: constraints from deformation twinning

Metamorphic albite (either in the form of porphyroblasts or neoblasts formed during recrystallisation and crack-seal processes) is generally either untwinned or has one or two twins that traverse the entire grain. These twins are probably growth or annealing twins (Vernon, 1965). Needle-shaped deformation twins extending only across parts of the albite crystal occur in veins and fine-grained mylonites, but these

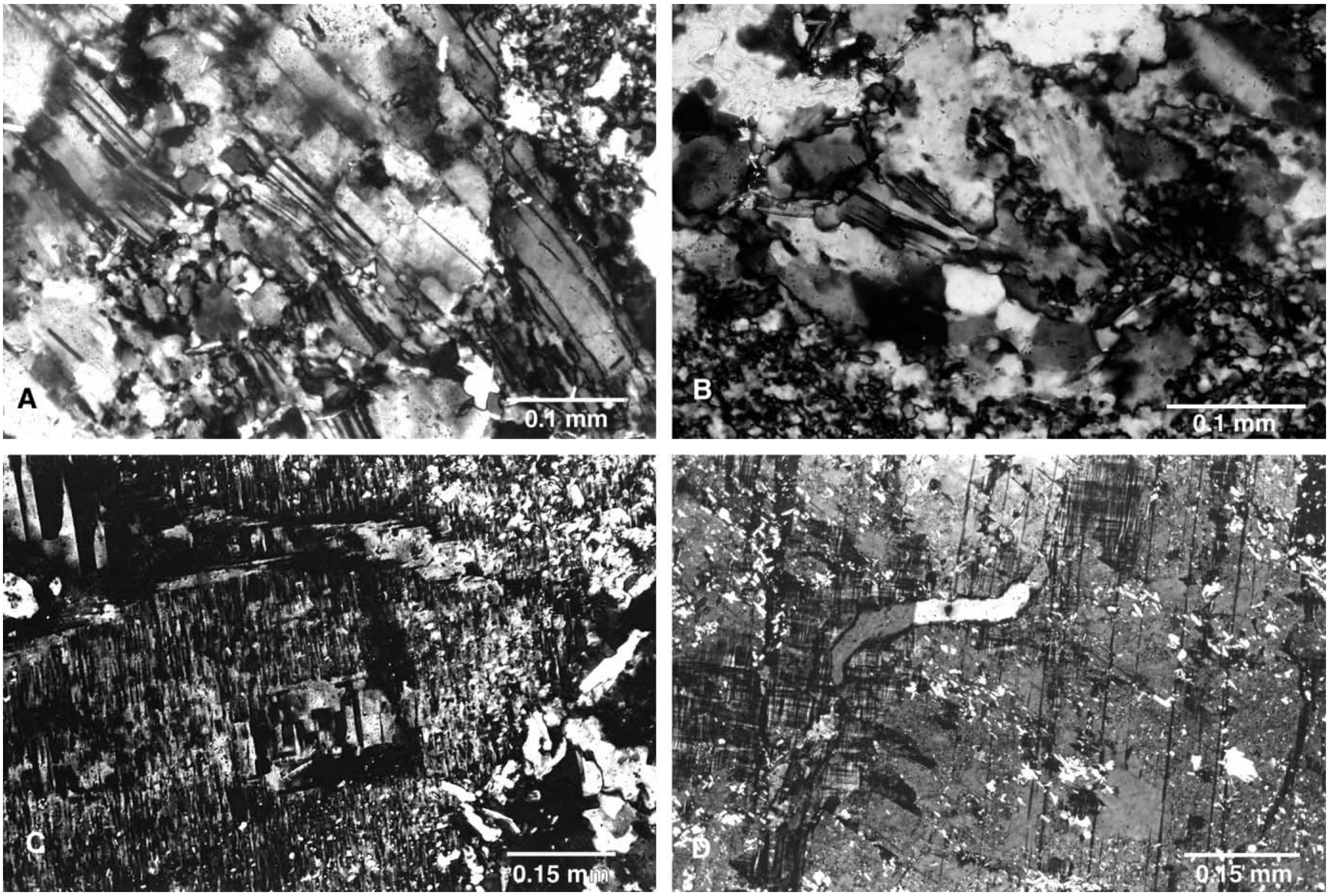

Fig. 11. (a) Deformed albite porphyroclast showing recrystallisation, undulose extinction, deformation twins and formation of some subgrains. Deformation included kinking and probably some microfracturing. The latter may have become the site of nucleation recrystallisation. Note irregular boundaries of the new grains and albite twins indicating grain/twin boundary migration. (b) Detail of a core-mantle structure of an albite porphyroclast near the intersection between the mantle and the mylonitic groundmass. Scanning electron microscope investigations confirmed that virtually all recrystallised grains are albite. (c) Detail of rim of chessboard albite around perthitic K-feldspar, which pseudomorphed an alkali feldspar phenocryst (possibly originally sodic sanidine). K-feldspar is partially transformed into tartan twinned microcline. The chessboard rim displays some diffuse M-twinning inherited from the K-feldspar host as is indicated by continuity of twin orientations from the microcline into the albite rim. (d) Antiperthitic albite. Note extension of the microcline from microfractures into the albite suggesting replacement.

deformation twins are relatively rare outside the shear zones (Figs. 9d and 11a). This is in accordance with the predicted difficulties in producing mechanical twins in low albite, since it requires a high shear stress to destroy the strong Al–Si bonds during Al/Si reordering (Starkey, 1963; Smith, 1974). Nevertheless, deformation twinning has been reported in low albite elsewhere (e.g. Mehta, 1979; Fitz Gerald et al., 1983). The scarcity of twins in metamorphic matrix albite outside the mylonite zones is in stark contrast to the polysynthetic albite twinning of the albite porphyroclasts (Fig. 9d). The polysynthetic twins are typically deformation twins, commonly not extending across the crystal and are wedge or needle shaped (e.g. Vance, 1961; Vernon, 1965). The twin boundaries are generally irregular in appearance, with remnants of twins floating as isolated patches within a larger crystal. These microstructures indicate twin boundary migration (Vernon, 1981). The simplest explanation for the formation of these deformation twins is that these phenocrysts were still at least partially disordered volcanic oligoclase (Noble, 1966; Smith, 1974, p. 364) when subjected to the early increments of D_1 deformation and M_1 metamorphism. During subsequent pseudomorphic replacement by albite, the phenocrysts partially to completely retained the mechanical twins (cf. Brown, 1989). In summary, these microstructures, and the preferential growth of M_1 pumpellyite and/or epidote-group minerals in the albite pseudomorphs after plagioclase (using Ca released during albitisation, Fig. 6b), suggest that albitisation of plagioclase took place during the early stages of regional deformation and metamorphism rather than during a pre-tectonic, hydrothermal alteration event.

8.2. Formation of chessboard albite

The alkali-feldspar porphyroclasts generally are patch-,

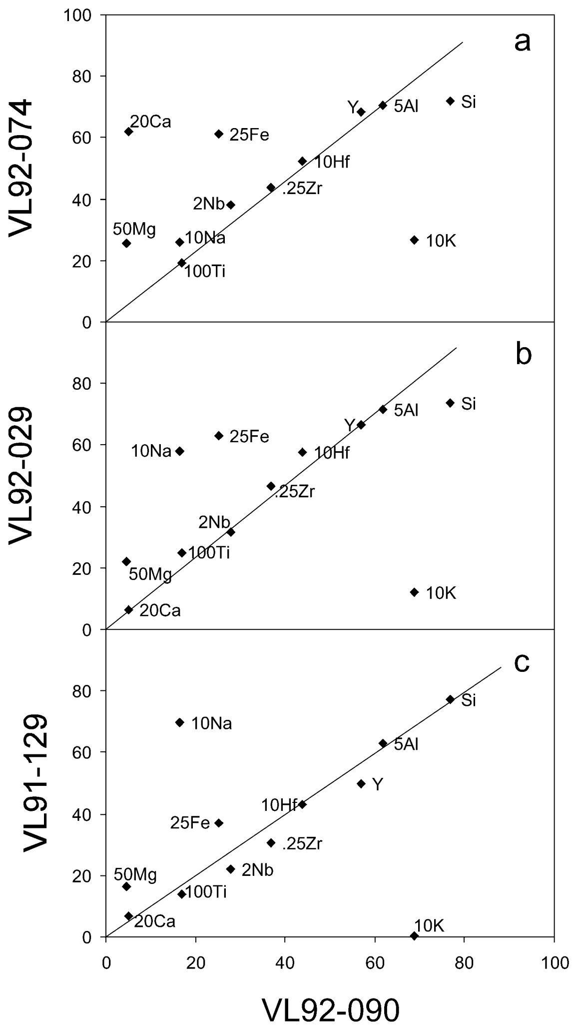

Fig. 12. Element concentrations of a relatively weakly strained FP3 volcanic rock plotted against progressively higher strained tectonites, culminating in ultramylonite (Vl91-129) from the roof-thrust shear zone according to the method of Grant (1986). Elements are arbitrarily scaled for presentation purposes, with the major elements recorded in wt% and the trace elements as ppm. Al, Ti, Nb and Zr lie on a straight line through the origin (isocon), demonstrating that these elements were immobile irrespective of the degree of alteration. Elements plotting above the isocon represent relative gains, while those below the isocon represent relative losses. Full chemical analyses are recorded in Appendix B.

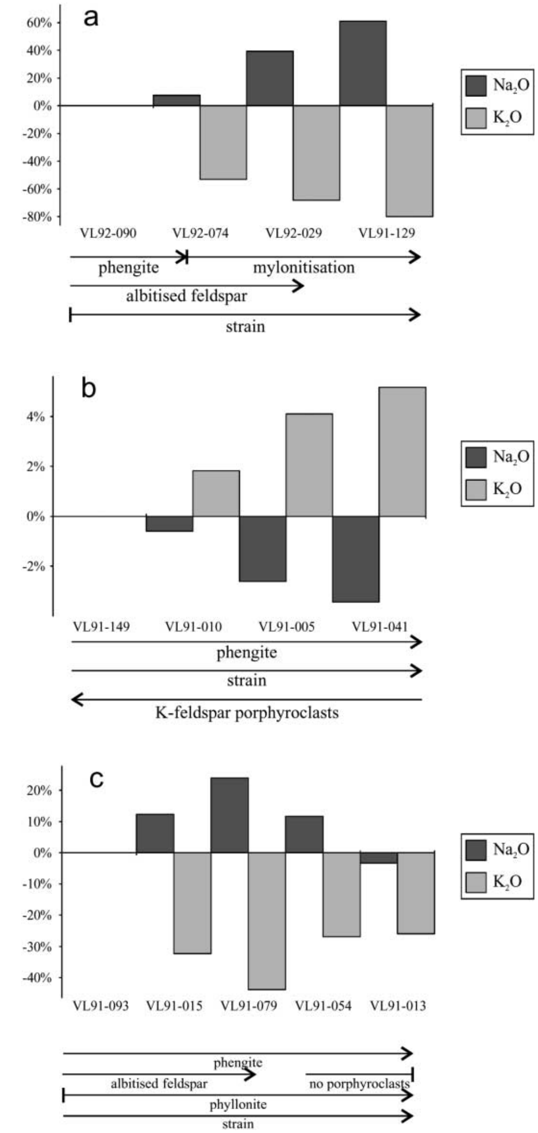

Fig. 13. Mass-balance profiles showing percentage change of total alkali concentration calculated on the basis of Al immobility, as determined by the isocon method of Grant (1986), along strain gradients in felsic volcanic rocks of the Spruce Lake nappe. For location of samples see Figs. 1 and 3 and Appendix A. Full chemical analyses are recorded in Appendix B. The relative strain was estimated from the degree of microboudinage and/or intensity of fabric development. (a) Strain profile through an early D_1 mylonite developed in FP3 felsic volcanic rocks that mark the contact between the Spruce Lake and Canoe Landing Lake nappes. Phengite, albitisation of K-feldspar and growth of pumpellyite/epidote took place during the earliest increments before mylonitisation started to breakdown the albitised porphyroclasts. Samples are compared with VL92-90. (b) Strain profile of FP2 phyllonites from the basal shear zone of the Spruce Lake nappe. This part of the shear zone is a late D_1 structure and experienced little or no previous albitisation. Samples are compared with VL91-149. As the strain represented in the rocks increases from VL91-149 to VL91-041 the amount of phengitic mica increases whilst feldspar porphyroclasts are destroyed. (c) Strain profile through FP1 felsic volcanic rocks from the basal thrust shear zone of the Spruce Lake nappe that initially experienced alteration into albitised tectonites (samples VL91-046 and VL91-054) prior to phyllonite development (samples VL91-058 and VL91-013). Samples are compared with VL91-015.

vein- or micro-perthites (Starkey, 1959). In the least deformed and least altered rocks, the albite patches seem to develop from vein- or micro-perthite by irregular broadening of the exsolved albite lamellae (Fig. 6a); thus, the vein- or micro-perthite probably acted as nuclei for the albite patches. All albite patches within one crystal have the same crystallographic orientation and commonly show a blocky, chessboard-like twin pattern. The twinning consists of square or rectangular shaped albite twin lamellae with low aspect ratios that wedge out either in pointed terminations, join with neighbouring lamellae into complex blocky geometries or are abruptly truncated on (001). The relationship between patch-perthite and chessboard albite has been observed in many other areas and is generally ascribed to partial replacement of alkali-feldspar by albite (e.g. Smith, 1974). Porphyroclasts within a thin section vary frequently between pure vein-perthite (5–25% albite) to a nearly complete replacement (80–90%) of the alkali-feldspar by chessboard albite. The patches of chessboard albite are locally so coarse that the remainder of the alkali-feldspar resembles highly irregular shaped veins. Some patch-perthites also have rims of chessboard albite (Fig. 10c), which is consistent with a replacement origin. Both the rims and patches of chessboard albite have inherited the crystallographic orientation of the alkali-feldspar because (010) and (001) are continuous from the K-feldspar into the adjacent albite. Moreover, where the alkali-feldspar displays tartan cross-hatching, chessboard albite inherited the M-twin pattern of the host, including some rare, diffuse pericline twins of the host, although the twins are wider in the patches compared with the rim. The diffuse pericline twins appear unstable and contain smaller scale albite twins. This type of chessboard albite structure is remarkably similar to the irregular geometries shown by the arrays of wedge shaped, stubby albite twins with tapering terminations that replace coarser, unstable pericline twin lamellae in microcline (McLaren, 1984, fig. 3; Fitz Gerald and McLaren, 1982). The M-twinning formed during transformation from a monoclinic to a triclinic symmetry.

The following five observations suggest that formation of chessboard albite took place during deformation. First, chessboard albite tends to form near, or in, zones of high strain and becomes more penetrative with increasing strain (a relationship observed in the same rocks by Starkey in 1959), although once formed there is a tendency to destroy the chessboard pattern by recrystallisation, probably involving twin boundary migration. Second, selective development of chessboard albite in only some of the pulled-apart segments of boudinaged alkali feldspar porphyroclasts indicate that the chessboard must have formed after tectonic segmentation. Third, the compositions of the albite patches and matrix metamorphic albite (Table 1) are identical and the temperature estimates determined by feldspar geothermometry are consistent with metamorphic temperatures calculated by other means (Currie and van Staal, 1999). Fourth, the predominance of tapering twins indicates that most are mechanical twins (Starkey, 1959). Fifth, chessboard albite structures are virtually absent in regular albite, both in matrix grains and porphyroblasts, but common in fibrous-vein albite, fractures within K-feldspar (Fig. 6a), and where larger extensional veins cut through an alkali-feldspar porphyroclast (Fig. 8b). These observations suggest that chessboard-albite formation is related to the interaction of alkali-feldspar with metamorphic fluids during deformation. Initially, this interaction first promotes formation of crosshatched low microcline from untwinned, intermediate microcline or orthoclase and second replacement or epitaxial new growth by chessboard-albite. Fibrous epitaxial growth is restricted to vein segments, which cut through alkali-feldspar porphyroclasts. Replacement and growth was followed by recrystallisation, partially removing the unstable, irregular-shaped, stubby albite twin arrays described by Fitz Gerald and McLaren (1982), so that similarly oriented twins coalesce into broader patches. This broadening and simplifying of the albite twins may well be driven by the tendency to remove the high-energy incoherent interfaces and terminations of tapering albite twins during recrystallisation (Vernon, 1965).

8.3. K-feldspathisation of albite

Apart from the observed albitisation of plagioclase and alkali-feldspar porphyroclasts, the examined rocks also contain rare evidence of replacement of albitised plagioclase by K-feldspar. Patches of microcline extending from veins and microfractures into albite porphyroclasts suggest a replacement origin rather than exsolution (Fig. 10d). Due to the difficulty of selectively altering plagioclase phenocrysts to albite without affecting the microcline patches and veins, albitisation must have taken place before replacement by microcline. D_1 microcline-bearing veins and overgrowths of microcline on D_1 boudinaged, untwinned alkali-feldspar porphyroclasts (Fig. 7c) indicates that K-feldspathisation was mainly a D_1 event, but postdates albitisation.

9. Alkali-metasomatism

Mass balance calculations were carried out to better understand the interaction between the fluids and the volcanic rocks hosting the D_1 shear zones, particularly with respect to why some rocks were transformed into albite-rich mylonites, whereas others became phengite-rich phyllonites. We compared the least deformed sample of each of the three felsic volcanic suites (FP1, FP2 and FP3) with highly-strained examples of the same suite from areas for which we have the best available field and petrographic constraints on spatial distribution of the rocks with respect to shear zones, relative finite strain and associated mineralogical changes.

Quartz and albite-rich mylonites (Fig. 9c) are particularly well preserved in the shear zone that marks the roof-thrust of the Spruce Lake nappe, which constitutes the boundary with the overlying Canoe Landing Lake and/or blueschist nappe (Fig. 1). The roof-thrust shear zone is mainly developed in the volcanic rocks of suite FP3 (Fig. 5). Samples displaying variable states of strain were collected, owing to a lack of continuous outcrop, along a composite profile across the shear zone. Phyllonites are characteristic of the basal shear zone of the Spruce Lake nappe (Fig. 7d). Nearly continuous access to the basal shear zone is provided by drill-holes in the immediate volcanic hanging wall of the Caribou massive sulphide deposit (Fig. 3). We used the chemical data of the volcanic rocks of suites FP1 and FP2 to monitor changes induced by increasing strain across this shear zone. The least strained tectonites developed from FP2 suite rocks show little to no albitisation of K-feldspar, whereas albitisation is characteristic of the shear zone-related tectonites developed in FP1 volcanic rocks.

Element concentration plots of variably strained rocks compared with a reference rock according to the method of Grant (1986) show that Al, Zr, Ti and Nb lie on a straight line through the origin (isocon), independent of the degree of strain and alteration (Fig. 12). This indicates that these elements were relatively immobile, which is consistent with the results of many other alteration studies (e.g. Grant, 1986; Selverstone et al., 1991; Streit and Cox, 1998; Lentz, 1999). Hence, these elements were used to monitor the alteration of the mylonites and phyllonites.

With increasing strain, the tectonites of the roof thrust shear zone show a progressive increase in Na, Fe and Mg, and loss of K and Si culminating in quartz- and albite-rich ultra-mylonites. Most relevant to this paper is the large gain in Na and loss of K (Na-metasomatism) reflecting a progressive increase in albitisation of alkali-feldspar (Figs. 12 and 13a).

The chemical changes displayed by the phyllonites of the basal shear zone of the Spruce Lake nappe (Fig. 7d) are complex and variable because, in part, they overprint previously Na-metasomatised tectonites (Fig. 13b and c). The degree of metasomatism associated with development of the phyllonites is generally much less than that in the mylonites. The phyllonites developed from FP2 volcanic rocks appear to display consistent slight gains in K and Mg, and minor losses of Na with increasing strain (Fig. 13b), however, the absolute changes in element concentrations are generally within analytical error. Si and Ca show no significant modifications. The same trend of K gain and Na loss is shown by the phyllonites developed from the FP1 volcanic rocks but the amount of K-metasomatism is more obvious. This type of metasomatism mainly relates to an increase in phengite content and destruction of feldspar (albite and K-feldspar) in the phyllonites.

We interpret our data set as follows. Na-rich aqueous fluids were preferably channelled along the roof-thrust related shear zones of the Spruce Lake nappe during its underthrusting beneath the Canoe Landing Lake and/or blueschist nappes (the burial stage of the M_1 P–T–t path). This is consistent with the marked oxygen isotope shift from low strain metabasalts to blueschists (Fig. 10) and the relative increase in Na, Ca and Mg with respect to Si (Dipple and Ferry, 1992). The timing of the Na-metasomatism relative to D_1/M_1 is constrained by: (i) the presence of pumpellyite in the albite-sealed intragranular fractures of partially albitised K-feldspar, and highly deformed albite-bearing veins. Generally, pumpellyite is partially replaced by epidote or stilpnomelane (Fig. 8b), suggesting that the pumpellyite-bearing veins formed early, during the prograde P–T path of M_1; (ii) the phyllonites do not show a late stage of albite porphyroblastesis; (iii) albite–quartz mylonites or tectonites are partially or completely transformed into phyllonites; and (iv) evidence of Na-metasomatism is best preserved in less-altered pods within phyllonites. Na-metasomatism promoted the transformation of K-feldspar into patch-perthite and eventually into chess-board albite. The apparent gains in Ca, Mg and Fe are consistent with the growth of pumpellyite or epidote, and phengite. Phengite probably also incorporated some of the K released during albitisation of K-feldspar. The remaining K was progressively flushed out of the rock by upward flowing fluids. The albitisation of K-feldspar follows the generalised reaction displayed below. The composition of the phengite is from Currie and van Staal (1999):

$$6KAlSi_3O_8 + \text{fluid}\ (2H_2O + 2Na^+ + (Fe, Mg)^{2+})$$
$$= 2NaAlSi_3O_8 + K_2(Fe, Mg)Si_7Al_4O_{20}(OH)_4$$
$$+ 5SiO_2 + \text{fluid}\ (4K^+)$$

Formation of the phengite-rich phyllonites, K-feldspar-rich veins and the rare partial replacement of albite porphyroclast by K-feldspar (Fig. 10d) are the textural manifestations of K-metasomatism. Petrography suggests the phyllonites formed in two ways. First, the phyllonites developed in felsic volcanic rocks that show little to no previous albitisation of alkali-feldspar (Fig. 13b); hence, they probably represent a new set of shear zone-related tectonites, developed after peak high-pressure metamorphism in the basal shear zone during out-of-sequence thrusting (van Staal, 1994; Currie and van Staal, 1999). The second, and probably more common, type are phyllonites formed in already existing Na-metasomatised tectonites (Fig. 13c). During shearing, the remaining albitised feldspars were progressively destroyed by dissolution and alteration to phengite. The K-metasomatism thus attempts to counter-balance the earlier Na-gains and K-losses, according to the following generalised reaction:

$$4NaAlSi_3O_8 + \text{fluid}\ (2H_2O + 2K^+ + (Fe, Mg)^{2+})$$
$$= K_2(Fe, Mg)Si_7Al_4O_{20}(OH)_4 + 5SiO_2 + \text{fluid}\ (4Na^+)$$

10. Summary and conclusions

In this study, we describe how a progressive, thrusting-related deformation (D_1) resulted in several generations of shear zones and associated folds. The shear zones in the felsic volcanic rocks of the Spruce Lake nappe formed during underthrusting and incorporation of the rocks into the Early Palaeozoic Brunswick subduction complex. These shear zones record an unusually complex history of alkali-metasomatism caused by large-scale infiltration of aqueous fluids. The fluids were probably generated mainly by large-scale dewatering (±dehydration) of

Table 2

Sample	VL91-005 FP2	VL91-010 FP2	VL91-013 FP1	VL91-015 FP1	VL91-041 FP2	VL91-046 FP1	VL91-054 FP1	VL91-058 FP1	VL91-129 FP3	VL91-149 FP2	VL92-029 FP3	VL92-074 FP3	VL95-030 FP3
Wt%													
SiO_2	74.61	74.77	72.84	71.81	73.47	71.22	72.57	71.82	77.50	74.99	73.96	71.73	70.70
TiO_2	0.17	0.17	0.24	0.26	0.18	0.26	0.24	0.24	0.14	0.16	0.25	0.19	0.28
Al_2O_3	13.17	13.06	14.02	14.49	13.34	14.73	13.95	13.58	12.60	12.84	14.35	14.00	14.30
Fe_2O_3t	1.30	1.52	2.24	2.70	1.58	2.34	2.54	2.37	1.48	1.82	2.53	2.44	2.10
MnO	0.02	0.02	0.04	0.04	0.03	0.04	0.05	0.06	0.01	0.02	0.05	0.03	0.03
MgO	0.45	0.32	1.64	1.08	0.34	0.97	0.93	1.02	0.33	0.23	0.44	0.51	0.30
CaO	0.24	0.27	0.45	0.63	0.32	0.77	0.52	1.07	0.34	0.23	0.32	3.09	0.51
Na_2O	1.17	1.34	0.61	2.07	1.11	2.38	1.94	1.14	7.00	1.37	5.83	2.60	1.10
K_2O	7.97	7.70	5.67	5.28	8.17	5.41	5.56	5.47	0.05	7.41	1.22	2.67	8.80
P_2O_5	0.17	0.18	0.19	0.19	0.18	0.18	0.18	0.16	0.22	0.19	0.22	0.24	0.31
Total	100.12	100.24	100.22	100.40	99.84	99.93	100.14	99.48	100.15	100.40	100.47	99.56	99.90
ppm													
Ba	365	372	385	452	287	319	400	334	37	430	417	394	1500
Be	2	2	3	3.2	2.2	3.1	3.1	3.1	0.8	1.5	1.4	4.2	3.5
Co	20	36	10	18	20	23	19	27	58	12	27	35	2.5
Cs	2.5	4.3	7.6	6.8	6.2	5	3.6	4.8	0.17	2.5	1.4	7.9	8.5
Ga	17	16	20	18	18	20	19	19	12	19	20	23	21
Hf	3.3	3.4	4.6	4.5	3.4	4.4	4.3	4.7	4.3	4.4	5.8	5.2	7.3
Mo	0.4	0.6	1.1	0.9	0.5	1.1	1.4	1.4	1.4	2.2	0.9	0.7	0.3
Nb	12	11	13	13	12	14	12	13	11	12	16	19	17
Pb	18	23	32	40	15	36	21	24	7	16	37	26	25
Rb	240	240	250	200	270	200	210	220	3.6	220	76	140	300
Sc	3.9	4.0	5.6	6.0	4.1	6.1	5.6	5.4	5.9	5.0	13.0	15.0	11.0
Sr	33	34	10	41	39	83	48	41	28	84	110	85	60
Th	14	15	22	19	15	19	16	16	13	16	17	18	32
U	6.8	6.5	7.2	7.5	7.6	7.8	6.7	6.9	8.2	7.3	6.2	5.7	7.7
V	2.5	2.5	5	8	2.5	8	7	7	2.5	2.5	8	2.5	2.5
Y	51	51	58	55	55	57	55	52	50	58	67	68	88
Zn	23	38	57	47	28	68	46	39	34	19	65	61	40
Zr	115	119	163	150	118	155	145	145	122	128	187	174	290

the deeper parts of the subduction complex, and then channelled upwards along the shear zones. These fluids were dominantly of seawater origin, and resulted in marked oxygen isotope alteration where fluid/rock ratios were highest.

The earliest D_1 shear zones formed during underthrusting and accretion of the underplated volcanic and associated sedimentary rocks and were mainly characterised by albite-rich mylonites; later-D_1 shear zones are represented by phengite-rich phyllonites that probably accommodated uplift after incorporation into the subduction complex. The mylonites formed by recrystallisation of albitised feldspar porphyroclasts. Albitisation and the unusual low temperature of albite recrystallisation ($<400°C$) were promoted by infiltration of Na-rich aqueous fluids. Unaltered K-feldspar generally behaved brittlely, suggesting that in the presence of abundant intergranular fluids, albite is significantly weaker than K-feldspar. Hence, metasomatic replacement of K-feldspar by albite weakened the rocks. The permeability required for the fluid infiltration was probably enhanced by continuous microfracturing, evident from the abundance of crack-seal vein complexes in the shear zones. The phengite-rich phyllonites accommodated late-D_1 out-of-sequence thrusting associated with emplacement of blueschists and associated high pressure rocks on top of lower pressure greenschist facies rocks (van Staal et al., 1990).

Mass-balance calculations suggest that the phyllonites were, at least in part, the product of a weak K-metasomatism that partly overprinted mylonites formed during an earlier, more profound phase of Na-metasomatism. The change from Na- to K-metasomatism does not necessarily reflect a major change in fluid composition with time. The aqueous fluid only has to provide a medium for cation exchange. A decrease in temperature is sufficient to increase the stability field of K-feldspar with respect to albite and muscovite (phengite) relative to both feldspars in the presence of a fluid (Simpson and Wintsch, 1989; Wintsch et al., 1995).

Acknowledgements

The structural work that is reported in this paper began while van Staal was a PhD student (1980–84) of Paul Williams. Van Staal wishes to thank Paul for sound scientific advice and numerous discussions over the years. Critical reviews by Greg Dipple and Joe White and internal

GSC reviews by Ken Currie, Jurgen Krauss, Steve Lucas and Jim Ryan improved the manuscript. John Stirling assisted with the electron microprobe analyses. The members of the Department of Earth Sciences of Oxford University, England, particularly John Dewey and Des McConnell, are thanked for discussions and hospitality while the first author was a visiting scientist. Chemical analyses were conducted at the Department of Earth Sciences, Keele University, England. Geological Survey of Canada contribution 1996214.

Appendix A

Drill-hole number and depth for the samples from the Caribou massive sulphide deposit

Sample number	Drill-hole	Depth (m)
VL91-005	109	8
VL91-010	109	301
VL91-013	109	200
VL91-015	109	500
VL91-041	108	135
VL91-046	108	267
VL91-054	38	34
VL91-058	58	138

Appendix B

Chemical analyses for the samples discussed in the text and Figs. 12 and 13. Details of the sampling and analytical methodology, as well as the elemental detection limits and accuracy are given in Rogers (1994) and Table 2.

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