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HIGH-GRADE METAMORPHISM IN THE WESTERN CAPE BRETON HIGHLANDS, NOVA SCOTIA, AND ITS RELATION TO TECTONISM*

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Abstract

Geothermobarometry and detailed mapping in the western Cape Breton Highlands of Nova Scotia indicate that upper-amphibolite-facies (~700°C, >8 kilobars) supracrustal and plutonic rocks structurally overlie Ordovician–Silurian supracrustal sequences that exhibit inverted metamorphic isograds. Maximum T–P values (~640°C, 6.4 kilobars) in the latter occur at the boundary between the two suites. Geothermobarometry based upon garnet – biotite – amphibole – epidote (± clinopyroxene) assemblages in the overlying suite and garnet – biotite – muscovite (rutile ± ilmenite ± staurolite) assemblages in the Ordovician–Silurian suite shows an abrupt increase in calculated P and T across the shear zone separating the two suites. The overlying rocks thus originated at considerably deeper levels in the crust than the underlying supracrustal suite. It is thus unlikely that the two suites are correlative, as previously supposed. Extensive early to middle Devonian plutonism and uplift accompanied and followed the peak of metamorphism. Devonian greenschist-facies (~400°C, 2.3 kilobars) shear zones give pressures similar to those of andalusite–cordierite hornfels aureoles around the late plutons. The observed metamorphism can be modeled by late Silurian thrust emplacement of a thick (20 km), hot allochthon. In this model, ~14 kilometers of denudation preceded development of the low-angle, extensional Margaree Shear Zone, along which a further 6 kilometers of cover was removed prior to Carboniferous sedimentation.

Keywords: metamorphism, tectonics, geothermobarometry, Cape Breton Highlands, Nova Scotia.

Sommaire

Une étude de la géothermobarométrie et une cartographie détaillée dans le haut plateau occidental de la Nouvelle-Écosse montrent que des roches supracrustales et plutoniques, recristallisées dans le faciès amphibolite supérieur, recouvrent une séquence de roches supracrustales ordovicienne et silurienne qui témoigne d'isogrades métamorphiques inversés. Les valeurs maximales de T et de P (~640°C, 6.4 kilobars) dans cette deuxième séquence se retrouvent à l'interface entre les deux suites. Une étude géothermobarométrique fondée sur les assemblages à grenat – biotite – amphibole – épidote (± clinopyroxène) dans la séquence supérieure, et à grenat – biotite – muscovite (± rutile ± ilménite ± staurolite) dans la séquence ordovicienne – silurienne, indique une augmentation abrupte dans les valeurs de P et de T calculées de part de d'autre de la zone de cisaillement qui sépare les deux suites. Les roches de la séquence supérieure se seraient donc formées à une profondeur considérablement plus grande dans la croûte que les roches supracrustales de la séquence inférieure. C'est donc dire que les deux séquences ne sauraient être équivalentes, comme celà avait été proposé. Un épisode répandu de plutonisme au dévonien précoce à moyen et un soulèvement ont accompagné et suivi le paroxysme métamorphique. Des zones de cisaillement dévoniennes, actives au faciès schistes verts (~400°C, 2.3 kilobars), indiquent une pression de formation semblable à celle indiquée dans les cornéennes à andalusite + cordiérite qui entourent les plutons tardifs. Notre modèle tectonique des assemblages métamorphiques fait appel à la mise en place, par chevauchement au Silurien tardif, d'un allochtone chaud, d'une épaisseur de 20 km. Quatorze de ces vingt km ont été érodés avant le développement de la zone de cisaillement extensionnelle de Margaree, à faible pendage. L'activité le long de cette zone est responsable de l'érosion d'une couche supplémentaire de 6 km avant le début de sédimentation au Carbonifère.

(Traduit par la Rédaction)

Mots-clés: métamorphisme, reconstruction tectonique, géothermobarométrie, Cap Breton, Nouvelle-Écosse.

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Introduction

The Cape Breton Highlands comprise a block of plutonic and metamorphic rocks surrounded by unmetamorphosed sedimentary and minor volcanic rocks of late Devonian to Carboniferous age (Fig. 1). The highlands form part of the plutonic and metamorphic core, or internal zone, of the Appalachian Orogen, but correlation with other parts of this core has proved to be difficult and controversial. On the basis of their plutonic and metamorphic character, early investigators assumed the highlands to be composed mainly of Precambrian rocks, with some relatively minor plutons of Devonian age (Fletcher 1885, Cameron 1948, MacLaren 1955, Neale & Kennedy 1975), all unconformably overlain by surrounding younger strata. The discovery that the highlands include late Ordovician to Silurian rocks (Currie *et al.* 1982, Jamieson *et al.* 1986) that are faulted against surrounding low-grade rocks (Currie 1977) profoundly altered this view. Radiometric data indicate metamorphism in late Silurian to early Devonian time (~415–385 Ma: Barr & Jamieson 1991, Jamieson *et al.* 1986), with associated and subsequent emplacement of granitic plutons (405–365 Ma: Jamieson *et al.* 1986, Dunning *et al.* 1990). Initial interpretations of metamorphism assumed either a single metamorphic sequence relatively slightly disrupted by faulting (Craw 1984, Plint & Jamieson 1989), or two distinct metamorphic sequences separated by a fault or unconformity (Currie 1987).

Recent remapping (Lynch *et al.* 1993, 1995, Lynch & Tremblay 1992, 1994) has clarified the structural history of the western highlands, whereas advances in geothermobarometry (Berman 1991, Mäder & Berman 1992, Mäder *et al.* 1994) permit more accurate estimates of metamorphic conditions. These developments allow a fruitful re-examination of the metamorphic history of the western highlands and its relation to the structural history.

Geological Setting

The pre-Carboniferous geology of Cape Breton Island can be divided into southeastern and northwestern portions (inset, Fig. 1). The southeastern portion (unit 2, Fig. 1), comprising carbonate-bearing Proterozoic sedimentary strata, late Proterozoic volcano-sedimentary sequences and correlative plutons, and Cambro-Ordovician strata containing Acado-Baltic faunas, forms a terrane typical of the Avalon Zone, which fringed Gondwana during early Paleozoic time (Strachan & Taylor 1990). The rocks to the northwest, including the Cape Breton Highlands, can be divided into five lithological associations: (1) Proterozoic rocks (unit 1, Fig. 1) including anorthosite, syenite, and tonalitic orthogneiss, giving igneous and metamorphic ages of ~980 to >1200 Ma (Miller *et al.* 1996), (2) late Proterozoic (~560 Ma, Dunning *et al.* 1990) calc-alkaline plutons, with narrow intervening screens of deformed Proterozoic, locally marble-bearing metasedimentary strata, both intruded by rare early Ordovician plutons (unit 3, Fig. 1), (3) variably deformed and metamorphosed late Ordovician to Silurian volcanic rocks and turbidites (Ordovician–Silurian suite; unit 4, Fig. 1), (4) deformed plutons of Silurian and older age, with enclaves of metasedimentary rocks (unit 5, Fig. 1), and (5) little deformed plutons (unit 6, Fig. 1) and volcanic rocks (Fisset Brook Formation and correlative rocks; unit 7, Fig. 1) of middle to late Devonian age. Barr & Raeside (1986, 1989) divided these rocks into four fault-bounded "terranes" which, they suggested, may have been widely separated until late Silurian time, when plutonism and metamorphism affected the whole of the highlands. The concept of widely dispersed "terranes" has provoked ongoing controversy. Stratigraphic overlap between "terranes" has been suggested for the latest Proterozoic units (Keppie 1990) and Ordovician–Silurian units (Lin *et al.* 1991, Lynch & Tremblay 1992, Lin 1993, Lynch *et al.* 1993, Chen *et al.* 1995).

Prior to recent remapping, many investigators assumed that steep strike–slip faults separated contrasting lithological associations, although the presence of major reverse faults has been known for many years (Milligan 1970, Currie 1977). Lynch (1996) introduced a new idea by proposing that the topographically and structurally highest parts of the Cape Breton Highlands (Fig. 1; units 1, 5, and part of 3) comprise a complexly folded and dissected nappe, roughly 100 by 50 km in size, as indicated on Figure 1. The rocks assigned to the supposed nappe consist mainly of deformed Silurian and older plutons, with lesser amounts of amphibolite-facies deformed supracrustal rocks and Proterozoic meta-igneous rocks. According to the nappe model, the Ordovician–Silurian suite (Fig. 1, unit 4) lies beneath the nappe, exposed around its edges and as tectonic windows within it. In the western Cape Breton Highlands (Fig. 2), the Ordovician–Silurian suite exhibits imbrication and inverted metamorphic isograds (Craw 1984, Currie 1987, Plint & Jamieson 1989) beneath a polydeformed thrust-surface (Lynch 1996). Structurally above this surface, the topographically highest parts of the highlands consist of deformed granitic plutons and lesser amounts of migmatitic metasedimentary rocks of uncertain provenance. We here consider the rocks above the thrust the "allochthonous high-grade suite". Volcanic rocks of the Ordovician–Silurian suite gave early to middle Silurian U–Pb zircon ages of emplacement, approximately coeval with emplacement of granitic plutons into the allochthonous high-grade suite (~435 Ma: Jamieson *et al.* 1986, Dunning *et al.* 1990, Chen *et al.* 1995).

Metamorphism and plutonism affected the whole of the highlands in latest Silurian and early Devonian

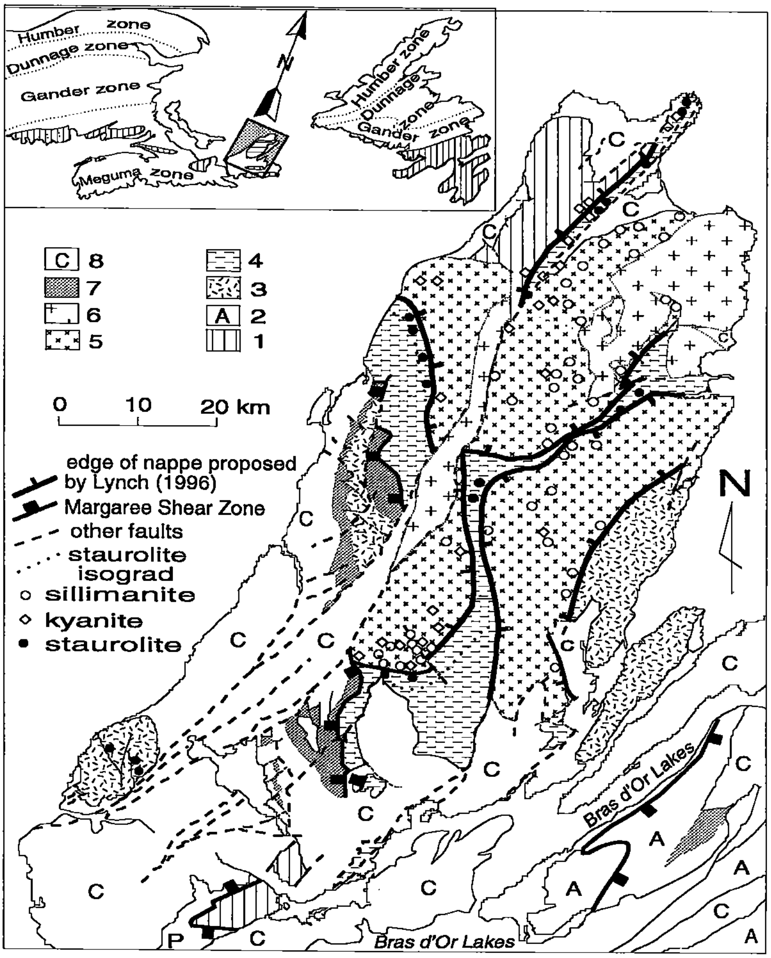


FIG. 1. Generalized geological map of northern Cape Breton Island (after Lynch *et al.* 1995). Units: (1) Proterozoic (~980–1220 Ma) metasedimentary and igneous rocks. The region southwest of Bras d'Or Lakes is of uncertain age. (2) Low-grade late Precambrian volcanic, sedimentary and plutonic rocks (Avalon Zone). (3) Late Proterozoic (~560 Ma) calc-alkaline plutons, with narrow screens of deformed Proterozoic metasedimentary strata and local early Ordovician plutons. (4) Late Ordovician to Silurian volcanic rocks and turbidites (Jumping Brook suite). (5) Deformed Silurian and older plutons, with rare metasedimentary relics. (6) Little-deformed granitic plutons of Devonian age (only the major plutons are shown). (7) Late Devonian bimodal volcanic rocks (Fisset Brook Formation). (8) Carboniferous sedimentary cover. The inset map shows the position of the main map (shaded rectangle) within the Appalachian orogen in Canada. The Avalon Zone is hachured with vertical lines. High-grade rocks of the Cape Breton highlands are hachured with horizontal lines.

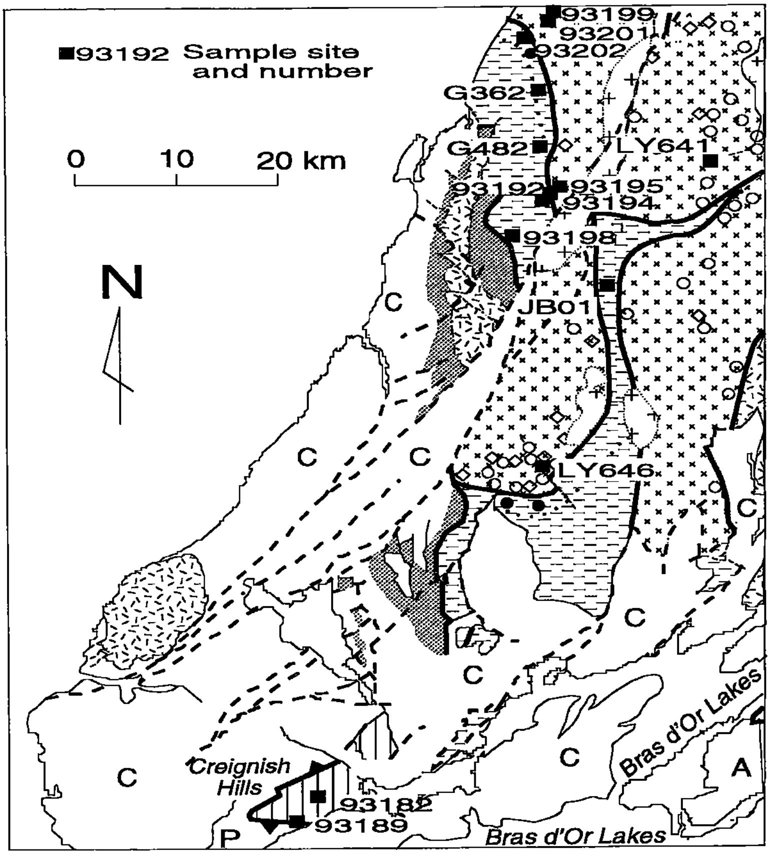


FIG. 2. Sample-location map. Patterns are the same as in Figure 1.

time. A U–Pb monazite age of 411 ± 2 Ma from kyanite – K-feldspar gneiss (Barr & Jamieson 1991) may date near-peak metamorphic conditions, whereas many $^{40}Ar–^{39}Ar$ plateau ages for hornblende and biotite fall in the range 395–370 Ma (Jamieson *et al.* 1986, Reynolds *et al.* 1989, Dallmeyer & Keppie 1993), which these authors interpreted as the time of cooling through the 400°C isotherm. Emplacement of plutons that penetrate both the Ordovician–Silurian suite and the allochthonous high-grade metamorphic suite accompanied and followed metamorphism (405–365 Ma: Jamieson *et al.* 1986, Dunning *et al.* 1990). These late plutons are essentially post-tectonic, tend toward A-type, and have andalusite–cordierite contact aureoles, suggesting emplacement at pressures of <2.5 kilobars (Plint & Jamieson 1989).

Exhumation of the plutonic rocks had commenced prior to deposition of the Late Devonian Fisset Brook Formation, which locally rests on low-grade Proterozoic plutonic rocks, but is nowhere found in stratigraphic contact with rocks of the Ordovician–Silurian or allochthonous high-grade suites. Exhumation to surface occurred mainly along the low-angle, extensional, greenschist-grade Margaree Shear Zone (Lynch & Tremblay 1994), which cuts the Fisset Brook Formation but predates lower Carboniferous sedimentation.

CONDITIONS OF METAMORPHISM

Previous investigators of late Silurian–Devonian metamorphism in western Cape Breton Island agreed that (a) metamorphic grade increases from west to east, (b) amphibolite-facies conditions (700–750°C, 8–10 kilobars) were reached early in the metamorphic history, and (c) isograds were disrupted and telescoped across faults during tectonic stacking (Craw 1984, Currie 1987, Plint & Jamieson 1989). These authors differed on whether the rocks affected by metamorphism could all be correlated. Craw (1984) and Plint & Jamieson (1989) argued that despite disruption by faults, both high- and low-grade rocks exhibit a single P–T path, and originally formed part of the

FIG. 3. Calcareous gneiss, allochthonous high-grade suite, northwestern Cape Breton Island (specimen 93195). Large, irregular poikilitic grains of garnet (G) in contact with amphibole (A), pyroxene (P), plagioclase (F), biotite (B) and calcite (C). Clear areas are quartz (Q). Plane-polarized light. Width of field of view: 4 mm.

Ordovician–Silurian suite. They considered contact metamorphism around plutons to be a later and minor phenomenon. On the basis of stratigraphic correlation and P–T determinations, Currie (1987) argued that the allochthonous high-grade rocks are not part of the Ordovician–Silurian suite, and had been metamorphosed earlier. He considered that the later and lower-grade metamorphic assemblages affecting the Ordovician–Silurian suite passed continuously into andalusite–cordierite-grade assemblages around the plutons.

In this paper, we consider (a) conditions of metamorphism within the Ordovician–Silurian and allochthonous high-grade suites, and relations between them, as well as (b) how late Devonian extensional deformation and exhumation are related to slightly earlier metamorphism.

The allochthonous high-grade suite was sampled at five points along its western edge (Fig. 2). Four samples were taken from thin layers of quartz-rich calcareous rocks, and a fifth from garnet amphibolite. The calcareous samples (described and located in the Appendix), collected over a distance of more than 50 km, represent rock types either not represented, or very rare, in the Ordovician–Silurian suite. Calcareous rocks in the allochthonous high-grade suite form discrete olive-green layers, or elongate boudins, up to 30 cm thick and tens of m long, composed of distinctive, hard,

FIG. 4. Garnet amphibolite, allochthonous high-grade suite, western Cape Breton Island (specimen 93192). Note the large grain of garnet (G) poikilitic with quartz (Q), and the aligned prisms of amphibole (A). Clear areas are plagioclase (P). Plane-polarized light. Width of field of view: 4 mm.

coherent, quartz-rich granoblastic rocks. Discontinuous layering on a scale of a few millimeters occurs between quartz-rich layers and amphibole + epidote-bearing layers (Fig. 3). The latter contain plagioclase, calcite, and large, fractured grains of pale pink to orange garnet. Clinopyroxene, complexly intergrown with epidote, occurs in two specimens. It was not found in contact with other ferromagnesian phases. The garnet amphibolite has a strong *L>S* fabric defined by elongate grains of amphibole and plagioclase (Fig. 4). Garnet is randomly distributed as small equant grains.

The mineral assemblages in both rock types are favorable for P–T determination using a multi-equilibrium technique (Berman 1991). This approach provides not only a P–T estimate, but also, if more than two independent equilibria can be computed, an estimate of error based on the scatter of the intersections. Data used to estimate P and T, given in Table 1, came from averaged rim compositions of pairs of minerals in contact, with the exception of clinopyroxene, found only in contact with epidote and quartz. Clinopyroxene was therefore excluded from P–T analysis, considerably reducing the scatter without significantly changing the estimated P–T. None of the minerals exhibits significant zoning, although a few grains of epidote exhibit a thin Al-poor rim. Compositions of grains of ferromagnesian minerals in contact do not differ significantly (more than 1σ) from compositions of isolated grains, suggesting

TABLE 1. RESULTS OF ELECTRON-MICROPROBE ANALYSES OF ASSEMBLAGES USED FOR P–T ANALYSES

Sample	SiO_2	TiO_2	Al_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	F	Cl	Total
93201												
am201	40.95	0.75	15.64	16.14	0.38	8.51	12.41	1.00	1.84	0.09	0.05	97.76
di201	52.23	0.05	0.58	10.28	0.43	11.64	24.44	0.12	0.01	0.00	0.00	99.78
grt201	38.05	0.16	21.52	25.63	2.87	3.67	8.13	0.00	0.02	0.00	0.00	100.05
pl201	48.18	0.00	32.22	0.27	0.00	0.00	15.48	2.87	0.09	0.00	0.00	99.11
ep201	37.62	0.07	22.80	12.85	0.09	0.04	23.41	0.00	0.01	0.00	0.00	96.89
93202												
bt202	36.30	2.11	19.23	17.00	0.13	10.89	0.01	0.30	9.27	0.05	0.02	95.31
ms202	47.23	0.61	36.72	0.80	0.02	0.60	0.01	1.30	9.48	0.00	0.02	96.79
grt202	37.64	0.02	21.30	29.87	5.48	3.98	1.96	0.00	0.00			100.25
pl202	59.11	25.77	0.06			7.47	7.30	0.17			99.88	
ilm202	0.04	52.10	0.34	45.85	0.99	0.08	0.04	0.00	0.01			99.45
ky202	37.63	0.02	62.19	0.22	0.02	0.02	0.02	0.04	0.10			100.26
rt202	0.00	99.17	0.08	0.12	0.01	0.02	0.04	0.00	0.01			99.45
LY641												
bt641	38.99	2.13	15.37	12.94	0.63	16.72	0.06	0.30	10.21	0.21	0.05	97.61
am641	41.81	0.69	15.90	15.15	0.27	10.51	11.06	1.93	0.39	0.00	0.01	97.72
ep641	37.68	0.01	22.59	13.22	0.05	0.01	23.44	0.01	0.01			97.02
grt641	38.17	0.10	21.60	23.49	4.72	5.18	6.43	0.00	0.01			99.70
pl641	54.52	28.59	0.05		10.86	5.31	0.10			99.13		
LY646												
am646	40.34	0.65	16.15	15.88	0.84	8.75	11.90	1.09	0.95	0.03	0.07	96.65
ep646	37.75	0.02	22.64	13.24	0.02	0.01	23.49	0.00	0.01			97.18
grt646	38.13	0.03	21.70	21.11	10.39	4.53	4.48	0.01	0.00			100.38
pl646	58.26	26.22	0.05			8.07	6.98	0.10			99.68	
G362												
bt362	35.73	1.77	18.61	19.34	0.00	9.70	0.03	0.24	9.10	0.49	0.03	95.17
ms362	46.88	0.70	35.27	1.28	0.03	0.74	0.00	1.58	9.30	0.00	0.00	96.25
grt362	36.79	0.04	20.86	37.12	0.56	3.47	1.27	0.02	0.06			100.24
pl362	64.05	21.82	0.01			3.86	9.75	0.04			99.53	
st362	27.24	0.47	55.09	14.79	0.20	1.46	0.00	0.01	0.00			98.79
ky362	37.63	0.12	61.87	0.27	0.02	0.02	0.02	0.01	0.02			99.98
G482												
bt482	36.32	1.72	17.78	19.39	0.13	11.95	0.04	0.27	9.02	0.30	0.06	96.98
ms482	46.50	0.27	35.37	2.24	0.00	0.56	0.02	2.10	8.32	0.00	0.06	95.42
grt482	36.79	0.04	20.86	36.56	0.56	3.07	2.27	0.02	0.06	0.00	0.00	100.23
pl482	63.45		22.41	0.11			3.96	9.17	0.61			99.71
st482	27.41	0.23	54.66	14.12	0.26	1.64	0.00	0.01	0.00			98.33
ky482	37.15	0.09	62.03	0.43	0.00	0.01	0.00	0.00	0.01			99.72
JB01												
btB01	36.05	1.85	19.67	18.68	0.02	9.70	0.11	0.15	8.43	1.67	0.11	96.44
msB01	46.65	0.30	35.85	0.77	0.00	0.64	0.03	1.39	9.51	0.22	0.00	95.36
grtB01	35.34	0.19	20.34	35.39	1.53	2.74	2.87	0.00	0.00	0.00	0.00	98.40
plB01	60.69	0.00	25.09	0.07	0.00	0.00	7.03	7.92	0.03	0.00	0.00	100.83
stB01	27.33	0.60	54.81	14.57	0.23	1.51	0.00	0.01	0.00	0.00	0.00	98.06
kyB01	37.48	0.01	61.89	0.37	0.02	0.00	0.00	0.00	0.00	0.00	0.00	99.97
93182												
am182	53.06	0.08	4.47	6.54	0.41	18.63	13.01	0.33	0.36	0.01	0.08	96.98
bt182	37.49	1.95	18.19	11.35	0.30	16.42	0.06	0.00	9.61	0.17	0.05	95.48
ms182	47.86	0.80	34.89	1.75	0.04	0.79	0.01	0.86	9.35	0.16	0.04	96.55
pl182	65.36	0.00	21.52	0.19	0.00	0.00	2.51	10.51	0.25	0.00	0.00	100.34
ep182	38.28	0.00	25.24	10.30	0.02	0.01	23.83	0.01	0.02	0.00	0.00	97.71
93189												
am189	56.01	0.04	0.83	8.95	0.17	18.72	13.16	0.13	0.00	0.00	0.00	98.01
ep189	38.45	0.03	27.64	7.06	0.12	0.03	23.96	0.00	0.03			97.32
chl189	25.69	0.07	19.17	28.77	0.44	13.47	0.04	0.03	0.03			87.71
pl189	60.44		24.97	0.09			6.58	7.64	0.24			
cal189				0.43	1.08	0.33	52.33					
93192												
am192	42.38	0.76	14.33	19.30	0.29	7.09	11.75	1.09	0.57	0.04	0.04	97.64
grt192	37.33	0.11	20.41	26.16	4.13	2.47	9.70	0.00	0.01	0.00	0.00	100.32
pl192	57.01	0.00	26.81	0.26			10.41	5.54	0.32			100.35
93194												
bt194	36.52	2.24	19.50	18.51	0.05	10.82	0.00	0.21	8.98	0.18	0.07	97.08
ms194	47.39	0.92	35.84	1.02	0.02	0.75	0.00	1.07	9.66	0.34	0.04	97.05
grt194	36.90	0.03	21.02	34.37	2.85	3.89	1.72	0.03	0.02			99.83
pl194	62.61		23.42	0.02			4.46	9.15	0.12			99.78
ilm194	0.31	52.07	0.07	45.51	1.39	0.01	0.00	0.00	0.00			99.36
rt194	0.00	99.29	0.07	0.34	0.03	0.02	0.03	0.00	0.02			99.80
sil194	37.57	0.02	62.74	0.24	0.01	0.03	0.03	0.00	0.00			100.64
93195												
am195	42.22	1.13	14.06	15.18	0.29	9.63	12.43	1.07	1.78	0.02	0.07	97.86
bt195	37.87	1.87	17.74	16.36	0.25	13.04	0.00	0.08	10.43	0.52	0.01	98.17
grt195	37.82	0.09	21.29	26.29	4.91	4.26	6.68	0.00	0.01			101.35
ep195	37.35	0.07	22.38	13.39	0.09	0.04	23.34	0.00	0.01			96.67
pl195	46.18		26.95	0.10			18.85	3.18	0.15			95.41
di195	49.29	0.07	1.10	19.96	0.30	5.79	23.16	0.14	0.01			99.82
93198												
am198	55.53	0.02	1.31	9.42	0.21	18.37	13.33	0.05	0.04	0.04	0.01	98.33
cal198				0.06	0.09	0.12	53.93					
ep198	38.33	0.05	26.30	9.73	0.09	0.00	25.02	0.00	0.01			99.53
chl198	28.65	0.01	19.43	14.77	0.06	24.18	0.02	0.01	0.02			87.15
pl198	60.44	24.97	0.09			6.58	7.64	0.24			99.97	
93199												
am199	53.59	0.11	4.51	6.59	0.39	18.25	12.88	0.33	0.36	0.01	0.08	97.02
bt199	37.11	1.87	18.01	13.08	0.31	15.38	0.05	0.00	9.53	0.22	0.09	94.64
ms199	48.33	0.81	35.23	1.77	0.04	0.80	0.01	0.87	9.44	0.17	0.02	97.47
pl199	65.36	21.52	0.19			2.51	10.51	0.25			100.34	
ep199	38.20	0.02	25.54	10.71	0.15	0.02	23.76	0.07	0.10			98.57

The symbols of Kretz (1983) are used; in addition, am is used to represent amphibole. Blank: not analyzed. See Appendix for specimen locations and descriptions. Total Fe reported as FeO. All compositions, quoted in wt.%, are the average of five or more determinations. The proportions of SrO, BaO, NiO, Cr_2O_3 and V_2O_5 were also established.

unusually rapid quenching during uplift. Results of P–T determinations are shown in Table 2 and Figure 5. All TWQ calculations use the following solution models: garnet: Berman (1990), Berman & Koziol (1991); biotite: McMullin *et al.* (1991); plagioclase: Fuhrman & Lindsley (1988); amphibole: Mäder *et al.* (1994); other minerals: ideal solid-solution.

With this exclusion, all five samples provide at least three independent equilibria, so that scatter of the intersections can be calculated. The samples give calculated pressures between 8.1 and 8.8 kilobars at temperatures of 700–745°C. The results have 1σ standard deviations of less than 40°C and 0.4 kilobars in T and P, respectively, indicating small scatter of the intersections. These results suggest that peak P and T in the allochthonous high-grade suite in this region everywhere exceeded 8.0 kilobars and 700°C. Owing to obvious retrogression and general lack of garnet, we found no samples of pelitic rocks appropriate for P–T determination, although several contain kyanite or sillimanite. Plint & Jamieson (1989) documented one pelitic specimen, and mention others (p. 415), giving P–T estimates in the range 9–10 kilobars and 700–750°C. Plint & Jamieson (1989) did not give sample locations, but R.A. Jamieson (written commun., 1996) has confirmed that these samples came from rocks which we consider to form part of the allochthonous high-grade suite.

Conditions of metamorphism in the low- to medium-grade parts of the Ordovician–Silurian suite

TABLE 2. PRESSURE–TEMPERATURE CONDITIONS FOR METAMORPHIC ROCKS FROM NORTHWESTERN CAPE BRETON ISLAND, AS CALCULATED BY THE METHOD OF BERMAN (1991)

Sample	T	dT	P	dP	n_r	n_i	n_x	Assemblage
				Allochthonous high-grade suite				
Calcareous gneiss								
93195	710	80	8.06	0.38	16	4	54	qtz,pl,grt,bt,hbl,ep,(di)
93201	700	23	8.18	0.26	24	4	103	qtz,pl,grt,hbl,ep,(di)
LY641	710	13	8.71	0.20	31	5	122	qtz,pl,grt,bt,hbl,ep
LY646	742	24	8.84	0.39	35	4	321	qtz,pl,grt,bt,hbl,ep
Garnet amphibolite								
93192	707	1	8.69	0.02	9	3	13	qtz,pl,grt,hbl
				Ordovician–Silurian suite				
Pelitic schist								
93194	611	49	6.34	0.73	11	3	26	qtz,pl,grt,bt,ms,sil,rt,ilm
93202	626	27	5.05	0.46	9	3	14	qtz,pl,grt,bt,ms,ky,rt,ilm
G362	610	17	5.93	0.26	6	3	12	qtz,pl,grt,bt,ms,(st)
G482	527	16	5.81	0.30	6	3	12	qtz,pl,grt,bt,ms,(st)
JB01	561	3	5.28	0.04	6	3	11	qtz,pl,grt,bt,ms,(st)
				Margaree Shear Zone (mol fraction of H_2O is assumed to be 0.98; see text)				
93182	362	-	2.39	-	3	2	1	qtz,pl,ep,bt,ms
93189	421	-	1.89	-	5	2	9	qtz,pl,tr,ep,cal,chl
93198	434	-	2.09	-	5	2	9	qtz,pl,tr,ep,cal,chl
93199	359	-	2.42	-	3	2	1	qtz,pl,ep,cal,chl

T: calculated metamorphic temperature in °C; dT: calculated 1 σ scatter in T (°C); P: calculated metamorphic pressure in kilobars; dP: calculated 1 σ scatter in P (kilobars). n_i: number of independent reactions; n_r: number of reactions considered; n_x: number of intersections considered in P–T analysis (intersections at >10° and within 1.5 σ of the calculated intersection).

Mineral abbreviations: qtz quartz, pl plagioclase, grt garnet, bt biotite, di diopside; hbl hornblende, ms muscovite, sil sillimanite, ky kyanite, tr tremolite, cal calcite, ep epidote, chl chlorite, rt rutile, ilm ilmenite, st staurolite.

The following solution models have been used; plagioclase: Fuhrman & Lindsley (1988); biotite: McMullin *et al.* (1991); muscovite: Chatterjee & Froese (1975); garnet: Berman & Koziol (1991); amphibole: Mäder *et al.* (1994). For other phases, ideal ionic solution has been assumed.

are well established by the work of Currie (1987) and Plint & Jamieson (1989). However, maximum pressures and temperatures attained within the suite are controversial, because Currie (1987) argued that there was a significant gap between these conditions and those within the allochthonous high-grade suite, whereas Plint & Jamieson maintained that there was a continuum of conditions, or had been prior to tectonic dislocations. The transition between the two suites lies within a broad retrograde shear-zone. Cataclastic deformation and retrogression make P–T determination within this shear zone impracticable. However, we sampled pelitic rocks of the Ordovician–Silurian suite (sample 93194) within 200 meters of a specimen from the allochthonous high-grade suite (93195), for which P and T are well defined. Another specimen of the Ordovician–Silurian suite (93202) lies about 2.5 km from a calibrated locality within the allochthonous high-grade suite (93201), but both lie less than 100 meters from the shear zone separating the two suites. Samples 93194 and 93202 both consist of coarse pelitic schist containing aligned biotite and muscovite, with a matrix of quartz and plagioclase, abundant disseminated, equant grains of garnet, and nebulous, thin quartzofeldspathic segregations (Fig. 6). Sample 93194 contains sillimanite, whereas sample 93202 contains kyanite. Both contain accessory rutile + ilmenite. The grains of garnet are slightly zoned, with a poikiloblastic core about 3% richer in spessartine component than the dense, thick rim. Minerals are otherwise essentially unzoned. P–T determinations based on four independent equilibria in these specimens (Table 2, Fig. 7) give well-defined intersections with similar temperatures (611 and 626°C) at pressures of 6.3 and 5.1 kilobars, respectively. Some high-grade parts of the Ordovician–Silurian suite consist of distinctive knobby schists containing the assemblage biotite – muscovite – staurolite – garnet – kyanite. Three P–T determinations were made on this assemblage (Table 2, Fig. 8) using the garnet – biotite and biotite – muscovite – garnet – plagioclase geothermobarometers. These gave similar temperatures and pressures (527–610°C, 5.3–5.9 kilobars). Addition of staurolite to the analysis, with assumed activity of H_2O of 1.0 (giving maximum P–T values), gives similar values, but with much larger scatter of intersections. These results are consistent with those obtained by Plint & Jamieson (1989) for "medium-grade" rocks in the same area (5.1–6.4 kilobars at 553–640°C). P–T determinations therefore suggest that peak P–T conditions in the Jumping Brook suite did not significantly exceed 6.4 kilobars and 640°C. These values are markedly lower in both P and T than those found for the allochthonous high-grade suite. The extent of the gap is shown in Figure 8.

For many samples of the Ordovician–Silurian and allochthonous high-grade suites, uncertainties in P and T estimates due to scatter in intersections of calculated equilibria are small (<0.4 kilobars and <30°C). These values do not give an estimate of the errors in determination of P and T, because they do not include uncertainties in chemical analyses or thermodynamic parameters. Since thermodynamic parameters are incorporated into a self-consistent database (Berman 1991), we assume that adjustments in these parameters will tend to shift all P and T estimates involving particular minerals by similar amounts. In particular, we assume that these adjustments will not systematically displace the pressures and temperatures for the Ordovician–Silurian suite relative to those of the allochthonous high-grade suite, because both depend heavily on the properties of garnet and biotite. Possible adjustments to the thermodynamic database are unlikely to affect the gap in P–T estimates in the Ordovician–Silurian and allochthonous high-grade suites, although they may shift both sets of estimates by an uncertain amount. To assess sensitivity of P–T estimates to analytical errors, we have recalculated the intersections with adjusted Fe/(Fe + Mg) for ferromagnesian minerals and adjusted An content for plagioclase. We increased or decreased these quan-

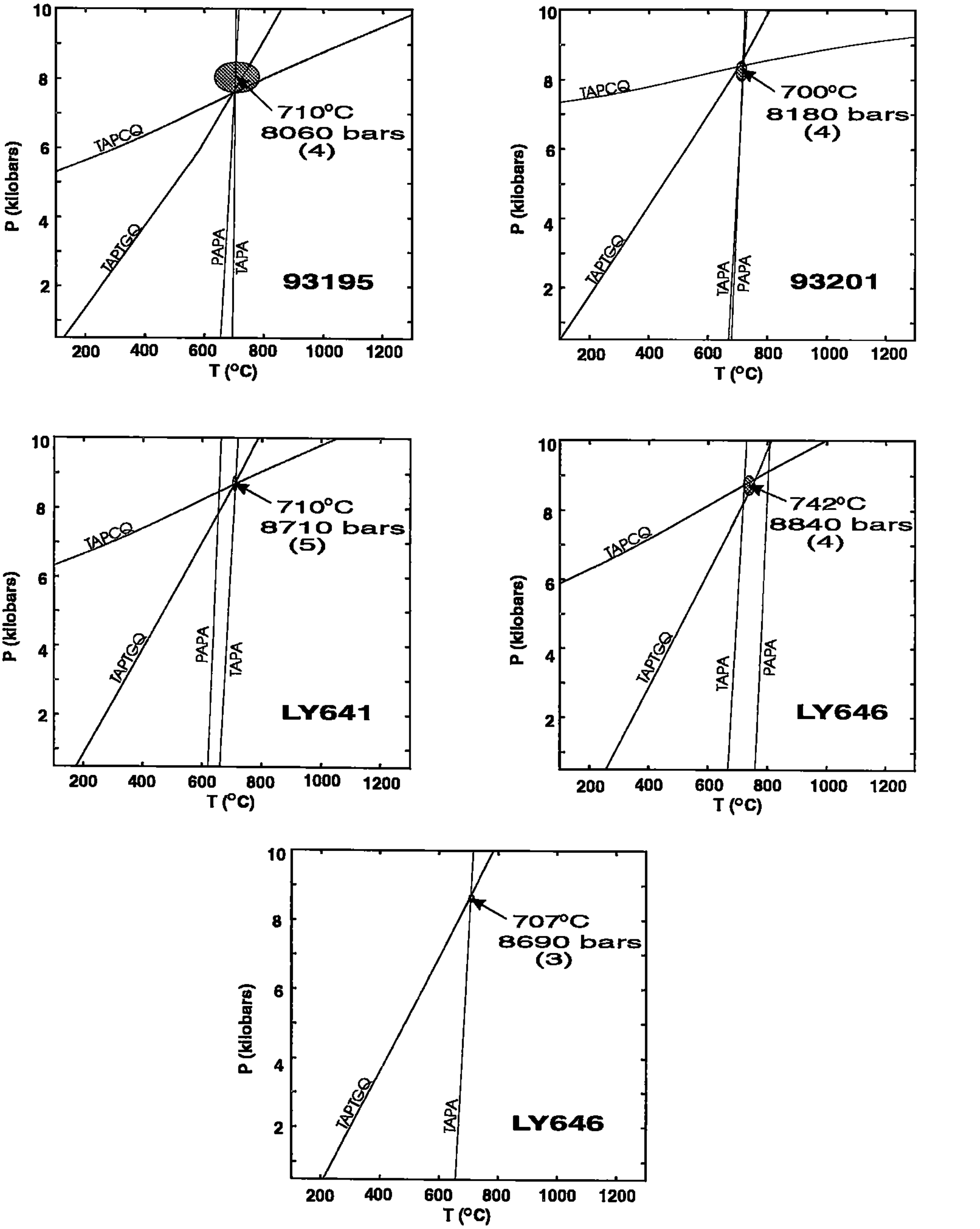


FIG. 5. Selected equilibria defining P–T conditions in calcareous gneiss and garnet amphibolite of the allochthonous high-grade suite. Shaded ellipses have axes equal to 2σ scatter in P and T, as calculated by the program INTERSX (Berman 1991). The P and T estimated by this program from all relevant equilibria lie at the center of the ellipse. The number of independent reactions is shown beneath the calculated P–T. The plotted equilibria are believed to be the most robust and best calibrated of the set considered. They are identified as follows: PAPA: pyrope + annite = almandine + phlogopite; TAPA: 3 tremolite + 5 almandine = 5 pyrope + 3 ferro-actinolite; TAPTGQ: 3 tremolite + 12 anorthite = 2 pyrope + 3 tschermakite + 4 grossular + 12 quartz; TAPCQ: tschermakite + 2 anorthite = pyrope + 2 clinozoisite + quartz. Figures 5, 7 and 9 assume the following solution models: garnet: Berman (1990), Berman & Koziol (1991), biotite: McMullin *et al.* (1991), plagioclase: Fuhrman & Lindsley (1988), amphibole: Mäder & Berman (1992). For other phases, ideal ionic solution was assumed.

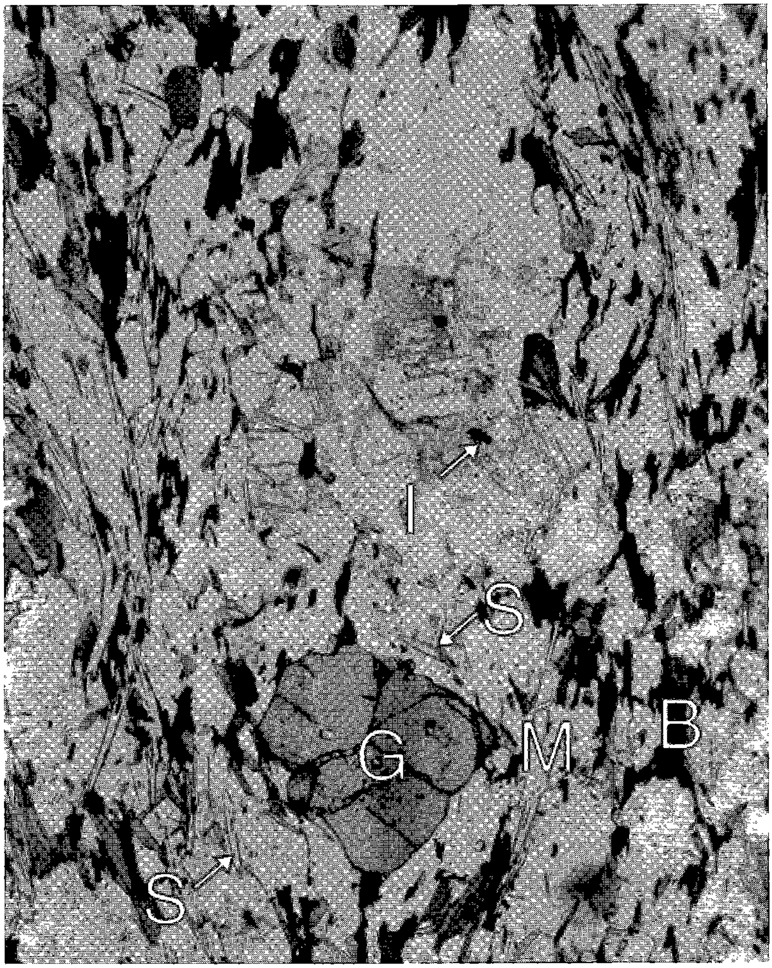


FIG. 6. Pelitic schist, Ordovician–Silurian suite, western Cape Breton Island (specimen 93194). The matrix consists of coarse quartz and twinned plagioclase, with aligned grains of intergrown biotite (B) and muscovite (M), which bend around large, irregular grains of garnet (G). Sillimanite (S) prisms appear to be concentrated around the garnet. Small opaque grains are ilmenite (I). Plane-polarized light. Width of field of view: 4 mm.

tities by 0.02 for one mineral at a time and examined the results. The calculated intersections moved by amounts varying from 15 to 70° and 150 to 800 bars, with displacements clustering near 40° and 600 bars. We consider that these values give a reasonable estimate of uncertainty due to analytical error. If one accepts that some uncertainty is allowed for the thermodynamic parameters, the P–T determinations are probably correct within 50° and 800 bars. Since the gap in Figure 8 between calculated P–T for the Ordovician–Silurian and allochthonous high-grade suites is roughly twice this amount (~75° and 2 kilobars), it is improbable that a continuum of conditions between them is represented in outcrop, as assumed by Plint & Jamieson (1989). We conclude that the data strongly suggest the existence of significantly differing conditions of metamorphism in the two suites.

Little attention has previously been paid to conditions of metamorphism during development of late Devonian extensional shear-zones, such as the Margaree Shear Zone, which contributed to denudation of the high-grade metamorphic suites, but themselves contain greenschist-grade assemblages. We sampled these structures in four places, two in the Cape Breton Highlands (specimens 93198, 93199), and two in the Creignish Hills (specimens 93182, 93189). All specimens are fine-grained greenish mylonitic schists, with felsic streaks. Two assemblages are

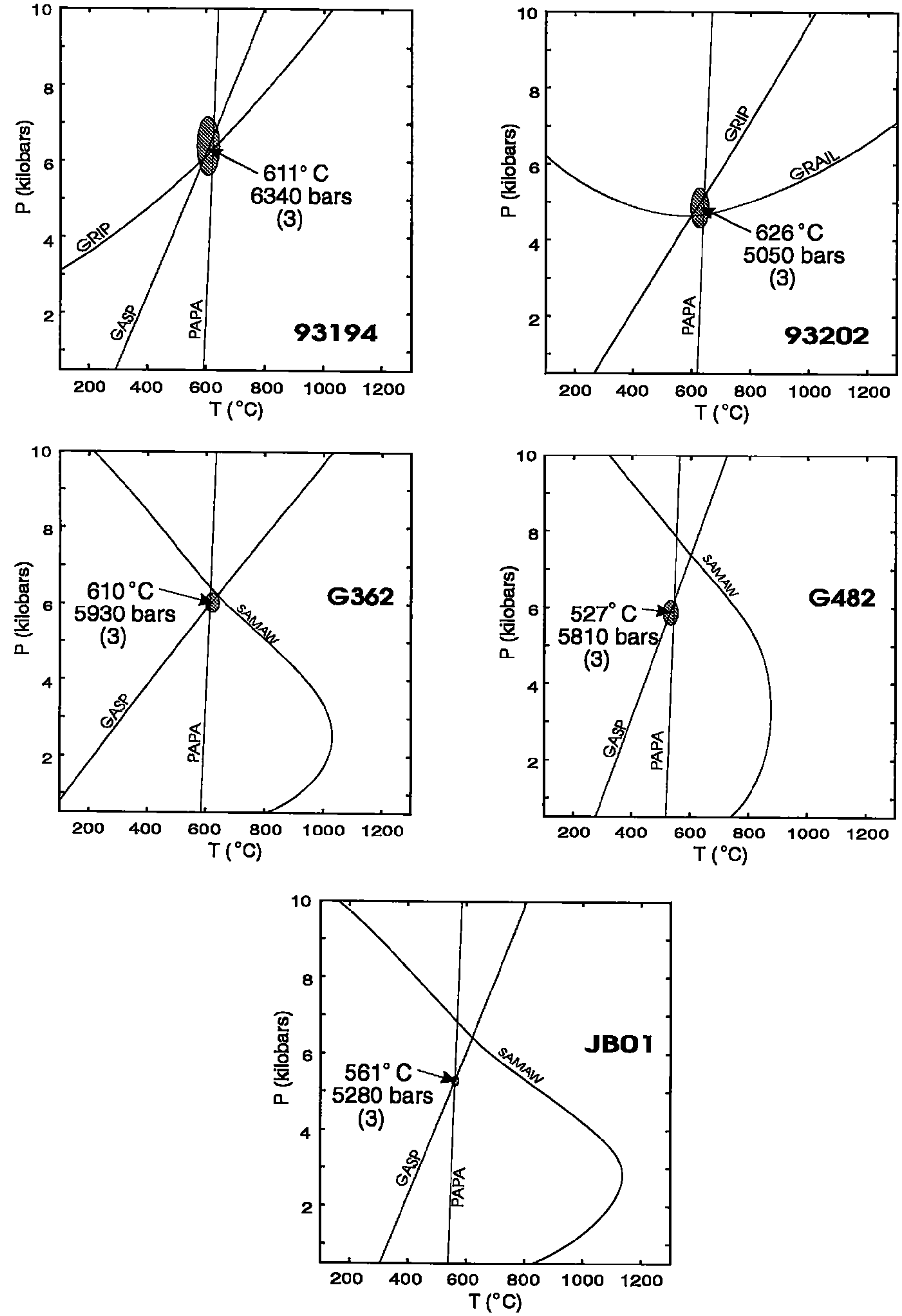


FIG. 7. Selected equilibria defining P–T conditions in the Ordovician–Silurian suite. Shaded ellipses have axes equal to 2σ scatter in P and T, as calculated by program INTERSX (Berman 1991). The P and T estimated by this program from all relevant equilibria lie at the center of the ellipse. The number of independent reactions is shown beneath the calculated P–T. The plotted equilibria are believed to be the most robust and best calibrated of the set considered, with the exception of the staurolite-containing reaction, shown for illustrative purposes only. Equilibria (in addition to those in Fig. 5) are identified as follows: GRIP: 3 anorthite + 6 ilmenite + 3 quartz = grossular + 2 almandine + 6 rutile; GASP: grossular + 2 sillimanite + quartz = 3 anorthite; GRAIL: 3 ilmenite + 2 quartz + sillimanite = almandine + 3 rutile; SAMA W: 12 staurolite + 23 annite = 31 almandine + 23 muscovite + 6 water.

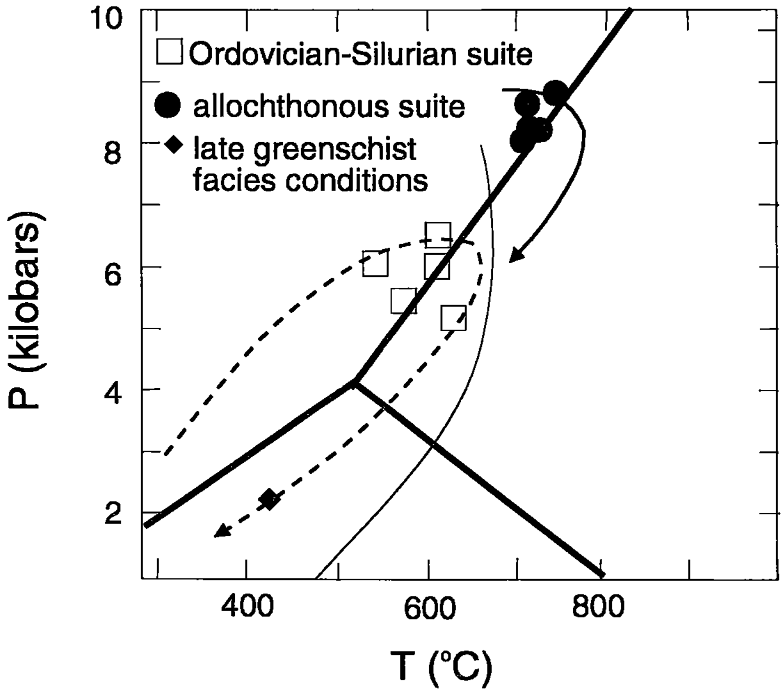


FIG. 8. Comparative P–T conditions in the allochthonous high-grade and Ordovician–Silurian suites. The heavy line shows fields of stability of the aluminosilicate polymorphs, whereas the light line shows the maximum stability of staurolite + quartz relative to almandine + aluminosilicate. The dashed line shows a possible P–T–t path for the Ordovician–Silurian suite, whereas the full line shows part of a possible P–T–t path for the allochthonous high-grade suite.

present in thin section, one a fine-grained laminated matrix of epidote, actinolite, chlorite and minor calcite (Fig. 9) around larger crystals of plagioclase, the other a mat of fine-grained aligned flakes of biotite and muscovite around small porphyroblasts of epidote. Recrystallized quartz and plagioclase also are present in both assemblages. In three samples, these phases appear to be essentially unzoned in composition (Table 1). However a fourth sample (93198) contains coarse, spectacularly zoned amphibole that varies from hornblende in the core, similar to amphibole grains in the high-grade allochthonous suite, to an Al-free actinolitic rim (Table 3). We conclude that this sample originated in the high-grade allochthonous suite, as did augen of high-temperature orthoclase-bearing ribbon mylonites, typical of the shear zones bounding the suite (Lynch & Tremblay 1994). We have confined P–T determination in sample 93198 to amphibole rims, which are assumed to be in equilibrium with the other low-grade phases in the specimen. All potentially useful equilibria in the mylonites involve H_2O and CO_2. The activity of these components must therefore be established. The observed mica-bearing assemblages

TABLE 3. CHEMICAL COMPOSITION OF ZONED AMPHIBOLE FROM SPECIMEN 93198

	1	*2*	*3*	*4*
SiO_2 wt.%	44.32	51.68	53.87	55.53
TiO_2	0.34	0.08	0.08	0.02
Al_2O_3	12.42	3.80	2.61	1.31
FeO	15.20	13.21	10.39	9.42
MnO	0.24	0.17	0.23	0.21
MgO	12.06	15.10	17.30	18.37
CaO	11.29	12.99	13.02	13.33
Na_2O	1.27	0.53	0.22	0.05
K_2O	0.25	0.09	0.06	0.04
F	0.00	0.00	0.00	0.00
Cl	0.01	0.02	0.01	0.00
BaO	0.02	0.07	0.02	0.04
Cr_2O_3	0.08	0.03	0.00	0.01
NiO	0.05	0.06	0.05	0.05
SrO	0.01	0.00	0.00	0.01
V_2O_3	0.09	0.01	0.06	0.02
Total	97.65	97.84	97.92	98.41

1. Colorless core of amphibole prisms (compare Table 1, compositions of amphibole of the high-grade allochthonous suite).
2-3. Compositions of the intermediate zones.
4. Rim of blue-green actinolite used for geothermometry.
Total Fe reported as FeO. Each composition reported is the average of at least five electron-microprobe spot-analyses.

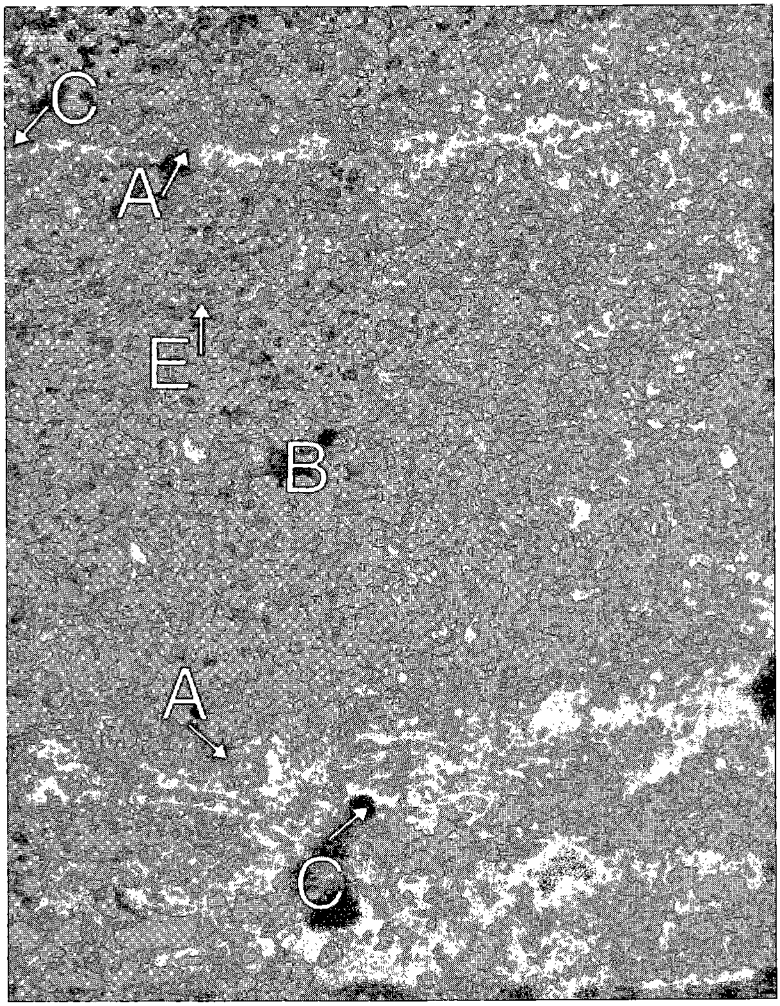


FIG. 9. Mylonitic schist, western Cape Breton Island (specimen 93189). The mosaic matrix (quartz + plagioclase) is cut by mafic septa composed of relatively large grains of amphibole (A), intergrown with chlorite (C). Epidote (E), calcite and biotite (B) are more evenly disseminated. Plane-polarized light. Width of field of view: 4 mm.

are insensitive to fugacity of H_2O, giving results varying by only a few degrees and a few hundred bars for mole fractions of H_2O between 0.95 and 0.98. However, the quartz – plagioclase – tremolite – epidote – chlorite – calcite assemblages give intersections only for temperatures between 385 and 440°C, and mole fractions of H_2O in the fluid phase between 0.94 and 0.99 (Fig. 10). We have, therefore, taken the mole fraction of H_2O to be 0.98, and taken the pressure to be given by the mica-bearing assemblages. The pressure of the chlorite-bearing aseemblages is not closely constrained, because the relatively low angle of the intersections makes them sensitive to small analytical errors.

Our estimates of pressure for greenschist-grade shear zones agree quite closely with those of Plint & Jamieson (1989) for andalusite–cordierite aureoles around late Devonian plutons [<2.5 kilobars according to Plint & Jamieson (1989)], indicating that pressures in the metamorphic rocks had fallen to <2.5 kilobars by late Devonian time from >5.5 kilobar in early Devonian time. This finding requires a decrease in overburden from ~20 kilometers to <7.5 kilometers

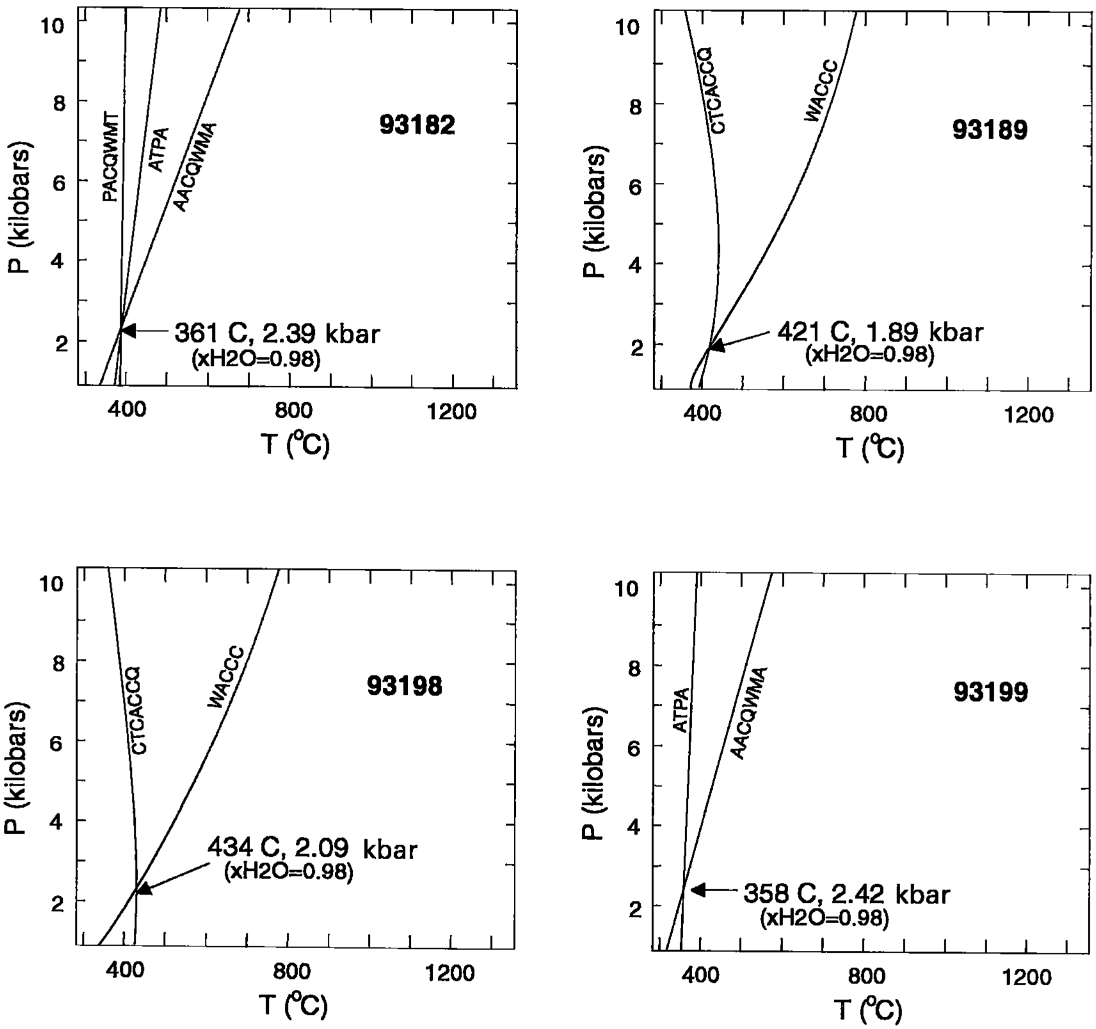
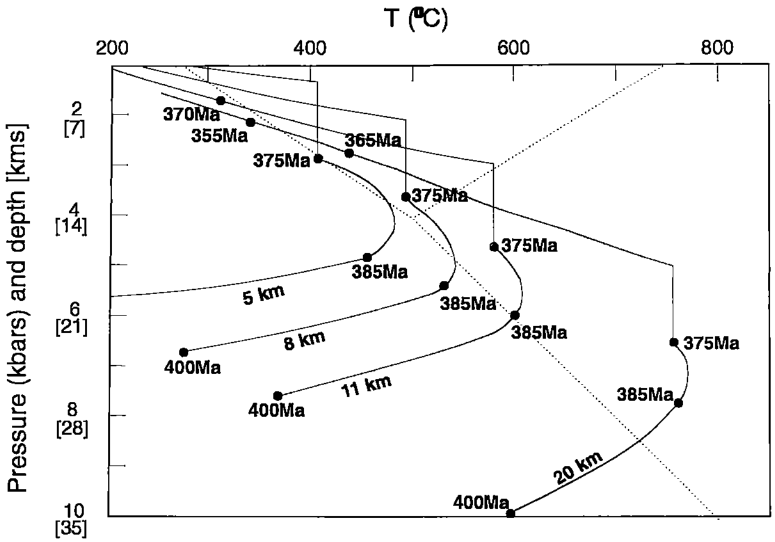


FIG. 10. Representative equilibria illustrating P–T conditions in the mylonitic schist. The assumed mole fraction of H_2O in the fluid phase is 0.98 (see text). Equilibria are identified as follows: ATPA: 5 annite + 3 tremolite = 5 phlogopite + 3 ferro-actinolite; PACQWMT: 5 phlogopite + 2 anorthite + 2 clinozoisite + 14 quartz + 2 water = 5 muscovite + 3 tremolite; AACQWMA: 5 annite + 2 anorthite + 2 clinozoisite + 14 quartz + 2 water = 5 muscovite + 3 ferro-actinolite; WACCC: water + 3 anorthite + calcite = 2 clinozoisite + CO_2; CTCACCQ: 6 clinozoisite + tremolite + 6 CO_2 = 8 anorthite + 6 calcite + clinochlore + 7 quartz.

between ~400 Ma and 375 Ma, a rate of uplift of about 500 meters per million years.

Discussion

Plint & Jamieson (1989) used a one-dimensional thermal model to conclude that a hot allochthon was necessary to explain isograds in the Ordovician–Silurian suite, and that tectonically driven denudation was required to explain its cooling history. We agree with these conclusions. The requirement for tectonic denudation can be seen from Figure 11, calculated using the program of Haugerud (1986). This model assumes a slab of crust 16 km thick emplaced at 400 Ma, a constant rate of erosion of 500 m/Ma (see above), and thermal properties (listed on Fig. 11) in the middle of the range considered by Plint & Jamieson (1989). The assumed rate of isostatic denudation lies in the midst of, but toward the high end, of the range discussed by England & Thompson (1984). Without an assumed 6 kilometers of instant uplift at 375 Ma, strata beneath the allochthon would still be in the greenschist facies in



Model parameters:

Thermal conductivity	2.50 W/m/K	Thickness of crust	32 km
Heat content	2.25x10 J/K/m^3	Erosion rate	500 m/Ma
Radiogenic heat production	2.00 μW/m^3	Thickness of thrust slab	16 km
Mantle heat flow	30 mW/m^2	Time of instant uplift	375 Ma
Thickness of heat-producing layer	16 km	Amount of instant uplift	6 km

FIG. 11. One-dimensional thermal model of western Cape Breton Island. A layer of crust 16 km thick is assumed to be emplaced on top of existing crust instantaneously at 400 Ma, following which erosion takes place at the rate of 500 m/Ma until 375 Ma, at which time 6 km of crust is instantaneously removed, and erosion continues at the same rate as before. The relevant thermal parameters are shown on the figure. The km figures on the P–T–t trajectories represent the distance below the base of the allochthon at the time of emplacement. The stability fields of the aluminosilicate polymorphs are shown in dotted lines. Calculations were made using the program of Haugerud (1986).

Carboniferous time (350 Ma), contrary to the observation that detritus of high-grade metamorphic rocks occurs in basal Carboniferous strata. We attribute this tectonic denudation to extensional movement on the Margaree Shear Zone and related structures, which can in turn be related to gravitational collapse of the orogen (Lynch 1996). In our model (Fig. 11), strata up to 20 kilometers beneath the allochthon reached a thermal maximum at 375–390 Ma and moderate pressures (4–7 kbars), and passed through the 400°C isotherm by 365 Ma. Rocks less than 11 km below the allochthon were at pressures of less than 3 kilobars by 375 Ma. Rocks originally more than ~8 km below the base of the nappe made a brief (4–20 Ma) transit through the sillimanite field. Such a thermal history could explain the local presence of significant sillimanite (specimen 93194) in the generally kyanite-bearing Ordovician–Silurian suite, although the observed distribution of Al_2SiO_5 polymorphs is not exactly in accord with the phase diagram shown on Figure 11. The generally unoriented character of the sillimanite may be due to formation during rapid uplift under conditions of overall extension.

Plint & Jamieson (1989) considered that all metamorphic rocks in the western Cape Breton Highlands exhibit a single P–T path. In their view, no part of their deduced hot allochthon remains. Our data make this model implausible. The gap in peak pressures of metamorphism between the Ordovician–Silurian suite and structurally overlying rocks (>1.5 kilobars) makes it evident that a continuum of metamorphic conditions did not exist, and suggests that the overlying rocks formed part of the hot allochthon invoked by Plint & Jamieson (1989). There is no geochronological evidence for metamorphism in the Ordovician–Silurian suite prior to 411 Ma, suggesting that this hot allochthon was emplaced at ~411 Ma. The maximum

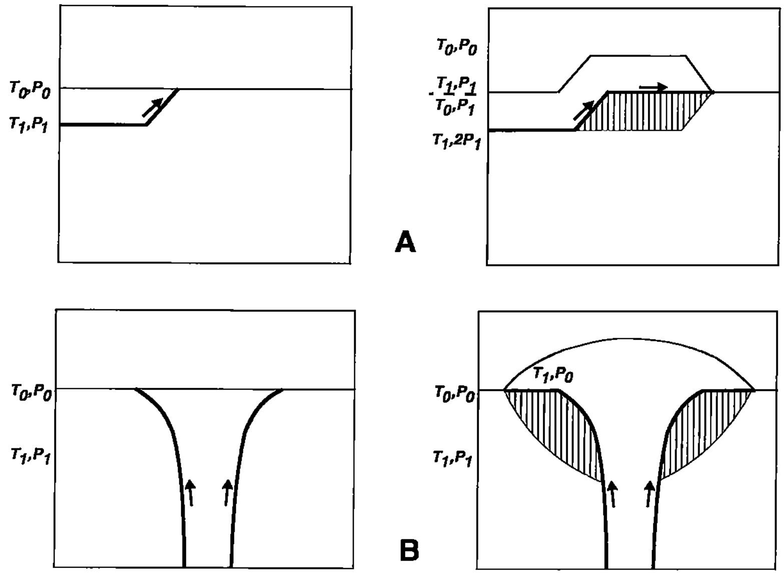


FIG. 12. Conceptual models of metamorphism by a hot allochthon. The metamorphosed region is hachured. A. Laterally transported allochthon. Assuming instantaneous transport, note that the surface overridden by the allochtnon, originally at T_o,P_o, is raised to pressure P_1 while the temperature remains at T_o pending thermal relaxation. Assuming originally horizontal isotherms, the level originally at T_1,P_1 goes to T_1, $2P_1$. Such an allochthon, therefore, tends to produce relatively high-pressure metamorphism. B. Vertically transported allochthon. It is assumed that the top of the allochthon spreads owing to gravitational collapse. Again assuming instantaneous transport, note that material originally at T_1,P_1 can reach the surface still at temperature T_1. The pressures reached below the allochthon depend on the details of the spreading, but are always less than in case A because the allochthon thins as it spreads. This type of allochthon thus tends to give rise to high-temperature metamorphism at relatively low pressures.

pressures observed in the Ordovician–Silurian suite (~6.5 kilobars) require an allochthon ~20 kilometers thick. The source of this allochthon remains conjectural. Lynch (1996) proposed an allochthon model based on horizontal transport (conceptual cartoon shown in Figure 12A), which involves early thrusting and late extensional tectonics. Lin & van Staal (in press) pointed out that the model of Lynch (1996) appeared to conflict with structural observations in the southern and eastern Cape Breton Highlands. Dallmeyer & Keppie (1993) suggested uplift of the central highlands on a positive flower-structure, possibly assisted by thermal buoyancy over a hypothetical thermal anomaly. Such a configuration could bring hot rock to the surface by a combination of flow and steep reverse faulting, as shown conceptually in Figure 12B. In this model, overthrusting in the western highlands resulted from gravitational collapse and spreading of the hot rock.

Fragments from middle Devonian greenschist-grade shear zones occur in basal Carboniferous strata. To reach the surface, these post-peak-metamorphic shear zones required 6–8 kilometers (~2–2.5 kilobars) of post-metamorphic denudation (uplift), mainly by extension on the Margaree Shear Zone (Lynch & Tremblay 1994). In both models, this post-metamorphic history can be explained by isostatic re-adjustment and gravitational collapse of the hot allochthon. The ~20 kilometers (6.4 kilobars) of denudation in Devonian time can be divided into ~14 kilometers (4 kilobars) prior to formation of the greenschist-grade shear zones, presumed to be due to isostatic uplift and erosion of the thickened crust, and an additional ~6 kilometers due to late Devonian displacement on the shear zones.

CONCLUSIONS

Peak conditions of metamorphism within quartz-rich calcareous rocks and garnet amphibolite of the allochthonous high-grade suite exceeded 8 kilobars and 700°C. Maximum P–T conditions in tectonically underlying rocks of the Ordovician–Silurian suite did not exceed 640°C, 6.4 kilobars. Geochronological data define an earliest Devonian age for the peak of metamorphism. Metamorphic rocks had been uplifted ~14 kilometers (~4 kilobars) by middle Devonian time. Extension on greenschist-grade shear zones (~415°C, ~2 kilobars), formed in middle to late Devonian time, caused an additional ~6 kilometers of uplift, returning the metamorphic rocks to surface prior to early Carboniferous sedimentation. These data can be explained by transport of a thick (~20 km), hot allochthon onto the Orodovician–Silurian suite, followed by rapid denudation due to isostatic uplift, followed by extensional faulting. Our data do not clearly distinguish between predominantly horizontal and predominantly vertical transport for the allochthon.

Acknowledgements

We are indebted to John Stirling for assistance with electron-microprobe analyses, and to Rob Berman for discussions regarding geothermobarometry. We are also indebted to Shoufa Lin for careful reading and correction of an earlier version of the manuscript, and for a preprint of his discussion (with C.R. van Staal) of allochthon models of the highlands. The manuscript has been greatly improved by critical reviews from Rob Raeside and Urs Mäder, and by editorial comments by Rob Berman. None of these persons are responsible for the interpretations expressed in this paper.

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APPENDIX: LOCATION AND DESCRIPTION OF ANALYZED SPECIMENS

93182 (45°57'13"N, 61°14'52"W, woods road on Skye Mountain). Lineated biotite mylonite with rootless sheath folds. Specimen is pale green, siliceous and calcareous. In thin section, the rock consists of a fine-grained mosaic of quartz and minor cloudy plagioclase, separated into lenses by braided, wispy, sinuous films of aligned muscovite and chloritized biotite, with scattered granules of epidote, titanite and carbonate.

93189 (45°55'28"N, 61°14'50"W, River Denys, aproximately 3 km upstream from the Trans-Canada Highway). Mylonite, layered on centimeter scale, of shattered granite in a dark chloritic matrix. In thin section, the matrix is a fine-grained mosaic of quartz and feldspar, with some slightly coarser lenses of quartz, all separated into elongate lenses by a network of chlorite in ragged, unoriented grains. This network contains abundant granules of magnetite and epidote and sparse carbonate. Carbonate also occurs in thin veinlets at relatively high angles to the foliation.

98192 (46°34'40"N, 60°52'32"W, Belle Cote road, northeast of Jim Campbell's Barren). Finely laminated, strongly lineated, medium-grained migmatitic garnet amphibolite cut by a multitude of layers and low-angle veins of pink medium-grained granitic rocks. In thin section, the rock consists of strongly aligned olive green hornblende, lensoid aligned quartz, and erratically twinned plagioclase. Garnet forms disseminated, dense, euhedral grains.

93194 (46°34'57"N, 60°52'12"W, Belle Cote road, northeast of Jim Campbell's Barren). Screen of coarse pelitic schist 4 m wide in migmatitic granite. The rock consists of coarse, aligned biotite, muscovite and quartz with disseminated large grains of garnet. In thin section, quartz and twinned plagioclase are elongate, polygonal and sutured. Biotite is red-brown with intergrown muscovite. Sillimanite forms euhedral scattered prisms that overgrow the cleavage. Garnet forms dense euhedra, with a small poikilitic core. Small angular grains of ilmenite and needles of rutile appear to be associated with sillimanite. Accessory tourmaline and altered cordierite are present.

93195 (46°35'03"N, 60°52'04"W, Belle Cote road, 2 km northeast of Jim Campbell's Barren). Mesocratic biotite schist with competent, little-deformed, calcareous layers up to 2.5 cm thick, which contain quartz, plagioclase, hornblende, biotite, garnet, diopside and calcite. In thin section, the rocks are granoblastic, but with a distinct fabric, and the minerals are cuspate and poikiloblastic, but unzoned. Clinopyroxene is colorless, and locally elongate across foliation. Epidote exhibits an anomalous blue interference color, and forms a symplectitic intergrowth with clinopyroxene. Weakly pleochroic amphibole forms both large prisms and interstitial material. Garnet forms large grains, with inclusions of amphibole and titanite. Carbonate is present in minor amount.

93198 (46°33'46"N, 60°53'48"W, Belle Cote road, Jim Campbell's Barren). A fissile, strongly lineated greenish black, fine-grained schist with visible needles of amphibole lying on the parting plane. In thin section, the rock consists of moderately aligned, locally radiating, plumose zoned amphibole in a chlorite-rich quartz–plagioclase mosaic matrix. Amphibole grades from a colorless core to a blue-green rim. Chlorite forms large poikiloblastic masses containing epidote and carbonate granules.

93199 (46°48'11"N, 60°50'48"W, Quarry on Cabot Trail 200 meters south of McKenzie's Mountain lookoff). Dark, mylonitic breccia with clasts of quartz and feldspar and polycrystalline fragments in a very fine-grained matrix. In thin section, the matrix consists of a very fine-grained, faintly foliated quartz–plagioclase matrix separated into microlithons by narrow muscovite–biotite septa that contain disseminated epidote and carbonate granules.

93201 (46°47'08"N, 60°50'03"W, 85 m north of the emergency cabin on the Cabot Trail). A prominent pale olive layer, about 20 centimeters thick, of weakly foliated quartz-rich hornblende – diopside – garnet rock within rusty biotite schist. Coarse granoblastic, almost nonpleochroic amphibole occurs together with slightly altered plagioclase, and poikilitic garnet. Aligned amphibole defines a weak foliation. Carbonate is present as relatively large but sparse grains. Diopside-rich clinopyroxene occurs intergrown with epidote. Titanite and microcline are common accessories.

93202 (46°44'55"N, 60°51'51"W, old Cabot Trail on French Mountain 250 m north of the old bridge over Fishing Cove River). The rock, a fairly homogeneous biotite–muscovite schist, contains stubby prisms of kyanite. In thin section, the rock is granoblastic, with a weak foliation due to moderately aligned lenses of polycrystalline quartz. Biotite is reddish brown and well disseminated. Round to subhedral grains of garnet exhibit a thick, inclusion-free rim and poikilitic core without helical structure. Plagioclase is well twinned. Kyanite forms poikiloblastic prisms associated with ilmenite dust and rutile needles.

G482 (46°40'06"N, 60°51'46"W, headwaters of Robert Brook). Coarse-grained silvery white schist with a knobby texture owing to large garnet and staurolite porphyroblasts, and clasts of recrystallized quartz. In thin section, abundant coarse flakes of muscovite and green biotite define the principal foliation, although plagioclase also is elongate. Garnet and staurolite are poikiloblastic with quartz. Kyanite is a minor component, as stubby prisms. Biotite and, locally, staurolite may be mildly chloritized.

G362 (46°42'40"N, 60°51'17"W, headwaters of Corney Brook). Coarse-grained, lineated pelitic schist with large porphyroblasts of garnet and staurolite. In thin section, abundant coarse muscovite and brown biotite, intergrown with quartz and plagioclase, define the foliation. Garnet and staurolite, poikiloblastic with quartz, exhibit sigmoidal trails of inclusions. Kyanite and tourmaline are minor components. Biotite and staurolite locally exhibit minor chloritization. (This sample lies within the shear zone separating high-grade and medium-grade suites).

JB01(46°29'41"N, 60°47'28"W, saddle of Second Fork and Coineach brooks). Green plagioclase – staurolite – biotite – muscovite schist with coarse quartz – garnet layers defining schistose fabric. In thin section, staurolite and garnet are inclusion-free. Biotite, locally chloritized, is less abundant than muscovite. Late, coarse porphyroblasts of chlorite are abundant. Trace amounts of illmenite and metamict zircon occur.

LY93-641 (46°38'35"N, 60°39'46"W, Highland Road 450 m south of Cheticamp Flowage). The rock is pale olive, hard, and strongly foliated on a mm scale, with small aligned prisms of amphibole, biotite and elongate quartz defining the foliation. In thin section, the rock is a seriate mosaic of quartz, epidote, blue-green amphibole, inclusion-free garnet, and mildly altered twinned plagioclase. Biotite occurs around amphibole. Epidote is colorless, and poikiloblastic with amphibole. Magnetite, calcite and microcline are widely distributed as small rounded grains.

LY93-646 (46°18'52"N, 60°53'17"W, Egypt Highland Road between Fionnar Brook and Middle River). In thin section, the rock is strongly foliated with packed, aligned prisms of greenish amphibole, interstitial biotite and slightly altered plagioclase, and disseminated equant grains of garnet and epidote. Garnet is commonly inclusion-free, but some grains contain a few large inclusions of quartz and plagioclase. Epidote locally forms a symplectitic intergrowth with amphibole. Calcite occurs as large anhedral grains. Magnetite dust occurs throughout.