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Fault-controlled emplacement of arc-related magmas along the Neoproterozoic northern Gondwanan margin: An example from the Antigonish Highlands, Nova Scotia

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Abstract

In the Antigonish Highlands, Nova Scotia, Neoproterozoic (ca. 620–605 Ma) magmatism is a local representative of regional arc-related tectonothermal activity that typifies Avalonia and related terranes located along the northern Gondwanan margin. Three distinct types of coeval magma are represented by volcanic rocks of the Georgeville Group and by syn- to post-tectonic plutons (e.g. Greendale Complex); (i) calc-alkalic subduction-related basaltic andesites and andesites, (ii) crustally-derived rhyodacites and rhyolites, and (iii) Fe–Ti rich continental tholeiites derived from the melting of subcontinental lithosphere. These magmas were probably generated in an extensional arc regime with deep-seated faults providing conduits for rising magmas. The Georgeville Group was polydeformed by thrusts, isoclinal folds and upright folds associated with movement along NE-trending faults very soon after it was deposited.

A 85-km-long 1–2 km wide magnetic anomaly transects Georgeville Group strata and connects outcrops of Fe–Ti rich continental tholeiites and ultramafic cumulate rocks of the Greendale Complex. The N–S orientation of this anomaly is compatible with coeval sinistral shear along NE-trending faults, and its magnetic signature suggests that it is a shallow (<1 km) complex of narrow dykes that fed the tholeiites. This suggests that continental tholeiites were emplaced in a local sinistral transtensional basin within the arc along the northern Gondwanan margin. Deformation of the basin-fill volcanic and sedimentary rocks was accompanied by a switch to dextral shear during basin inversion.

Regional studies indicate that the kinematic reversal and the cessation of magmatism is a local expression of the Neoproterozoic–Early Paleozoic termination of subduction and generation of a transform fault along the northern Gondwanan margin, possibly by ridge–trench collision, analogous to the Cenozoic evolution of the San Andreas Fault system in the western United States.

Keywords: Neoproterozoic; Magmatism; Tectonics; Avalonia; Gondwana; Structure

1. Introduction

A number of recent studies on modern arcs have indicated the importance of extensional regimes in the generation of arc magmas (e.g. Hyndman et al., 2005) and that the composition and ascent of these magmas are profoundly influenced by faults and fractures whose orientation reflects plate boundary conditions (Tibaldi, 1992; Spinks et al., 2005). In the late Neoproterozoic, subduction beneath the northern margin of West Gondwana (Amazonia and West Africa cratons) produced voluminous arc-related magmatism, much of

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which occurred in extensional settings (O'Brien et al., 1983, 1996; Keppie, 1985; Murphy and Nance, 1989; Quesada, 1990; Strachan and Taylor, 1990; Strachan et al., 1996; Linnemann and Romer, 2002). If similar relationships between magma type and structural setting can be identified in these Neoproterozoic arcs, they may yield constraints on plate boundary conditions along the northern Gondwanan margin.

Many of the regions affected by this subduction effect have since been rifted from the northern Gondwanan margin, and were incorporated as terranes in Paleozoic orogenic belts of Laurentia and Baltica. However, their origin along the northern Gondwanan margin is confirmed by abundant paleomagnetic and faunal data which also show that the terranes (collectively termed "peri-Gondwanan") formed at considerable latitudinal distance from Laurentia and Baltica (e.g. Cowie, 1974; Theokritoff, 1979; Johnson and Van der Voo, 1986; Van der Voo, 1988; McKerrow et al., 1992; Landing, 1996; Cocks, 2000; McNamara et al., 2001; MacNiocaill et al., 2001; van Staal et al., 1998). A wealth of varied geologic data strongly identify Neoproterozoic and earliest Paleozoic tectonothermal linkages and possible former continuity between peri-Gondwanan terranes in Atlantic Canada and the eastern United States (West Avalonia, Carolinia), Europe (e.g. East Avalonia, Cadomia, Iberia, Bohemia, Carpathians) and Central America (e.g. Oaxaquia, Yucatan, Chortis).

Recent overviews of the Neoproterozoic tectonothermal evolution of these peri-Gondwanan terranes may be obtained in Hibbard et al. (2002), Nance et al. (2002), Keppie et al. (2003) and Murphy et al. (2004). Together, they record a protracted (ca. 780–600 Ma) history of subduction, followed at ca. 600–550 Ma, by a diachronous transition to a continental transform fault environment, which terminated orogenic activity, in a manner analogous to that of the San Andreas fault system of the western United States (Murphy et al., 1999, 2000; Nance et al., 2002; Keppie et al., 2003). This record implies that the peri-Gondwanan terranes faced an open ocean throughout the Neoproterozoic and Early Paleozoic, such that their evolution provides important constraints on continental reconstructions for that time interval. This is a pivotal interval in the Earth's history during which a succession of events, including widespread orogeny, profound changes in ocean geochemistry, and an explosion in biological activity, led to irreversible global change (e.g. Dalziel, 1997; Knoll, 1992; Hoffman et al., 1998).

Despite broad similarities, each peri-Gondwanan terrane has characteristics that require a more precise documentation of the relationship between magmatism and deformation along the Gondwanan margin. This paper focuses on this relationship in the late Neoproterozoic rocks of the Antigonish Highlands, Nova Scotia, Canada which is a part of West Avalonia, and shows that the evolution of this region involves an intimate interplay of basin development and inversion, shear zone-related deformation, and arc-related magmatism. This style of interplay may be a local expression of a more general relationship within Avalonia and related terranes along the northern Gondwanan margin and provides an example of how plate boundary conditions along that margin can be deduced.

2. Geology of the Antigonish Highlands

The Antigonish Highlands of Nova Scotia lie within the Avalon terrane (or West Avalonia) of the northern Appalachians, which presently occupies much of the eastern seaboard of North America between Newfoundland and New England (Fig. 1A, Williams, 1979). Although there is general agreement that West Avalonia lay along the northern Gondwanan margin in the Neoproterozoic its precise location (adjacent to Amazonia or West Africa) is controversial (see McNamara et al., 2001; MacNiocaill et al., 2001; Murphy et al., 2002). Resolution of this controversy does not affect what follows, and West Avalonia is shown adjacent to Amazonia in Fig. 1B.

The Antigonish Highlands (Fig. 2) are bounded to the northwest by the Hollow Fault and to the south by the Chedabucto Fault and are predominantly underlain by Neoproterozoic (ca. 620–605 Ma) volcanic, sedimentary and plutonic rocks that are a local manifestation of the regional arc-related tectonothermal activity that typifies West Avalonia. Older basement is not exposed. Neoproterozoic stratified rocks are assigned to the Georgeville Group which consists of a ca. 4000 m thick succession of volcanic rocks overlain by a thick sequence of turbidites and minor volcanic rocks (Fig. 3; Murphy and Keppie, 1987). Plutonic rocks range from mafic to felsic in composition, and include the Greendale Complex which is spectacularly exposed along the coastal section in the northernmost highlands.

The lowermost formations in the Georgeville Group (Chisholm Brook and Keppoch Formations) occur in the northern and southern highlands, respectively (Fig. 3). In the northern highlands, the Chisholm Brook Formation is dominated by interbedded calc-alkalic mafic to intermediate flows, minor Fe–Ti rich, tholeiitic mafic flows and interbedded carbonate sedimentary rocks which were deposited in a shallow marine environment (Murphy and Keppie, 1987; Murphy et al., 1990). In the southern highlands, the Keppoch Formation is dominated by

Fig. 1. (A) Location of West Avalonia and related peri-Gondwanan terranes in an Early Mesozoic (Pangean) reconstruction. AH = Antigonish Highlands. (B) Neoproterozoic location of West Avalonia along the Amazonian margin of northern Gondwana (see Nance et al., 2002; Keppie et al., 2003; Murphy et al., 2004). Ch: Chortis, Ox: Oaxaquia, Y: Yucatan, F: Florida.

three distinct types of interbedded volcanic rocks including calc-alkalic basaltic andesites and andesites, Fe–Ti rich basalts and felsic rhyodacites and rhyolites. Minor interbedded red arkosic sandstones and conglomerates indicate deposition in a subaerial environment. The calc-alkalic rocks are chemically very similar to those of the Chisholm Brook Formation and both suites are thought to be derived from the mantle wedge above a subduction zone (Murphy et al., 1990). The Fe–Ti rich basalts are continental tholeiites and are also compositionally similar to the tholeiitic mafic flows in the Chisholm Brook Formation and are thought to be derived from the subcontinental lithosphere. The felsic rocks have chemical and isotopic signatures typical of crustally-derived magmas (Murphy et al., 1990). U–Pb (zircon) data from a Keppoch Formation rhyolite yielded an age of 618 + 2 Ma (Murphy et al., 1997b) that is interpreted as the depositional age of the formation.

The highest formations occur in the central highlands, and were deposited in a submarine environment, suggesting that the overall Georgeville Group stratigraphy records the development of a sedimentary basin (Murphy

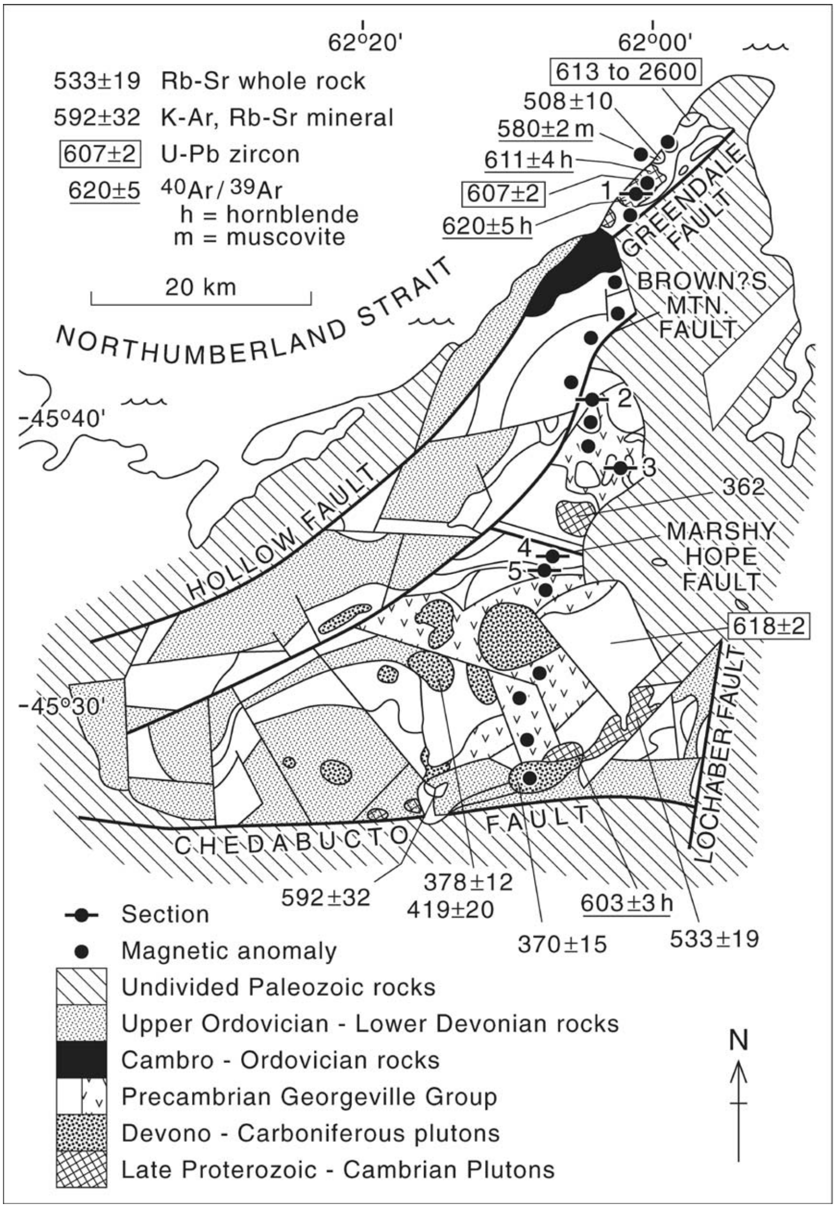

Fig. 2. Summary geological map of the Antigonish Highlands (summarized from Murphy et al., 1991). The axis of the total field magnetic field anomaly is from Geological Survey of Canada (1982). Location of magnetic profiles (see Fig. 4) are: (1) Greendale; (2) Browns Mountain, (3) Deadmans Lake, (4) Beaver Mountain A; (5) Beaver Mountain.

and Keppie, 1987). Correlations across the basin suggest that the formations are broadly coeval, and represent facies variations from the flanks to the centre of the basin (Murphy et al., 1991). The overlying sequence of turbidites range includes greywacke, mudstone and minor matrix-supported conglomerate and is thought to represent turbidite deposition in submarine channels and fans within a rifted arc basin setting (Murphy and Keppie, 1987). Geochemical and isotopic data (Murphy and McDonald, 1993) indicate that these sedimentary rocks have compositions consistent with mixing between the three volcanic rocks types, implying that they were largely derived from the cannibalization of the arc, with only a minor component of extra-basinal detritus. U–Pb (Thermal Ionization Mass Spectrometry, TIMS) single grain detrital zircon analyses taken from a turbidite sample yielded a range of ages, the youngest of which is 613 ± 5 Ma (Keppie et al., 1998), which provides a maximum age for the deposition of these rocks. Minor volcanic rocks interbedded with the turbidites (Clydes-

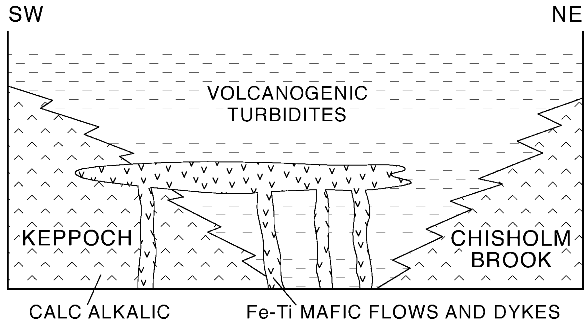

Fig. 3. Schematic diagram showing the distribution of the lithologic units and the facies relationships in the Georgeville Group. The Fe–Ti rich mafic volcanic rocks are associated with feeder dykes.

dale Formation) are Fe–Ti tholeiites (Figs. 3 and 4), very similar to the Fe–Ti tholeiites in the Keppoch Formation (Murphy et al., 1990).

The Georgeville Group is polydeformed by three phases of folding. F_1 isoclinal, possibly recumbent folds with variably developed cleavage that was accompanied by lower greenschist facies (chlorite-grade) metamorphism. This was followed by N–S F_2 and E–W F_3 upright, open to close folds (Murphy et al., 1991). F_1–F_3 fold superposition produce type 2 (mushroom) interference patterns with north–south F_1 closures, indicating that the trend of the F_1 hinge line was N–S prior to F_3. Since F_2 produced shallowly plunging N–S upright folds, this was likely the trend of F_1 prior to F_2.

The Greendale Complex is the most voluminous of several Neoproterozoic (ca. 610–605 Ma) mafic to felsic plutonic bodies. Exposed in the northernmost highlands between the NE-trending Hollow and Greendale faults

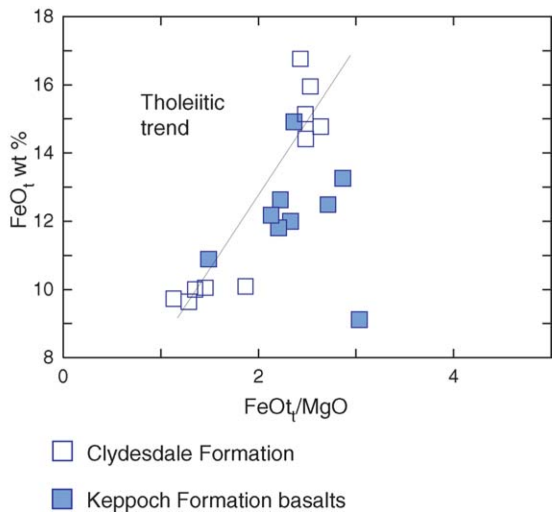

Fig. 4. Comparative geochemistry of basaltic rocks of the mafic rocks of the Keppoch and Clydesdale Formations emphasizing their FeO^t-rich compositions.

(Hollow–Greendale Fault system) (Fig. 2), the complex predominantly is semi-circular in plan and predominantly consists of small stocks and dykes of ultramafic, mafic and felsic compositions. Although the predominant lithology is fine to coarse grained hornblende gabbro to diorite, there are also abundant hornblende pegmatites, lamprophyres and fine grained felsic dykes and pods (Murphy et al., 1997a). U–Pb (titanite) data from a hornblende pegmatite yield an age of 607 + 2 Ma, and is interpreted as the age of emplacement of the complex (Murphy et al., 1997b).

3. Structural control on the basin evolution

The age of deformation in the Neoproterozoic rocks is constrained by the contact relationships between the Georgeville Group and the Greendale Complex. The deformation throughout most of the Antigonish Highlands produced a relatively weak and low-grade D_1 regional cleavage. Between the Hollow and Greendale Faults, however, this regional fabric is locally overprinted by an intense mylonitic foliation associated with narrow shear zones (generally <50 cm wide) with dextral kinematic indicators developed in Georgeville Group host rocks adjacent to the contacts with the Greendale Complex (Murphy et al., 2001). In some localities, the mylonitic foliation extends across the contact into the Greendale Complex at a high angle. In other localities, the foliation is truncated at the contact, and randomly-oriented contact metamorphic growth of hornblende in the wall-rock overprints the shear zone fabrics. These field relationships suggest that the intrusion of the Greendale Complex was syn- to late tectonic with respect to dextral shear on the NE-trending Hollow–Greendale Fault system. The emplacement of dykes and layers within the Greendale Complex is also consistent coeval shear along the bounding Hollow and Greendale faults during emplacement in which intrusion was associated with progressive and repeated extensional failure at the roof of the magma chamber (Murphy and Hynes, 1990).

These relationships together with the geochronological data imply a limited time interval (ca. 5 Ma) between deposition of the Georgeville Group, regional deformation, dextral shear along the Hollow–Greendale Fault system and emplacement of the Greendale Complex (Fig. 5). In this context, the N–S orientation of coeval regional F_1 and F_2 folds within the Georgeville Group is also consistent with an origin by dextral shear along the NE-trending faults including the Hollow–Greendale fault system (Fig. 5B; Murphy et al., 1991).

As the NE-trending fault system was active in the Neoproterozoic, it may have played a role during depo-

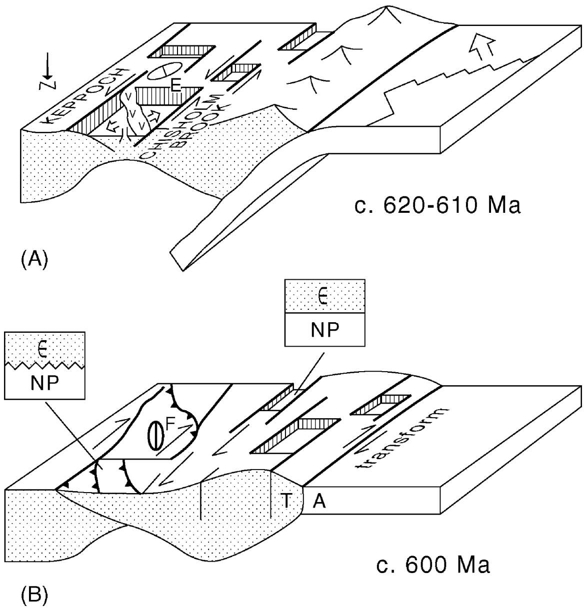

Fig. 5. Schematic model showing (A) how the turbidites of the Georgeville Group were deposited in an arc rift developed during oblique subduction and sinistral shear on NE-trending faults (see Murphy et al., 1999). The Fe–Ti rich mafic volcanics are inferred to have been emplaced in a N–S orientation, giving rise to the magnetic anomaly with a N–S orientation. Note that this direction is parallel to the extensional plane (E) in the instantaneous strain ellipse generated by sinistral movement along NE–SW faults. (B) As subduction along the northern Gondwanan margin was succeeded by transform fault development, dextral shear resulted in basin inversion and N–S F_1 and F_2 folds parallel to the plane of flattening (F) in the instantaneous strain ellipse.

sition of the Georgeville Group and opening of the basin. For example, a pronounced N–S trending total field and vertical magnetic anomaly between 2 and 2.5 km wide underlies the Georgeville Group in the southern and central Antigonish Highlands (Geological Survey of Canada, 1982; Fig. 2). Field observations in the vicinity of the anomalies indicate that these localities are predominantly underlain by Fe–Ti mafic flows and coeval dykes of the Keppoch Formation (southern highlands) and Clydesdale Formation (central highlands). Although the Clydesdale Formation stratigraphically underlies the Keppoch Formation, the above age data and constraints from contact relationships suggest these formations are related as lateral facies equivalents.

4. Fe–Ti rich rocks

As magnetic anomalies in the Antigonish Highlands are spatially associated with Fe–Ti rich mafic rocks, their mineralogy and lithogeochemistry are briefly summarized (see Murphy et al., 1991, 1997a for details). Fe–Ti rich basalts predominantly occur in the Keppoch and Clydesdale formations as interlayered flows with calc-alkaline basaltic andesites and rhyolites (Keppoch Formation) and with turbidites (Clydesdale Formation). These rocks are described in detail in Murphy et al. (1991). They are generally medium to fine grained containing microphenocrysts of augite and (more rarely) orthopyroxene, plagioclase and a fine grained matrix dominated by magnetite, ilmenite and plagioclase that exhibits flow textures. The basalts have been subjected to low grade alteration, as evidenced by the plagioclase composition (albite) and by the presence of chlorite, epidote and hydrated iron oxides.

Microprobe analyses of augite and orthopyroxene indicate that most pyroxenes are iron-rich ($Mg^{\#} = 50$), consistent with a crystallization in a fractionated magma (Table 1). Microprobe analyses of magnetite consistently yield low total wt.% oxides (typically 92–95 wt.%) suggesting that are partially altered to hematite and/or hydrated iron oxides. Ilmenite, on the other hand, has high total wt.% oxides, and with Fe/Ti cation ratios close to 1 (Table 1).

The high FeO^t (e.g. Fig. 4) and TiO_2 (2.0–4.0 wt.%) and moderate light rare earth (LREE) enrichment (Fig. 6A and D) exhibited by the mafic rocks are attributed to fractionation of a tholeiitic magma (Murphy et al., 1990). Trace element compositions bear a closer resemblance to E-MORB than to N-MORB (Fig. 6B, C, E and F) although the negative Nb and Ti anomalies suggest some contamination either by continental crust or a subduction component. Either source of contamination is possible in the rifted ensialic arc environment.

5. Greendale Complex

The lithologies and geochemistry of the Greendale Complex are described in detail elsewhere (Murphy et al., 1991, 1997a). The complex predominantly consists of hornblende-plagioclase mafic to intermediate rocks and subordinate felsic rocks that collectively exhibit calc-alkaline trends and have highly variable textures typical of appinite suites (Murphy et al., 1997a). The Greendale Complex also exposes an important ultramafic component (about 10%) as discontinuous sheets and pods (which are thought to have been layers formed early in the cooling history of the complex that were subsequently tectonically dismembered, Murphy and Hynes, 1990). They contain 10.0–12.5 wt.% FeO^t, and as their field occurrence matches the location of the N–S magnetic anomaly in the northernmost highlands, they are herein interpreted to be the source of that anomaly. Ultramafic rocks are medium to

Table 1
Representative mineral analyses (wt.% oxides and stoichiometry) of silicate minerals (orthopyroxene and clinopyroxene) and oxides (ilmenite and magnetite) from mafic flows and coeval dykes within the Georgeville Group

Sample MS	27-1:1a	27-3:1a	27-20:2a	27-20:2b	27-2:1a
Orthopyroxene					
SiO_2	51.55	52.63	52.76	52.58	52.77
TiO_2	0.46	0.34	0.28	0.29	0.28
Al_2O_3	2.84	1.75	1.51	1.62	2.04
Cr_2O_3	0.12	0.15	0.15	0.17	0.06
FeO	25.12	25.61	26.38	25.99	25.69
MnO	0.69	0.89	0.79	0.82	0.73
MgO	14.56	14.81	15.00	15.00	14.96
CaO	2.68	1.80	1.25	1.86	1.59
Na_2O	0.35	0.17	0.18	0.16	0.23
K_2O	0.00	0.06	0.02	0.07	0.00
Total	98.36	98.20	98.31	98.56	98.36
Si	1.99	2.03	2.04	2.03	2.03
Ti	0.01	0.01	0.01	0.01	0.01
Al	0.13	0.08	0.07	0.07	0.09
Cr	0.00	0.00	0.00	0.01	0.00
Fe	0.81	0.83	0.85	0.84	0.83
Mn	0.02	0.03	0.03	0.03	0.02
Mg	0.84	0.85	0.86	0.86	0.86
Ca	0.11	0.07	0.05	0.08	0.07
Na	0.03	0.01	0.01	0.01	0.02
K	0.00	0.00	0.00	0.00	0.00

Sample MS	18-3:1a	18-1:1a	18-5:4a	14-7:3a	13-2:3a
Clinopyroxene					
SiO_2	52.64	52.99	52.87	51.17	52.62
TiO_2	0.25	0.21	0.27	0.99	0.06
Al_2O_3	0.88	1.18	1.67	3.16	0.29
Cr_2O_3	0.12	0.13	0.10	0.22	0.10
FeO	8.01	7.99	8.64	8.97	11.84
MnO	0.70	0.62	0.66	0.25	0.44
MgO	14.70	14.93	14.04	13.63	10.81
CaO	22.52	22.20	22.20	21.58	24.91
Na_2O	0.27	0.36	0.31	0.61	0.17
K_2O	0.03	0.00	0.00	0.05	0.01
Total	100.13	100.60	100.77	100.61	101.25
Si	1.96	1.96	1.96	1.90	1.98
Ti	0.01	0.01	0.01	0.03	0.00
Al	0.04	0.05	0.07	0.14	0.01
Cr	0.00	0.00	0.00	0.01	0.00
Fe	0.25	0.25	0.27	0.28	0.37
Mn	0.02	0.02	0.02	0.01	0.01
Mg	0.82	0.82	0.78	0.76	0.61
Ca	0.90	0.88	0.88	0.86	1.01
Na	0.02	0.03	0.02	0.04	0.01
K	0.00	0.00	0.00	0.00	0.00

Table 1 (*Continued*)

Sample MS	27-5:4a	14-1:2b	14-7:1a	13-5:3a	2-3:4a
Ilmenite					
SiO_2	0.00	0.02	0.01	0.00	0.00
TiO_2	51.67	50.39	51.47	48.87	50.42
Al_2O_3	0.00	0.02	0.07	0.02	0.06
Cr_2O_3	0.36	0.28	0.28	0.31	0.29
FeO	46.10	47.11	46.08	43.68	43.07
MnO	2.51	2.60	2.30	7.16	5.32
MgO	0.05	0.08	0.10	0.00	0.05
CaO	0.14	0.04	0.05	0.13	0.15
Na_20	0.01	0.04	0.02	0.02	0.01
K_20	0.05	0.00	0.00	0.06	0.00
Total	100.89	100.58	100.38	100.23	99.36
Si	0.00	0.00	0.00	0.00	0.00
Ti	0.98	0.96	0.98	0.95	0.97
Al	0.00	0.00	0.00	0.00	0.00
Cr	0.01	0.01	0.01	0.01	0.01
Fe	0.97	1.00	0.98	0.94	0.92
Mn	0.05	0.06	0.05	0.16	0.12
Mg	0.00	0.00	0.00	0.00	0.00
Ca	0.00	0.00	0.00	0.00	0.00
Na	0.00	0.00	0.00	0.00	0.00
K	0.00	0.00	0.00	0.00	0.00

Sample MS	18-5:3a	14-3:2a	13-4:4a	11-9:1a	2-3:5a
Magnetite					
SiO_2	0.00	0.00	0.07	0.00	0.00
TiO_2	0.18	2.02	1.23	1.07	14.83
Al_2O_3	0.25	1.25	0.65	2.04	0.74
Cr_2O_3	0.45	0.56	1.07	5.59	0.81
FeO	93.02	89.91	91.51	86.12	75.62
MnO	0.12	0.21	0.28	0.37	2.12
MgO	0.00	0.04	0.02	0.08	0.03
CaO	0.09	0.08	0.13	0.10	0.24
Na_2O	0.04	0.01	0.03	0.00	0.04
K_2O	0.00	0.06	0.00	0.07	0.00
Total	94.16	94.16	95.00	95.44	94.41
Si	0.00	0.00	0.00	0.00	0.00
Ti	0.00	0.02	0.01	0.01	0.12
Al	0.00	0.02	0.01	0.03	0.01
Cr	0.00	0.01	0.01	0.05	0.01
Fe	0.98	0.92	0.94	0.85	0.70
Mn	0.00	0.00	0.00	0.00	0.02
Mg	0.00	0.00	0.00	0.00	0.00
Ca	0.00	0.00	0.00	0.00	0.00
Na	0.00	0.00	0.00	0.00	0.00
K	0.00	0.00	0.00	0.00	0.00

Orthopyroxene and clinopyroxene are normalized to six oxygens, ilmenite is normalized to three oxygens, and magnetite (because of alteration) is normalized to one oxygen.

coarse grained, and are dominated by brown amphibole oikocrysts (magnesiohastingsite, classification of Leake et al., 2003) up to 12 mm in length. Olivine ($Fo_{80–78}$), augite ($Mg^{\#} = 86–68$), and bronzite ($Mg^{\#} = 73–71$) occur as poikilitic inclusions within amphibole that range in diameter from 1 to 4 mm and are enclosed by oikocrystic amphibole giving the rock an overall poikilitic texture. Interstitial plagioclase ($An_{65–79}$), phlogopite and minor accessory phases, including Fe-rich chromite and par-

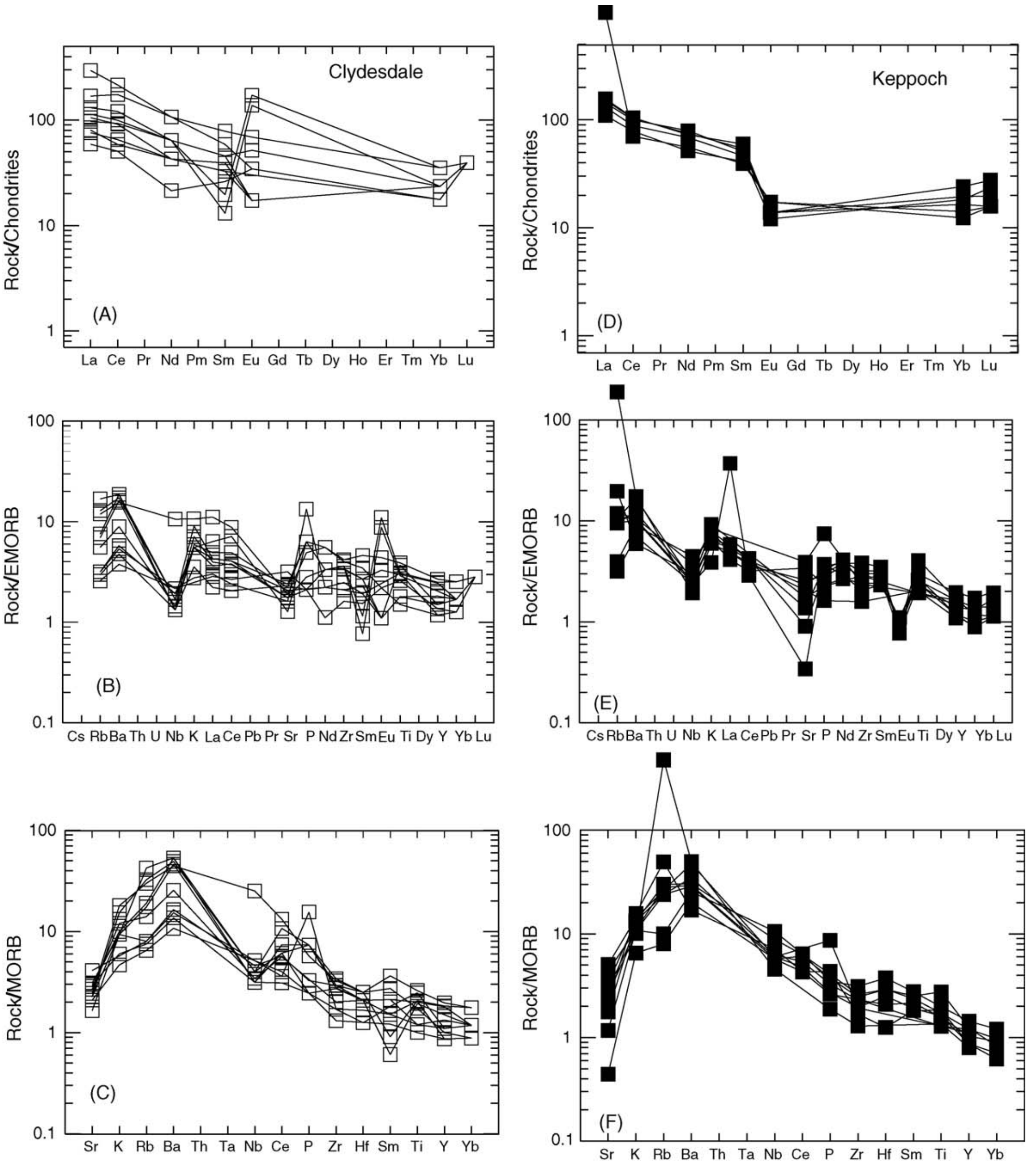

Fig. 6. Spidergrams showing trace and REE element compositions of the mafic tholeiitic rocks of: (A–C) Clydesdale Formation and (D–F) Keppoch Formation (normalizing values after Sun and McDonough, 1989; Hofmann, 1988).

tially altered magnetite account for less than 5%. On multi-element normalized plots, the ultramafic rocks exhibit low total REE and magmatophile elements such as Zr, reflecting their cumulate composition. Greendale Complex rocks exhibit slight to moderate enrichment of LREE relative to HREE. Taken together, the content and composition and lithogeochemical signature is consistent with derivation from a tholeiitic magma. The lack of Nb or Ti anomalies, however, suggests limited contribution from either continental crust or a subducted component.

6. Aeromagnetic data

The N–S magnetic anomaly identified on total field and vertical gradient magnetic maps (Geological Survey of Canada, 1982) is one of the most striking geophysical features in the Antigonish Highlands. These maps were derived from and aeromagnetic survey of the Antigonish Highlands and the surrounding areas conducted during 1977 and 1978. The survey was flown at 300 m above the ground at 300 m line spacings with a gradiometer mounted on a Queenair aircraft, by the Geological Survey of Canada as part of the DREE 1974-9 program (Hood et al., 1979; Geological Survey of Canada, 1982). This technique used two high sensitivity magnetometers mounted 3 m vertically apart and produced total field and vertical gradient magnetic maps (Geological Survey of Canada, 1982), which were published as magnetic contours on a topographic base and also with the magnetic data coloured and superimposed upon a topographic base with geological boundaries marked. In general, the total magnetic field values in Georgeville Group Neoproterozoic rocks vary from 55,000 to 55,500 γ (1 γ = 1 nT) in the less magnetic rhyolites and turbidites to ca. 55,500–57,000 γ in the mafic flows and related dykes.

Field investigations showed that the location of the N–S magnetic anomaly predominantly corresponded with exposures of basalt and associated dykes. In order to constrain the origin of the anomaly, magnetic susceptibility measurements were made on mafic samples selected from these formations using a hand-held susceptibility meter (Model KT-9 Kappameter manufactured by Exploranium G.S. Limited, Mississauga, Ontario). The magnetic susceptibility of a rock is roughly equivalent to the weighted average of the susceptibility of individual minerals times the percentage of each mineral in the rock. The KT-9 has a sensitivity of 1×10^{-5} S.I. units, which enables it to measure small variations in the contents of magnetic minerals (such as magnetite, titanomagnetite, ilmenite and pyrrhotite). The coil gets 90% of its signal from the first 20 mm of the sample, so that local heterogeneities in mineral content should not distort the measured values. Each sample was measured four times at approximately orthogonal orientations. Although a few samples displayed minor intra-sample variation (typically less than 20% from the mean), and this anisotropy in magnetic susceptibility was not related to any recognizable fabrics in the rocks, suggesting that any anisotropy would not significantly affect regional magnetic trends. An average of the four measurements was taken as representative of each sample (Table 2).

Using these values, it is clear that the magnetic susceptibility varies significantly from sample to sample, from a highest value (MS-14) of 50.1×10^{-3} S.I. to the lowest value of MS-1 (0.1×10^{-3} S.I.). The magnetic susceptibility values are strongly influenced by the amounts of Fe-bearing minerals identified in these rocks, especially magnetite and ilmenite and by visible low grade alteration of those minerals. However magnetic susceptibility values display no obvious correlation with abundances of elements e.g. (FeO^t, TiO_2, V, Nb) that are concentrated in these minerals (Tables 2 and 3), sug-

Table 2
Magnetic susceptibility measurements on mafic samples from outcrops in the vicinity of the magnetic anomaly using a hand-held susceptibility meter

Sample	Average MS
MS 1	0.1
MS 2	14.8
MS 3	0.1
MS 4	2.2
MS 5	4.8
MS 6	2.1
MS 7	3.7
MS 8	0.9
MS 9	8.2
MS 10A	6.9
MS 10B	3.2
MS 11	22.1
MS 12	29.5
MS 13	27.1
MS 14	49.7
MS 15	20.2
MS 16	11.9
MS 17	9.5
MS 18	38.4
MS 19	17.4
MS 20	0.1
MS 21	0.4
MS 22	2.4
MS 23	0.1
MS 24	0.2
MS 25	0.2
MS 26	14.5
MS 27	0.2
MS 28	0.3
MS 29A	0.2
MS 29B	1.6
MS 30	0.2
MS 31	0.3
MS 32	0.2
MS 33	0.3
MS 34	0.1
MS 35	0.1
MS 36	0.1
MS 37	4.4

Model KT-9 Kappameter manufactured by Exploranium G.S. Limited, Mississauga, Ontario. The magnetic susceptibility of a rock is roughly equivalent to the weighted average of the susceptibility of individual minerals times the percentage of each mineral in the rock. The susceptibility meter (Model KT-9 Kappameter) has a sensitivity of 1×10^{-5} S.I. units. Each sample was measured four times at approximately orthogonal orientations. Minor intra-sample variation was identified (typically less than 20% from the mean). The values shown are the average of the four measurements taken on each sample.

Table 3
Depth and width estimates (m) of the magnetic body along various profiles from north (Greendale) to south

Profile	Half-Width	Peters Rule	Thalen Rule	Mag susceptibility	
				Depth	Width
Greendale	100	960	650	228–375	104
Browns Mtn	75	690	770	172–288	75–63
Deadmans Lake	45	425	750	267–414	48–42
Beaver Mtn A	12	1310	1500	395–530	111–287
Beaver Mtn C	950	815	770	576–952	139–150

All estimates in metres. The first three columns give depth estimates to the top of the magnetic body by analyzing different aspects of the curve of the magnetic profile. The fourth column provides estimates of the shape of the body by inverting the data using an assumed magnetic susceptibility value. The width of the body is given from top to bottom. The estimates in column 4 were derived using a typical value for the magnetic susceptibility of basalt (2×10^{-3} S.I., see text).

gesting that the extent of alteration has an overriding influence on measured magnetic susceptibility values. This is consistent with petrographic and microprobe evidence that the original mineralogy has been variably affected by low grade alteration. It is uncertain to what extent the alteration visible in surface samples is representative of the alteration at depth, and so a typical value for basalt (2×10^{-3}) was used to model the anomaly.

The dimensions of the anomaly suggest the presence of a north–south dyke system that acted as a feeder to the mafic volcanics. Five E–W profiles up to 2.5 km in length were constructed across various segments across this anomaly (Fig. 7). Depth estimates to the anomaly may be estimated to within 20% by several empirical methods that analyze the shape of the profile (such as by using horizontal width at some fraction of the peak anomaly, Milsom, 1989) or by using magnetic susceptibility measurements to invert the profile. Although these methods assume that the anomaly is due to a single sheet, rather than a series of dykes, the estimates are considered to give a generalized indication of depth to the intrusive mafic bodies.

According to (i) the Half-Width Rule, the depth D to the magnetic anomaly = V_2 width between the flanks of the anomaly, (ii) the Peters Rule $D = 0.625 \times$ horizontal distance between the points of tangency of the lines whose slope is half of the slope at the point of inflection and (iii) to the Thalen Rule, the magnetic source depth is $0.7B$ where B is the horizontal distance from the edge of the anomaly to the peak (Beck, 1991; Milsom, 1989). However, the near symmetrical magnetic profiles in each locality suggest that the feeder system is relatively steep. With the exception of the Deadmans Lake profile, depth estimates from these methods (Table 3) are reasonably consistent, and collectively suggest that the buried mafic body or bodies lie at depths between ca. 450–1500 m. There is no apparent trend with depth along the length of the anomaly. Differences in depth estimates between the various methods suggest that the anomaly may be a composite signature of thin dykes separated by screens of less magnetic host rock, rather than a single feeder dyke.

Although the basalts exhibit a considerable range in magnetic susceptibility (Table 3), in modeling of the anomalies in each profile (Fig. 7), a basalt magnetic susceptibility value of 2×10^{-3} S.I. was used. The value selected affects the depth estimates shown for the mafic bodies, but not the main conclusion of the model, which illustrate that the anomalies can be explained by relatively narrow, near-surface N–S trending mafic intrusions. This method yields depth estimates to the top of the mafic body ranging from 170 to 600 m. These estimates are considerably shallower than those of the other methods, possibly reflecting a more complicated geometry than the single dyke model used, and/or that the magnetic susceptibility value used is lower than the average. With the exception of Beaver Mountain C, the depth of the anomaly seems to be between 110 and 160 m and width estimates range from 42 to 287 m.

The variations in depth and dimension estimates suggest that the geometry of the magma plumbing system is more complicated than can be expressed by the various methods used to analyze the signature. Nevertheless, more generally, it is clear that the anomalies represent a concentration of shallow (<1 km depth), narrow (<300 m width) mafic bodies that lie beneath Fe–Ti rich mafic flows and co-genetic dykes exposed within the Georgeville Group stratigraphy.

Assuming these intrusions are feeders to the mafic flows and dykes, they are probably related to local extension that accompanied the opening of the basin into which the Georgeville Group strata were deposited. The N–S orientation of this anomaly is compatible with

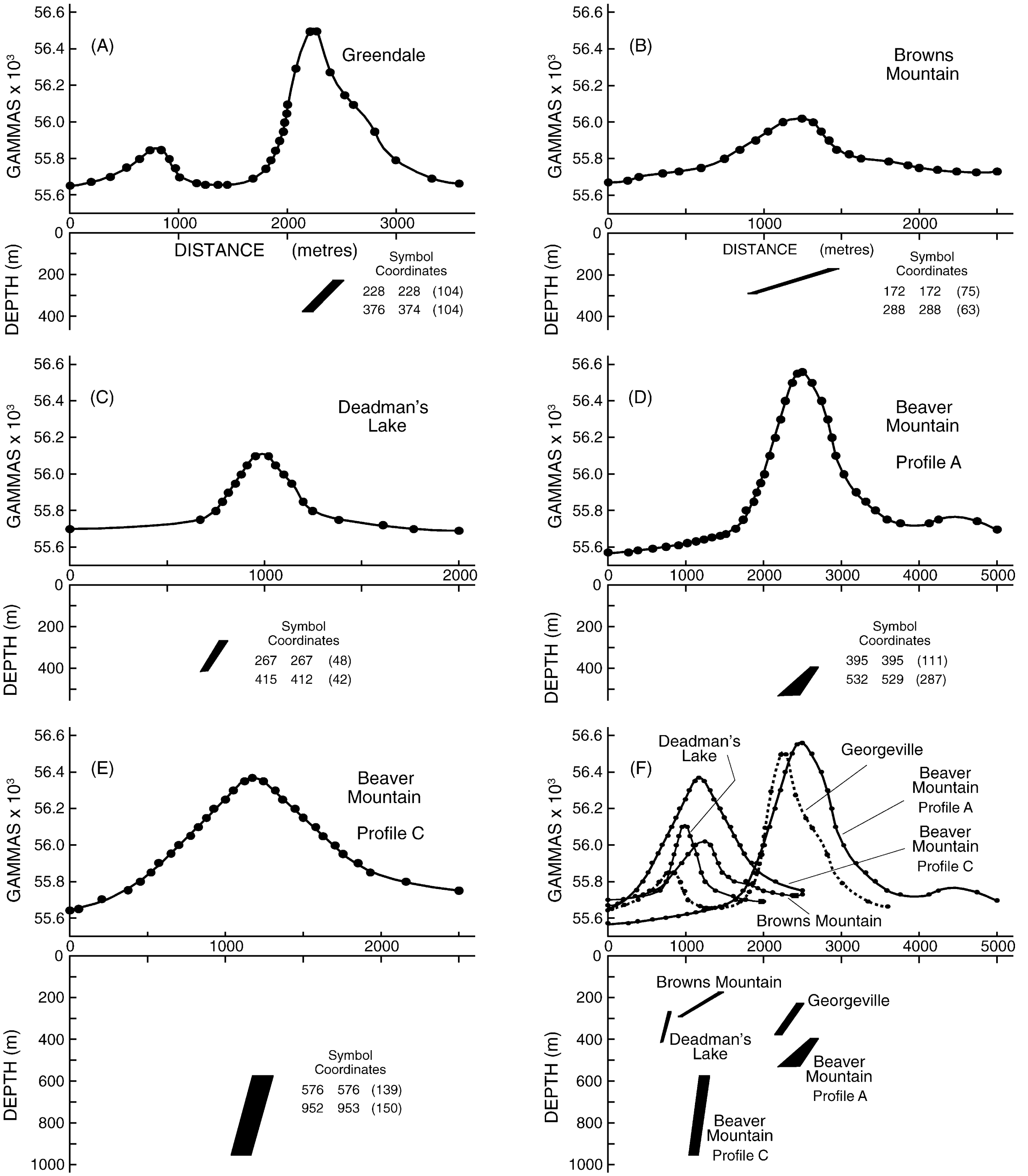

Fig. 7. Modeling of N–S geophysical anomalies along five profiles (see Fig. 2 for locations, and text for details). Note differences in the vertical and horizontal scale mean that the true dip is significantly steeper than the inclination apparent on the profile models. In each profile, the location of the four edges of the anomaly is given in metres. (A) Greendale (Profile 1, Fig. 2); (B) Browns Mountain (Profile 2); (C) Deadmans Lake (Profile 3); (D) Beaver Mountain A (Profile 4); (E) Beaver Mountain B (Profile 5). (F) is a composite of profiles and location of their respective anomaly. For each profile, the "*y*-axis" value gives the depth, and the "*x*-axis" value gives the distance from background values to the edges of the anomaly.

coeval sinistral shear along NE-trending faults during basin formation, in which the dyke orientation is parallel to the instantaneous plane of extension. Placed in a regional context, this suggests that continental tholeiites were emplaced in a local sinistral transtensional environment during the development of a basin within the Avalonian arc along the northern Gondwana margin.

The subsequent deformation, however, that produced N–S F_1 recumbent and F_2 upright folds, suggests that inversion was probably accompanied by a switch in the sense of shear from sinistral to dextral kinematics, in which the N–S orientation of the folds represented the instantaneous plane of flattening. A similar switch in the sense of shear has been documented in coeval Avalonian rocks in the adjacent Cobequid Highlands in mainland Nova Scotia using kinematic data measured along Neoproterozoic shear zones (Nance and Murphy, 1990).

7. Summary and discussion

Several recent studies have emphasized the control of strike-slip structures in the rise of magmas in the extensional regimes in modern volcanic arcs such as Andes (De Silva, 1989), NE Japan (Sato, 1994), and Mexico (Tibaldi, 1992). In the Taupo Volcanic Zone of northern New Zealand, Spinks et al. (2005) show that there is a correlation between the amount of extension and the composition of volcanic rocks produced. Highly active calderas of predominantly felsic magmatism form in regions of high orthogonal extension, whereas, andesitic magmas in regions of greatest transtension.

The extensional regime within the Neoproterozoic Georgeville arc shows a similar relationship between magma type and structural setting. The volcanic geochemistry and stratigraphy of the Georgeville Group is consistent with their formation in a rift basin within a volcanic arc (Murphy et al., 1990). Although the Antigonish Highlands is dominated by arc-related calc-alkalic rocks, tholeiitic mafic flows and related hypabyssal intrusions which occur in the Keppoch and Clydesdale formations may be related to extension within the arc. Their field occurrence is spatially associated with strong positive total field and vertical gradient N–S trending magnetic anomalies (Geological Survey of Canada, 1982) that are represented in outcrop by Fe–Ti rich tholeiitic basalts and ultramafic cumulate rocks of the Greendale Complex and can be explained by the presence of shallow intrusions of mafic composition.

These basaltic rocks are characterized by strong enrichment in FeO, TiO_2, and higher in incompatible elements such as Zr, Y and LREE. Their compositions are typical of Fe-enriched tholeiitic suites that underwent fractionation of olivine, clinopyroxene and plagioclase (Murphy et al., 1990). These magmas are thought to have been generated in an transtensional regime within the arc, with deep-seated faults providing conduits for magmas from various depths.

The limited time gap between the deposition and regional deformation of the Georgeville Group, shearing and the intrusion of the plutonic complexes is consistent with a strike-slip setting within the Neoproterozoic volcanic arc (Murphy et al., 1991). The opening of the basin is related to sinistral motion (during which the continental tholeiites were emplaced) and dextral motion has been related to basin inversion within the arc (Murphy et al., 1997b). During basin opening, sinistral motion of NE-trending faults produced N–S oriented extensional fractures that were exploited by Fe–Ti rich tholeiites, producing a N–S oriented magnetic anomaly.

Although there is evidence in many areas of Avalonia and related terranes that tectonic evolution and magmatism are inter-related and inter-dependent (e.g. Murphy and Hynes, 1990; Strachan et al., 1996; Gibbons and Horák, 1996), precise documentation on the influence of tectonic processes on the generation of magma, as well as its ascent towards the surface is less well understood. The Hollow–Greendale Fault zone may provide an example of the influence of the kinematics of fault systems and the development of conduits and ascent of magma along the Neoproterozoic northern margin of Gondwana.

Q&A section

Question by P. Clift: Can you think of any modern analogues to your proposed arc basin and do these modern systems have the chemical diversity that you have recorded in the Antigonish highlands?

Murphy answers:

One of the key problems understanding the link between tectonics and magmatism in these environments is, that unlike ancient arcs, the subsurface plumbing system and its relationship to regional structures are rarely exposed. Although the details are controversial, the general relationship between the structural evolution of modern rifted island and continental arcs and the composition and style of coeval volcanism is becoming increasingly apparent. For example, Tibaldi (1992) shows that intra-arc environments generate anomalously wide belts of magmatism, which he suggests are linked kinematically to adjacent strike-slip faults that are generated by plate boundary forces. Spinks et al. (2005) show that the regions with the greatest extension in the intra-arc Taupo Volcanic Zone of New Zealand produce active calderas dominated by felsic magmatism, whereas areas with the greatest transtension produce less active andesitic stratovolcanoes. The implication is that these varying structures provide conduits to magmas ascending from different structural levels. Paterson and Tobsich (1992) and Petford et al. (2000) show that strain rate along these structures influences magma composition and eruption style.

Structures with the modern rifted arc are apparently controlled by plate boundary kinematics rather than by local stresses (Tibaldi, 1992). This general relationship is used in the Antigonish Highlands to interpret that the emplacement of the Fe–Ti rich mafic magmas along a N–S fracture may be related to sinistral movement along adjacent strike-slip faults.

Several studies emphasize the importance of intra-arc strike-slip tectonics in controlling the rise of magmas in modern arcs (e.g. central Andes, De Silva, 1989; NE Japan, Sato, 1994; Ecudaor, Tibaldi and Ferrari, 1990; Mexico, Tibaldi, 1992; Kamchatka, Baluyev and Perepelov, 1988). Magmatism in the Kamchatka Volcanic Belt, for example, is associated with regional transcurrent faults that reach a depth of 40 km and it is suggested that the depth of magma generation is controlled by these faults (Baluyev and Perepelov, 1988). Tibaldi (1992) attributes the obliquity between the Mexican Volcanic Belt and the Middle American Trench to transcurrent motion along adjacent faults. Deep penetrating faults in intra-arc environments create favourable conditions for rapid ascent of mafic magmas with little crustal contamination (e.g. Ferrari et al., 1989), like the Antigonish continental tholeiites. In the Antigonish Highlands, magma ascent is thought to have been facilitated by the generation of a N–S fracture, which was generated by sinistral movement long NE–SW major faults (such as the Hollow Fault) which in turn is linked to plate boundary forces in a similar manner to modern arcs.

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