Last Gasp of the Grenville Orogeny: Thermochronology of the Grenville Front Tectonic Zone near Killarney, Ontario

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Last Gasp of the Grenville Orogeny: Thermochronology of the Grenville Front Tectonic Zone near Killarney, Ontario[1,2]

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ABSTRACT

We present U-Pb (titanite, zircon) and $^{40}Ar/^{39}Ar$ (hornblende, mica, K-feldspar) data from a transect across the western part of the Grenville Front Tectonic Zone (GFTZ) near Killarney, Ontario. High-grade metamorphic assemblages (~1450 Ma) in this part of the GFTZ pre-date the Grenvillian orogeny and were primarily exhumed, with little or no metamorphic overprinting, by Grenvillian deformation. The titanite and zircon data form a discordant array with an upper intercept of 1454 ± 8 Ma and a lower intercept of 978 ± 13 Ma. These data are interpreted in terms of partial lead loss during a short-lived thermal event that increased in intensity from west to east across the transect. $^{40}Ar/^{39}Ar$ data from hornblende indicate cooling through ~450°C at ~993–979 Ma, multiple diffusion domain models for the interpretation of discordant K-feldspar spectra indicate cooling through ~365–340°C at 990–960 Ma, and muscovite data indicate cooling through ~320°C at ~930 Ma. Biotite data are not easily interpreted owing to the effects of partial resetting and/or excess ^{40}Ar. The thermochronological data suggest that a thermal event with peak temperatures of 500–600°C affected the GFTZ at ~980 Ma, followed by very rapid cooling to ~350°C. We interpret the data in terms of a tectonic model involving rapid exhumation of GFTZ rocks (in response to erosion) in the hangingwall of a crustal-scale shear zone developed during a ~980 Ma episode of convergence.

Introduction

The Grenville Front is one of the most prominent tectonic features of the eastern Canadian Shield (Wynne-Edwards 1972; Rivers et al. 1989), extending from the coast of Labrador to the shores of Lake Huron, and beneath younger cover at least as far south as Ohio (e.g., Culotta et al. 1990). Typically marked by moderately to steeply dipping, NW-verging thrust faults and associated mylonite zones (e.g., Davidson 1986*a*, 1986*b*; Owen et al. 1988), it separates rocks deformed and metamorphosed during the ~1160–980 Ma Grenvillian orogeny (south and east of the Front) from a variety of Archean and Proterozoic rocks (north and west of the Front). In the western part of the orogen the Grenville Front Tectonic Zone (GFTZ), immediately southeast of the Grenville Front, is characterized by a sharp increase in metamorphic grade, so that in some places upper amphibolite to granulite facies gneisses are exposed within a few kilometers of the Front (e.g., Davidson and Bethune 1988; Indares and Martignole 1989). In the eastern part of the orogen a distinctive Grenville Front Tectonic Zone is not well developed, and the increase in metamorphic grade within the Parautochthonous Belt (Rivers et al. 1989) occurs over several tens of kilometers (e.g., Rivers 1983; Owen et al. 1988).

Despite two decades of investigation, the role of the Grenville Front and GFTZ in the Grenvillian orogeny remains unclear. The remarkably straight trend for over 1500 km (figure 1), prominent gravity and magnetic anomaly gradients (e.g., Rivers et al. 1989), and the position at the northwestern limit of Grenvillian deformation point to the fundamental importance of this zone in Grenvillian tectonics. However, the Grenville Front is clearly

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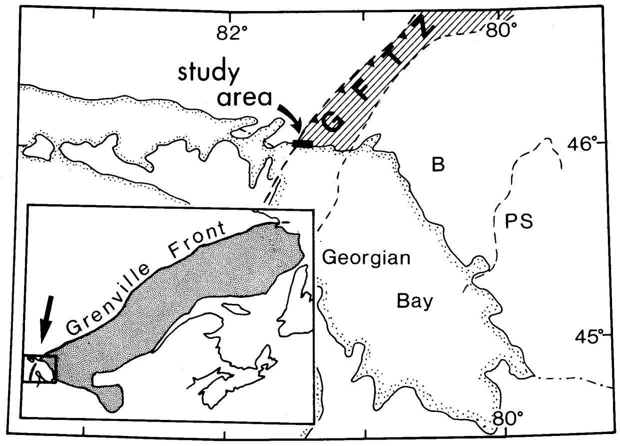

Figure 1. Location map of the study area in relation to the major tectonic elements of the southwestern Grenville Province. GFTZ = Grenville Front Tectonic Zone; B = Britt domain; PS = Parry Sound domain; barbed line = Grenville Front. Inset shows the position of the area relative to the Grenville orogen as a whole.

not a suture since distinctive rock units can be traced across it for up to ~200 km (e.g., Rivers et al. 1989). Kinematic indicators indicate NW-directed thrusting along the entire length of the Grenville Front, but the timing of this deformation is not generally well defined and may differ from place to place. U-Pb geochronologic data indicate that a significant event affected rocks of the GFTZ at ~980–1000 Ma (e.g., Krogh 1989, 1991; Childe et al. 1992), although samples from close to the Grenville Front generally show a high degree of discordance and/or inheritance, or may even lack significant Grenvillian overprinting. In some places pre-Grenvillian metamorphic assemblages are relatively unaffected for a few kilometers within the GFTZ, and in many places high-grade metamorphism can be shown to pre-date the Grenvillian orogeny. These effects can vary along strike, even in fairly small areas where the overall trend of the Grenville Front does not change significantly (e.g., Indares and Martignole 1989).

A major problem in Grenvillian geology is thus to reconcile the similarities in the orogen-scale features of the Grenville Front and GFTZ with considerable local variation in detail. What is the reason for the dramatic change in metamorphic grade over such a short distance? Were the high-grade rocks metamorphosed during the Grenvillian orogeny, or simply exhumed by Grenvillian deformation? In either case, what tectonic process is responsible for producing this effect along much of the northwestern boundary of the orogen? Understanding the timing of deformation, metamorphism, and exhumation in the GFTZ is central to answering many of these questions.

This paper reports U-Pb and $^{40}Ar/^{39}Ar$ dates from a transect across the western half of the GFTZ near Killarney, Ontario, along the north shore of Georgian Bay (figure 1; part of the Beaverstone terrane of Rivers et al. 1989). The area is well exposed and relatively accessible, with good geological and geochronological control both within the GFTZ (Davidson 1986*a*; Davidson and Bethune 1988; Bethune 1989) and to the south along Georgian Bay (e.g., Culshaw et al. 1988, 1991*a*). GLIMPCE seismic reflection profile "J" (Green et al. 1988) imaged deep crustal structure along strike ~75 km southwest of the study area. The goal of the study was to document the thermal history of the GFTZ for comparison with metamorphic and thermochronologic data from the Central Gneiss Belt and elsewhere in the GFTZ, and to constrain tectonic models for the GFTZ within this part of the Grenville orogen.

Geological Setting

The study area includes three major geological units: Huron Supergroup rocks of the Southern Province, felsic volcanic and high-level plutonic rocks of the Killarney Complex, and granitoid orthogneisses, amphibolites, and paragneisses of the Grenville Front Tectonic Zone (figure 2). The mainly clastic Huron Supergroup, deposited in the interval 2500–2200 Ma on the margin of the Archean Superior craton, is considered to have been metamorphosed at greenschist to lower amphibolite grade (Card 1978) during the Penokean orogeny (1890–1830 Ma, Van Schmus 1980). Felsic volcanic and plutonic rocks of the Killarney Complex formed at ~1740 Ma (van Breemen and Davidson 1988) and were metamorphosed under greenschist facies conditions prior to the intrusion of ~1400 Ma pegmatites (Krogh 1971; Davidson 1986*b*) and the 1238 Ma Sudbury diabase dike swarm (Krogh et al. 1987; van Breemen and Davidson 1988).

The Grenville Front, here a mylonite zone up to 50 m wide, truncates the Killarney Complex near the western end of Collins Inlet (figure 2). Metamorphic grade changes from greenschist facies in the Killarney Complex to epidote amphibolite facies in the Grenville Front mylonites. Several NE-striking mylonite zones with a northwesterly thrust sense occur along the transect within the GFTZ (Davidson and Bethune 1988). Foliations in this part of the transect are SE-dipping with a down-dip lineation, and kinematic indicators consistently show "top to the northwest." To the

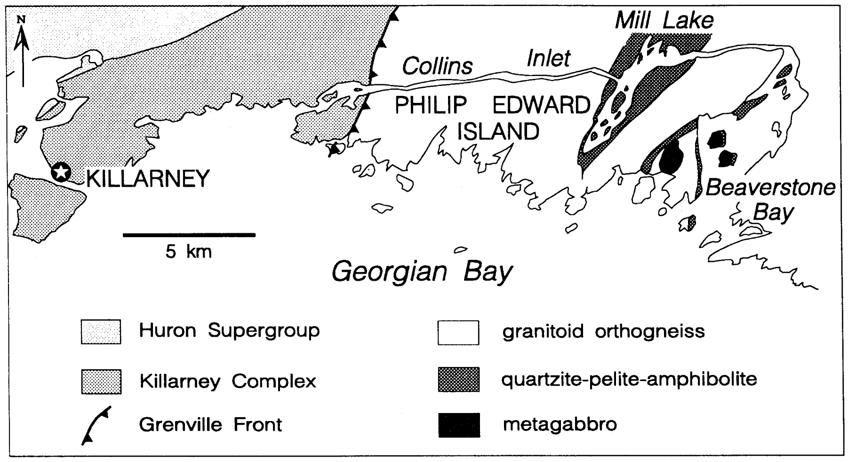

Figure 2. Simplified geological map of the Killarney area (modified after Davidson and Bethune 1988). Sampled transect runs along Collins Inlet from the Grenville Front to Beaverstone Bay (see figures 3–6 for details).

northeast the geology becomes more complex—the Grenville Front is marked by a series of thrust faults and mylonite zones, some formed at relatively low temperature, that affect a variety of supracrustal rocks. It is locally modified by younger brittle structures, including normal faults (Bethune and Davidson 1988; Davidson and Ketchum 1993).

Rocks of the GFTZ between the Grenville Front and Mill Lake (figure 2) consist mainly of variably foliated, pink, granitic orthogneisses, locally retaining evidence of a volcaniclastic protolith (Davidson and Bethune 1988; Haggart 1991). Minor, layered, hornblende-bearing, quartzo-feldspathic gneisses of uncertain parentage and attenuated amphibolitized dikes are also present. At Mill Lake and Beaverstone Bay, pelites, semi-pelites, quartzites, and metabasites (figure 2; Davidson and Bethune 1988) mainly preserve upper amphibolite facies assemblages; however, relict orthopyroxene in metabasite, and the assemblage K-feldspar + sillimanite + garnet in pelite suggest local preservation of granulite facies assemblages (Haggart 1991). A high-strain zone on the west side of the metasedimentary package may be responsible for the appearance of granulite facies assemblages at Mill Lake. Between Mill Lake and Beaverstone Bay, granodioritic gneisses dominate. Some mafic dikes in this region are garnet and/or orthopyroxene-bearing, and the gneisses are slightly migmatitic. An east-west trending, post-tectonic mafic dike, presumed to be part of the ~590 Ma (Kumarapeli 1992) Grenville diabase dike swarm, is exposed locally along Collins Inlet.

Geologic and geochronologic data indicate that orthogneisses within the GFTZ along Collins Inlet can be correlated with rocks west of the Grenville Front. U-Pb zircon data from granitic gneisses (1730 Ma, Krogh et al. 1971; ~1715 Ma, Davidson et al. 1992) and local preservation of volcaniclastic textures (Davidson and Bethune 1988; Haggart 1991) suggest that some GFTZ gneisses were derived from the Killarney Complex. It is also possible, but less likely, that some granitoid orthogneiss was derived from ~1450 Ma granitic plutons, which are common in the Britt domain to the south (e.g., van Breemen and Davidson 1988; Corrigan 1990; Culshaw et al. 1991*a*) and which are represented immediately north of the study area by the ~1470 Ma Bell Lake granite (van Breemen and Davidson 1988; Davidson and Bethune 1988). Paragneiss and migmatite from the study area and from Tyson Lake, 30 km to the northeast, have yielded ages of ~1445 Ma (U-Pb monazite, Bethune et al. 1990) and ~1453 Ma (U-Pb zircon, Krogh 1989), consistent with the age of pre-Grenvillian granulite facies metamorphism in the Britt domain (Ketchum et al., 1992). Grenvillian metamorphism produced garnet + orthopyroxene-bearing assemblages in 1238 Ma Sudbury database dikes at ~1032–985 Ma (Bethune et al. 1990; Bethune 1991; Bethune pers. comm. 1992). However, LaTour and Fullagar (1986) concluded that the effect of the Grenvillian Orogeny at Coniston, 100 km northeast of the study area, was limited to a minor thermal event affecting pre-existing mylonites. Existing data therefore define a significant pre-Grenvillian history for the western end of the GFTZ, but the nature and timing of Grenvillian metamorphism, deformation, and exhumation remain unclear.

Methods

Samples were obtained from all the major lithologies along Collins Inlet, including granitic orthogneiss, paragneiss, and amphibolite. Handpicked, mono-mineralic separates were prepared using standard magnetic and heavy liquid techniques. Where possible, multiple mineral separates (zircon, titanite, hornblende, mica, K-feldspar) were obtained from single or closely spaced samples (figures 3, 4, 5, 6). Petrographic descriptions of all dated minerals are presented in Appendix I. The Appendices may be obtained from *The Journal of Geology* free of charge upon request.

U-Pb dating of titanite and zircon was done at the Jack Satterley Geochronology Laboratory at the Royal Ontario Museum, Toronto. The techniques used in this study follow those outlined by Krogh (1973, 1982) and Schärer et al. (1986). Eleven titanite fractions and one zircon fraction, distinguished on the basis of morphology, were separated from eight samples taken from 1 to 16 km east of the Grenville Front (figure 3; Appendix IA). Analytical data are given in Appendix II.

$^{40}Ar/^{39}Ar$ dating of hornblende, muscovite, biotite, and K-feldspar was conducted at Dalhousie University using analytical procedures described by Muecke et al. (1988). Samples CI-37H, GF-8H, GF-35H, GF-7B, GF-35B, GF-19B, and GF-8B were analyzed in a modified MS10 mass spectrometer coupled to a quartz tube heated by an external Lindberg furnace. The remaining samples were analyzed in a VG3600 mass spectrometer coupled to an internal tantalum resistance furnace of the double-vacuum type. Hornblende MMhb-1, with a K-Ar age of 519 ± 3 Ma (Alexander et al. 1978), was the standard used for all analyses. Complete analytical data are presented in Appendix III.

U-Pb Results

Results for the titanite and zircon fractions are plotted on a single concordia diagram (figure 3) to illustrate the systematic distribution of their $^{207}Pb/^{206}Pb$ ratios. Ten of 12 fractions fall along a single discordia, with a 74% probability of fit (using the method of Davis 1982), between an upper intercept of 1454 ± 8 Ma and a lower intercept of 978 ± 13 Ma. There is a strong correlation between degree of discordance and distance from the Grenville Front, with data plotting near the upper intercept coming from samples closest to the Front (inset, figure 3). In the case of sample TK83-40, the only sample from which both zircon and titanite were obtained, the zircon data plot close to the upper intercept, and the titanite data plot close to the lower intercept (TK83-40Z and TK83-40a respectively, figure 3). Sample 91ND-91B, from the southeastern end of Beaverstone Bay, plots 0.25% below the concordia curve at 975 ± 3 Ma (figure 3). The upper intercept is consistent with both the age of granulite facies metamorphism (Krogh 1989; Bethune et al. 1990; Ketchum et al. 1992) and the age of granitoid plutons in the region (van Breemen and Davidson 1988; Corrigan 1990). The lower intercept of 978 ± 13 Ma is similar to the poorly defined ~1032–985 Ma age for growth of metamorphic zircon in the Sudbury dikes (Bethune 1991; Bethune pers. comm. 1992). The samples were taken from a varied suite of gneisses that may have included ~1450 Ma plutons as well as ~1450 Ma high-grade metamorphic rocks. Accordingly, not all units

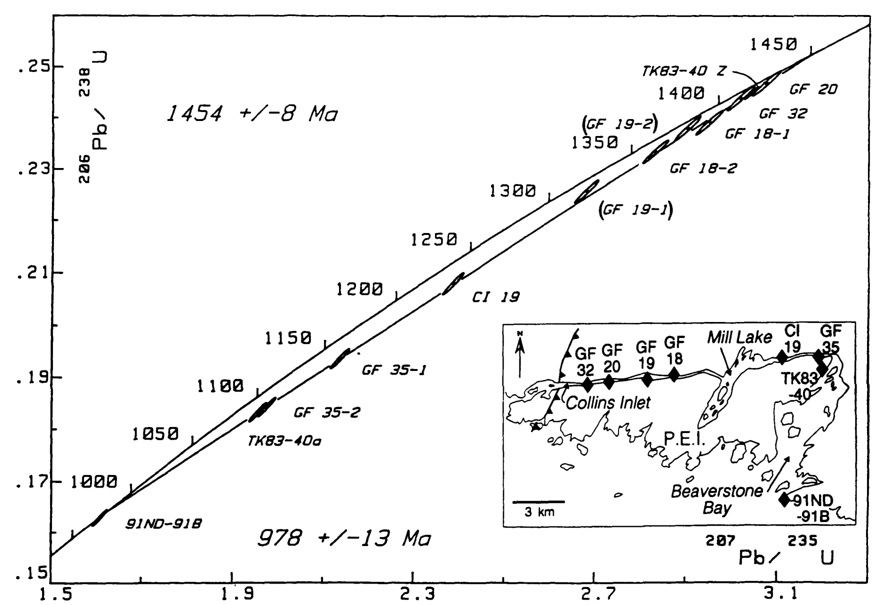

Figure 3. U-Pb data from one zircon (TK83-40Z) fraction and 11 titanite fractions (all other data). Discordia line includes all data except the two titanite fractions from GF-19 (numbers in parentheses), as discussed in the text. Note correspondence between distance from Grenville Front (inset) and titanite discordance; samples furthest from the Front plot closest to the lower intercept.

along the transect may have had identical titanite ages prior to the onset of Grenvillian deformation, and consequently there is no reason to require the inclusion of any single titanite fraction when drawing a discordia line through the data. Of the eight samples, GF-19 is significantly less deformed than the others (Appendix IA) and may represent a somewhat younger pluton. Data from this sample have been excluded from the calculation of the discordia line shown in figure 3, although the two fractions (GF-19-1, GF-19-2) show the same pattern of discordance as other titanites.

The correlation between distance from the Grenville Front and degree of titanite discordance has three possible explanations. New titanite growth at ~980 Ma could have produced a mixed population of old and new grains or could have formed new rims on old grains. In either case, systematic generation of new titanite in greater amounts with increasing distance from the Grenville Front would be required to produce the observed pattern. However, the titanites dated were from granodioritic, amphibolitic, and granitic gneisses, in which the stability fields and volumes of titanite produced by various reactions probably differ. Most fractions consisted of grains of similar morphology, and evidence for metamorphic rims is lacking.

An alternative explanation is that at ~980 Ma the samples were affected by an episode of lead loss which increased in intensity to the east. Its effects were strong enough to reset titanite from the eastern end of the transect substantially (samples CI-19, GF-35, TK83-40a, 91ND-91B), but not strong enough to induce significant discordance in zircon from one of these samples (TK83-40Z). The processes governing diffusion of lead in titanite are not well understood, and few examples exist in the literature of titanite discordance arrays caused by variable degrees of lead loss. Hanson et al. (1971) attributed a discordant titanite array in the contact aureole of a pluton to the short-lived (<1 m.y.) thermal effects of intrusion. In the western Norwegian Caledonides, U-Pb data from 38 titanites from orthogneisses lie on a single discordia line along which discordance and metamorphic grade are strongly correlated (Tucker et al. 1987; Tucker 1991). The inferred duration of the metamorphic event responsible for the discordance was ~4 m.y. The results of these studies suggest that discordance arrays are caused by brief, primarily thermal events, although strain-induced recrystallization may facilitate lead diffusion (Tucker et al. 1987). This type of array does not represent closure in response to slow cooling, because evolution of U-Pb systems wholly or partly open to diffusion for any duration will not produce a colinear array (Tucker et al. 1987), and simple cooling could not explain the zircon and titanite data from sample TK83-40.

By analogy with these interpretations, we suggest that the discordance array from the GFTZ transect resulted from partial opening of the U-Pb system during a thermal event that increased in intensity (duration and/or temperature and/or depth) toward the east. Although deformation within the GFTZ may have facilitated lead loss in general, there is no systematic increase in strain from west to east along Collins Inlet (Frarey 1985; Davidson and Bethune 1988), nor any correlation between intensity of strain and degree of discordance in the dated samples (Appendix IA). The duration of the event cannot be precisely determined from the present data. However, as radiogenic lead remained in titanite despite temperatures high enough to initiate diffusion, it must have been brief, presumably within the limits imposed by the error on the lower intercept (e.g., Tucker et al. 1987).

$^{40}Ar/^{39}Ar$ Results

Hornblende. Hornblende age spectra are shown in figure 4. The three samples closest to Beaverstone Bay yielded plateaus (as defined by Fleck et al., 1977) with ages of 996 ± 5, 994 ± 5, and 979 ± 9 Ma. $^{37}Ar/^{39}Ar$ ratios over each of these plateaus are similar, and in the two cases where microprobe data are available, the measured Ca/K ratios agree well with the $^{37}Ar/^{39}Ar$ data (figure 4). Both the (weighted) mean age of 993 ± 4 Ma and the age of the easternmost sample (GF-35H; 979 ± 9 Ma) overlap with the U-Pb discordia lower intercept value.

There are two possible interpretations of these results. The mean age of 993 Ma can be interpreted as the time of cooling through hornblende closure temperatures (450 ± 50°C; Onstott and Peacock 1987; Baldwin et al. 1990). In this case, the thermal event that induced partial lead loss in titanite apparently caused complete resetting of the $^{40}Ar/^{39}Ar$ system in hornblende in the eastern part of the transect. Alternatively, if the near-concordant 975 Ma titanite age from sample 91ND-91B dates the thermal event, the best approximation to a hornblende cooling age may be the 979 ± 9 Ma date from sample GF-35H. In this case the ~995 Ma ages represent either incomplete resetting or the

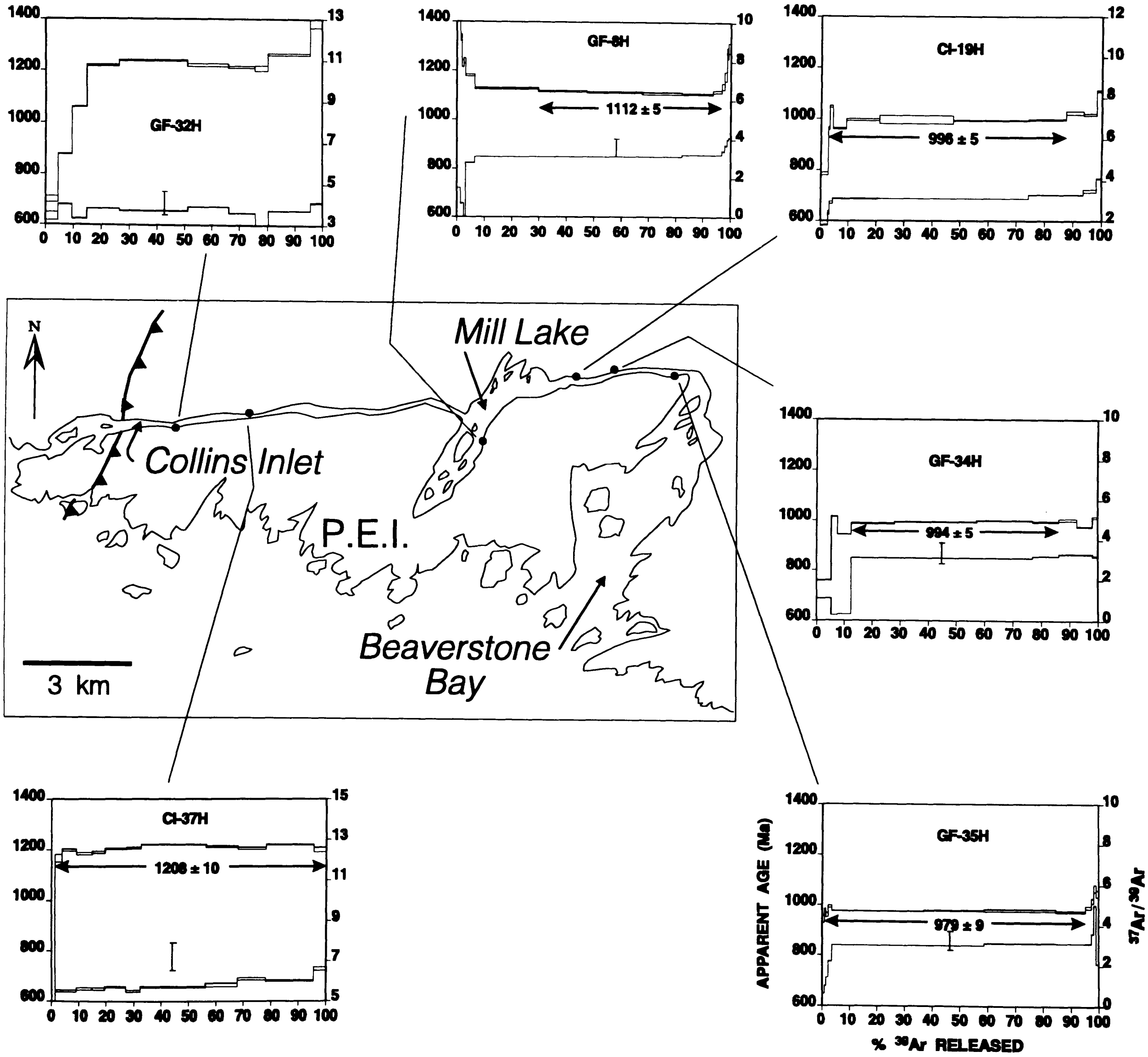

Figure 4. $^{40}Ar/^{39}Ar$ age spectra (upper graphs) and measured $^{37}Ar/^{39}Ar$ ratios (lower graphs) for hornblendes. Vertical lines associated with the $^{37}Ar/^{39}Ar$ plots (except CI-19H) show the $^{37}Ar/^{39}Ar$ ratios (and errors) inferred from microprobe Ca/K analyses. All age spectra are plotted at the same scale to emphasize the variation in age along the transect. Half-heights of open rectangles indicate the 1 σ relative (between-step) uncertainties.

effects of excess ^{40}Ar. Either interpretation is compatible with other data from the transect, as discussed below.

Samples nearest the Grenville Front (GF-32H and CI-37H) yielded discordant spectra without age plateaus. For CI-37H, the $^{37}Ar/^{39}Ar$ ratio inferred from microprobe-determined Ca and K concentrations is higher than the measured isotope ratios (figure 4). In thin section, hornblende is altered along cleavage planes and fractures to an unidentified cryptocrystalline brown phyllosilicate and locally intergrown with biotite. Either feature could be responsible for the apparent K enrichment and/or Ca loss. For GF-32H, the agreement between measured and inferred $^{37}Ar/^{39}Ar$ ratios is much better. In spite of these differences, the total gas ages of the two samples are very similar at ~1200 Ma. This age is probably not the time of cooling following a geologic event, since neither sample is close to any of the ~1238 Ma Sudbury dikes, and there are no other rocks of about this age known in the vicinity. It can be argued that an episode of gas overpressure during cooling (e.g., Foland 1979; Roddick et al. 1980) has contaminated the samples with excess ^{40}Ar and thus raised their ages. However, since these hornblendes have dissimilar K_2O

contents (1.24% K_2O for GF-32H, 0.72% for CI-37H) but similar ages, we think it more likely that the data reflect the partial loss of radiogenic ^{40}Ar at ~980 Ma, the time of the proposed thermal episode. This is in contrast with the complete loss observed farther east and consistent with the observed pattern of titanite discordance.

Sample GF-8H, from Mill Lake in the central part of the transect, has a broad U-shaped spectrum typical of hornblendes that contain excess argon (e.g., Harrison and McDougall 1981). In the present case, however, the apparent $^{37}Ar/^{39}Ar$ ratios (figure 4) suggest that the initial and final high ages are not produced by the principal hornblende phase in the rock. In ideal circumstances, excess argon can be detected and its effect removed using the isotope correlation method (Roddick et al. 1980). When the present data (excluding the high initial and final ages) are treated in this manner (i.e., $^{36}Ar/^{40}Ar$ plotted against $^{39}Ar/^{40}Ar$), the blank-corrected points plot in a tight cluster close to the $^{39}Ar/^{40}Ar$ axis and do not define a straight line. Consequently, we interpret the plateau value of 1112 ± 5 Ma as a partially reset age, comparable to those obtained from samples farther to the west.

Mica. Muscovite and biotite age spectra are shown in figure 5. Muscovite from paragneiss on the north side of Mill Lake (ND-88; figure 5) is associated with calcite veinlets and fringes K-feldspar porphyroclasts; it clearly post-dates the main foliation and is interpreted as a late-stage retrograde phase. The sample yielded a nearly concordant age spectrum with a plateau at 930 ± 5 Ma, interpreted to represent the time of cooling through 320 ± 40°C (Zeitler 1985).

With the exception of sample GF-8B, the biotite spectra are discordant. The mean ages (as defined in figure 5) range from ~950 Ma to ~1250 Ma, and no age trend along the transect is apparent. Because the ages of samples GF-19B and GF-35B exceed the hornblende values at comparable points along the transect, and because the apparent ages of neighbouring samples (GF-7B and GF-8B) are quite different, the common and variable occurrence of excess ^{40}Ar in biotite appears likely, as has been noted in other studies in the Grenville Province (e.g., Dallmeyer and Rivers 1983; Anderson 1988; Owen et al. 1988; York et al. 1991). We conclude that rocks in the Mill Lake area (and presumably also to the east) were heated sufficiently at ~980 Ma to completely reset biotite, and that subsequent cooling to the closure temperature (~300°C) occurred no earlier than ~950 Ma.

K-feldspar. All the K-feldspar age spectra are discordant (figure 6). Initial apparent ages of ~600–800 Ma increase to values of ~900–1000 Ma at the

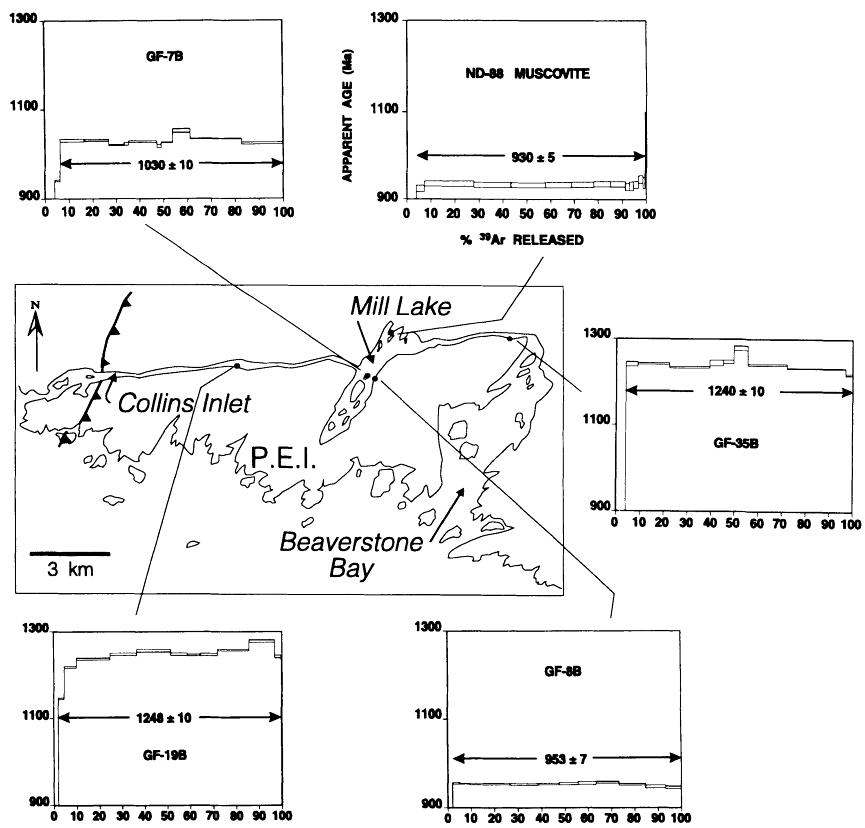

Figure 5. $^{40}Ar/^{39}Ar$ age spectra for biotite and muscovite, plotted on the same vertical scale to emphasize the variations in ages. The age indicated for muscovite is the plateau age. Error bars as for figure 4.

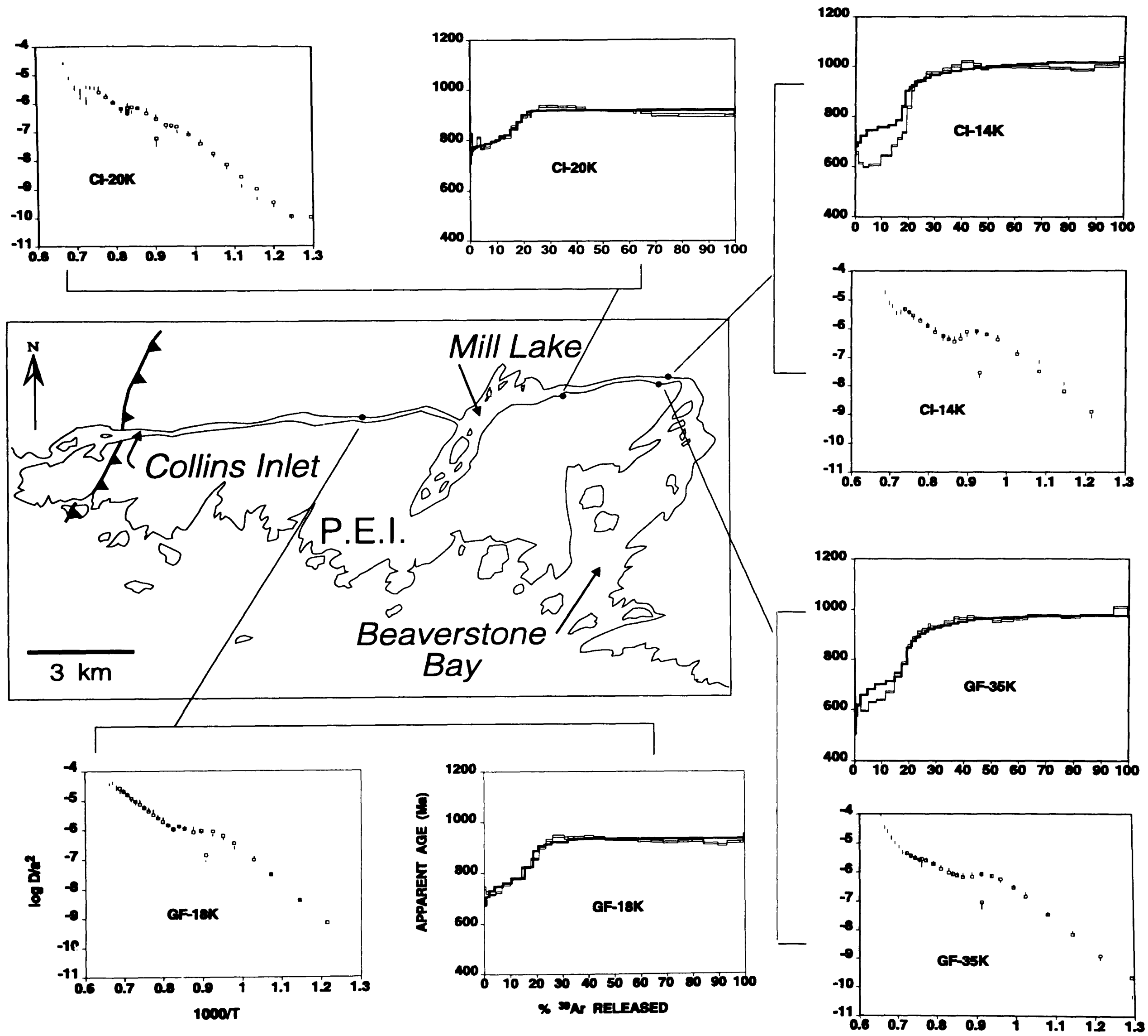

Figure 6. $^{40}Ar/^{39}Ar$ age spectra and Arrhenius plots for K-feldspars. In age spectrum plots, observed data are represented by open rectangles representing 1 σ errors (as for figures 4 and 5), and model results are plotted as single line segments. For the Arrhenius plots, D = diffusion coefficient, a = effective diffusion radius (equivalent to domain size), T(°K) = laboratory extraction temperature. Values (and associated errors) derived from the step-heating data are represented by short vertical line segments, and model results are represented by open squares. See text for discussion.

highest extraction temperatures, a common pattern in K-feldspar spectra from the Grenville Province (Hanes et al. 1988; Culshaw et al. 1991*b*; Cosca et al. 1991) and elsewhere (e.g., Lovera et al. 1989).

We have interpreted our data using multidomain diffusion theory (Lovera et al. 1989, 1991; Harrison et al. 1991), which assumes that argon in K-feldspar is released from discrete sets of domains differing in size and possibly also in activation energy. To accompany the apparent age spectra, Arrhenius plots based on the ^{39}Ar step-heating data were calculated for all four samples (figure 6). In order better to define the Arrhenius plot, one point was produced by "cycling" (Lovera et al. 1991), that is, by reducing the temperature by about 100°C at an appropriate time in each analysis and increasing the heating time to 5–10 hrs. Model Arrhenius plots, assuming three-domain spherical geometry, were generated for each feldspar using the extraction temperatures and heating times from the respective step-heating analyses. In each case, model parameters (E, activation energy; D_o/ρ^2, ratio of frequency factor to square of domain radius;

F, the fractional contribution of each domain to the total gas) were varied in a systematic fashion. The optimum solution was where the sum of the squares of differences between the model and the data-based Arrhenius plots was a minimum. For each final model Arrhenius plot, a corresponding model age spectrum was calculated; an optimum fit to the observed spectrum was achieved by varying the model ages of the domains and the assumed cooling rates. Domain closure temperatures were calculated from the equation derived by Dodson (1973).

The measured age spectra obtained from samples GF-35K and CI-14K are similar, with ages increasing in a regular fashion from initial values of ~600 Ma to final values of ~975–1000 Ma. As these initial values are very similar to the age of the Grenville dike swarm (Kumarapeli 1992), we suggest that heat from this event caused gas loss from these feldspars, in greatest proportion from the smallest domains. No attempt was made to model this gas loss. Consequently, the initial segments of the two model age spectra shown in figure 6 are arbitrary. Modeling of the data-derived Arrhenius plots was terminated at the point in each analysis where there was a marked deviation from linearity, an indication of the onset of incongruent melting. In GF-35K, this break occurred at 1110°C, in CI-14K at 1080°C. Past these points in the ^{39}Ar release (~50% for GF-35K, ~45% for CI-14K), the model age spectra can be expected to fit the observations only in an average sense.

Sample CI-20K, about 3 km west of these two, has a distinctly different age spectrum (figure 6). There is no evidence of the ~600 Ma overprint, and the maximum age attained is much lower at ~935 Ma. This peak age, reached relatively early in the spectrum, is followed by values which decrease to ~900 Ma, in contrast to the patterns obtained from samples GF-35K and CI-14K (see further discussion below). The fourth feldspar sample, GF-18K, is located between Mill Lake and the Grenville Front. Its apparent age spectrum and Arrhenius plot are similar to those from CI-20K (figure 6). The ~980 Ma thermal event evidently completely reset K-feldspar in this part of the transect.

Thermal History

The thermal history of the GFTZ at the eastern end of the transect (Mill Lake–Beaverstone Bay) is depicted in figure 7. Tucker et al. (1987) showed that lead diffuses in titanite at temperatures as low as 480°C, and that complete resetting occurs between 575 and 640°C, even for very brief heating.

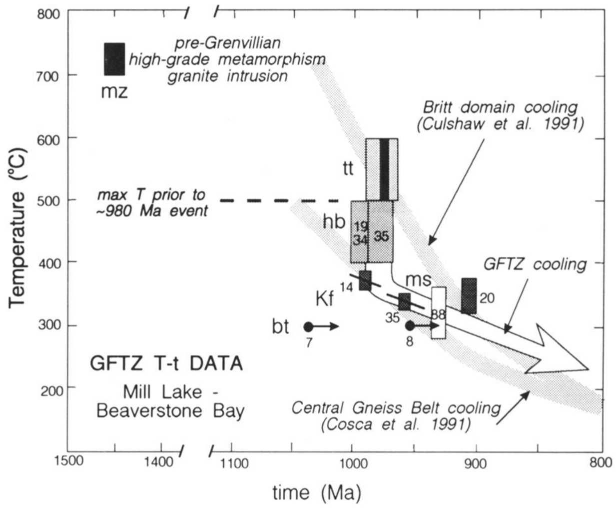

Figure 7. Temperature-time data for the GFTZ between Mill Lake and Beaverstone Bay. Time of high-grade metamorphism (U-Pb monazite) from Bethune et al. (1990). Titanite (tt) range based on lower intercept shown in figure 3 as discussed in text; black bar is the age of the sample closest to the lower intercept (91ND-91B). Hornblende (hb) ranges shown for three samples (19 = CI-19H; 34 = GF-34H; 35 = GF-35H); data from 19 and 34 may not represent true cooling ages, as discussed in text. Muscovite (ms; white box) represents data from sample ND-88 (figure 5); closure temperature used is 320 ± 40°C (Zeitler 1985). K-feldspar (Kf) age and closure temperature ranges based on large domain model results, as discussed in text (14 = CI-14K, 35 = GF-35K, 20 = CI-20K; figure 6); short lines through points 15 and 35 represent approximate cooling rates based on the models. As discussed in text, biotite (bt) ages are maxima owing to probable excess ^{40}Ar (7 = GF-7B, 8 = GF-8B; figure 5). The broad arrow represents a generalized GFTZ cooling path consistent with these data. See text for further explanation.

This implies peak temperatures of 500–600° C for the thermal event that affected titanite, as shown in figure 7.

The array of titanite data along a single discordia demonstrates that no other events caused discordance in titanite between ~1450 and ~980 Ma (cf. Tucker et al. 1987). This does not rule out GFTZ deformation during this time interval but precludes heating of GFTZ rocks in the study area above ~500°C. Consequently, high-grade metamorphic rocks and associated structures cut by the 1238 Ma Sudbury dikes (e.g., Bethune 1989) must be part of (or older than) the ~1450 Ma event.

Hornblende plateau ages are plotted at 450 ± 50°C, the assumed closure temperature (cf. Onstott and Peacock 1987; Baldwin et al. 1990). The muscovite plateau age is plotted at 320 ± 40°C (Zeitler

1985), and biotite data from Mill Lake (figure 5) are plotted at a nominal closure temperature of 300°C. The arrows indicate that due to the probable presence of excess ^{40}Ar, the plotted biotite ages must be considered maximum values.

Calculated closure temperatures for the largest K-feldspar domains are plotted with the corresponding domain model ages in figure 7. Model results for GF-35K yielded closure temperatures of 340 ± 8°C at 960 Ma, and for CI-14K 365 ± 13°C at 990 Ma; the associated uncertainties in these data (figure 7) reflect the sensitivity of the least-squares minima to changes in model parameters. The younger of the two, GF-35K (340°C at 960 Ma) was a cleaner separate and also produced a better fit of model to observed age spectra (figure 6); we therefore regard the results from this sample as the more reliable. For CI-20K, the best-fit models tend to give relatively high values for closure temperature and cooling rate (e.g., ~350 ± 30°C at 15°/m.y.), apparently inconsistent with the relatively low model age (900 Ma). Since the shape of the age spectrum for this sample (figure 6) is not that of a single stable feldspar phase that has undergone slow cooling, and therefore violates the criteria on which these model calculations are based (e.g., Lovera et al. 1989), we conclude that a cooling rate cannot be determined for this sample and that the closure temperature can only be approximated (figure 7).

Between Mill Lake and the Grenville Front, temperature was high enough to induce some discordance in titanite, but not high enough to fully reset hornblende. Biotite shows the effects of excess ^{40}Ar, and muscovite and K-feldspar were fully reset. Although a *T-t* history cannot be determined on the basis of these data, it is unlikely that temperature exceeded 500°C for any significant period, and could have been below 500°C if resetting of isotopic systems in this part of the transect was enhanced by strain (e.g., Desmons et al. 1982; Berry and McDougall 1986).

The data (notably coexisting titanite, hornblende, and K-feldspar from sample GF-35) show that the region between Mill Lake and Beaverstone Bay cooled rapidly from peak temperatures of ≤600°C to temperatures of ~350°C, before slowing to ~1°C/m.y. (figure 7). This pattern contrasts with the cooling histories from the Britt domain, 30 to 80 km southeast of the study area (Culshaw et al. 1991*b*), and the Central Gneiss Belt (Cosca et al. 1991), which both show much lower cooling rates over the interval 500–300°C (figure 7). Although these studies were based on regional sample distributions, and thus may not have detected variations in cooling rates over small areas, it is more likely that the contrast in cooling rates is a real effect related to the different tectonic settings—orogenic front vs. orogenic interior. Preliminary data from the GFTZ immediately east of the transect (Grant 1993) indicate that the zone of rapid cooling may not extend east of Beaverstone Bay. What process could produce the brief thermal pulse (necessary to explain the titanite data) followed by very rapid cooling in this part of the GFTZ without producing similar effects in the interior of the orogen?

Tectonic Interpretation

The thermal history shown in figure 7 suggests that the rocks now exposed along the Collins Inlet transect were heated briefly to temperatures of 500–600°C at 978 ± 13 Ma and cooled rapidly through ~350°C. The absence of any intrusions in this part of the GFTZ between 1238 Ma and 590 Ma rules out magmatic heat as the source of the disturbance. Metamorphism of the Sudbury diabase in the GFTZ 30 km northeast of the study area suggests peak *P-T* conditions of ~580–780°C and 3–8 kbar were achieved at ~1032–985 Ma (Bethune et al. 1990; Bethune 1991). Structural and seismic data indicate a tectonic regime dominated by NW-directed thrusting (Davidson 1986*a*, 1986*b*; Davidson and Bethune 1988), with SE-dipping reflectors coincident with the GFTZ extending into the lower crust along strike from the study area (Green et al. 1988). It therefore seems reasonable to assume that the thermal disturbance and subsequent rapid cooling were related to tectonic processes associated with thrusting.

Geodynamic models of convergent orogens (Beaumont et al. 1992; Willett et al. 1993) suggest that detachment and underthrusting (subduction) of the lower lithosphere at the crust-mantle boundary creates a pair of step-up shear zones (figure 8*a*) and associated mechanical wedges, one facing toward the "fixed" foreland ("retro-wedge"), and one facing into the direction of material transport ("pro-wedge"). As convergence proceeds, the models predict development of an antiform in the hangingwall of the retro-ward step-up shear zone; erosion of the upper surface of the retro-wedge accentuates development of this structure and leads to rapid exhumation of hangingwall rocks. Migration of the step-up shear zone into the foreland causes rocks previously in the footwall to be exhumed; high erosion rates on the retro-wedge limit crustal thickening and cause the shear zone

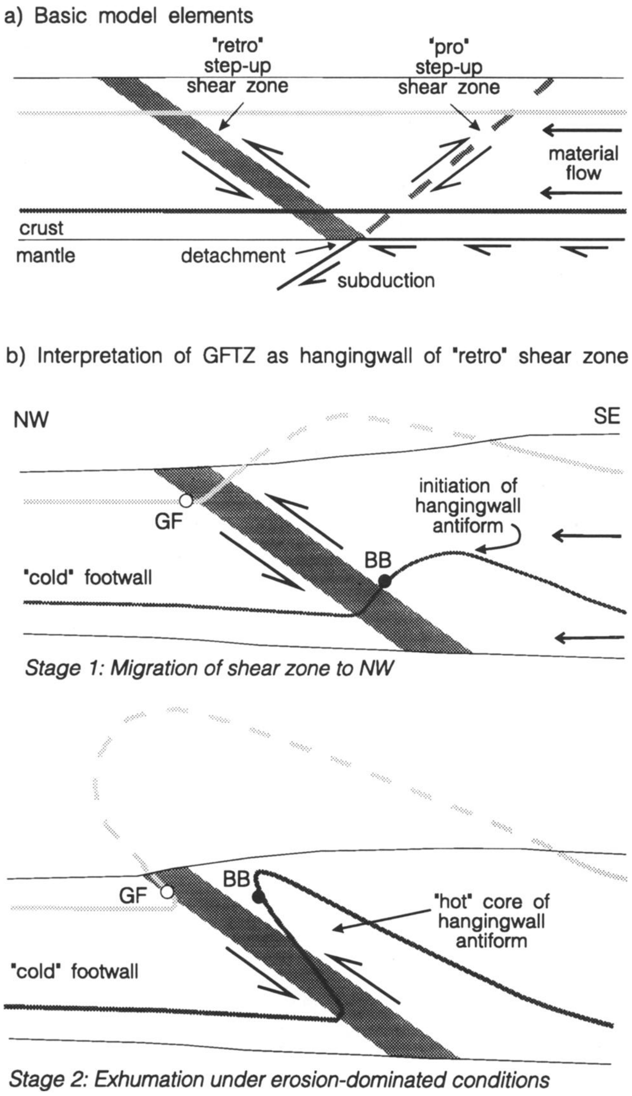

Figure 8. Tectonic interpretation. (*a*) Geometric configuration of "pro" and "retro" step-up shear zones and detachment point (after Beaumont et al. 1992; Willett et al. 1993; Beaumont and Quinlan 1993) at onset of convergence, detachment, and underthrusting. Horizontal arrows show direction of material transport. The horizontal reference lines represent the pre-980 Ma crustal levels of rocks now exposed at the Grenville Front and Beaverstone Bay (GF and BB in lower panels); these do not correspond to real geological horizons but can be considered approximations to isotherms. The location of the "retro" step-up shear zone relative to the present-day Grenville Front at the onset of this stage of convergence is not known, nor are the positions of the detachment point and "pro" step-up shear zone, which could have been some distance to the southeast. (*b*) Interpretation of the GFTZ in the study area as the hangingwall of the "retro" step-up shear zone. Points GF and BB correspond to rocks now exposed at the Grenville Front and Beaverstone Bay, respectively. *Stage 1.* Migration of "retro" step-up shear zone into transect area at ~980 Ma, and initial development of hangingwall antiform. *Stage 2.* Exhumation of hangingwall rocks under erosion-dominated conditions; shear zone has stopped migrating into the foreland, and hangingwall antiform becomes rapidly accentuated by erosion. Model results on which this interpretation is based (model 1, figure 2, of Beaumont and Quinlan, 1993) suggest a time interval between stages 1 and 2 of ≤5 m.y., with total convergence of ~100 km. See text for full discussion.

to stop propagating into the foreland. The type of crustal structure predicted by the models bears a remarkable resemblance to the structure of the GFTZ shown on GLIMPCE seismic profile "J" (Green et al. 1988; fig. 20 of Beaumont and Quinlan 1993).

Figure 8 illustrates a tectonic interpretation based on these models; details of the numerical model used as an analogy can be found in Beaumont and Quinlan (1993). We interpret the Grenville Front to be the northwestern boundary of the step-up shear zone associated with the retro-wedge, since geological data indicate northwesterly transport of material from the interior of the Grenville orogen toward the stable foreland (represented in the transect area by the Southern Province). We envisage rocks now at the Grenville Front (GF) and Beaverstone Bay (BB) to have been at upper and lower crustal depths respectively (corresponding to $T < 500°C$ in the stable foreland; depth range compatible with metamorphic data from the Sudbury diabase, Bethune 1991) prior to the propagation of the step-up shear zone (reference lines, figure 8*a*). Migration of the step-up shear zone toward the foreland at ~980 Ma brings these points into the shear zone (stage 1, figure 8*b*) and may have caused some crustal thickening and consequent burial of mid- to lower crustal rocks. Rapid cooling of rocks in the transect region occurs in response to erosion of the upper surface of the retro-wedge. This causes the shear zone to stop migrating, and accentuates development of the hangingwall antiform, leading to extremely rapid exhumation (≤5 m.y.) of rocks in the hangingwall (stage 2, figure 8*b*). Rocks at GF undergo substantial shear but relatively little exhumation, whereas rocks at BB are brought rapidly into the upper crust. If rocks at position BB were in the lower crust prior to the migration of the shear zone into the western GFTZ (as depicted in figure 8), they would have been initially hotter than rocks at GF, and would also have been heated by deeper rocks brought up in the core of the hangingwall antiform during exhumation. This interpretation is capable of explaining both the short-lived thermal event that induced partial lead loss in titanite and the rapid cooling of rocks at the eastern end of the transect. It does not require deep burial of the entire GFTZ by overthrusts, nor extreme crustal thickening in this region at this time.

This model can account for both the rapid cooling and the GLIMPCE seismic data, but how does it compare with the geological evidence? In the absence of well-defined marker horizons within the mainly granitoid orthogneisses of the GFTZ, it is not possible to identify an "antiform," and therefore not possible to test this geometric prediction of the model. The presence of discrete shear zones within the transect area (for example, at Mill Lake) suggests the possibility of structural breaks within the GFTZ. Although continuum mechanics models do not allow for discontinuities, this type of heterogeneous strain is to be expected in real rocks and can be regarded as compatible with the overall interpretation. The characteristic GFTZ strain pattern, with fabrics outside shear zones parallel to those within shear zones, is compatible with model predictions, since model elements undergo both pure and simple shear (Beaumont et al. 1992). We therefore regard the model shown in figure 8 as generally consistent with the geology, although there are differences in detail.

What change in orogenic boundary conditions induced this episode of convergence? Was this an isolated, discrete event, or merely the last stage in a continuum? What are the implications of this interpretation for the orogen as a whole? A discordant titanite array with a lower intercept of 993 Ma, obtained from ~1650 Ma metaplutonic rocks south of the Grenville Front in central Labrador (Thomas et al. 1986), suggests a similar tectonic regime may have operated in this part of the orogen, despite the differences in the present distribution of metamorphic rocks between the two regions. Geochronological data from many other places along the Grenville Front indicate a thermal event at ~990 Ma (e.g., Krogh 1991; Childe et al. 1992). Given along-strike variations in the pre-Grenvillian rocks of the foreland, and likely variations in erosion and exhumation rates along the GFTZ, our interpretation could explain many of the local variations in GFTZ geology as well as the general features of this zone. However, more data are needed on cooling histories of other parts of the GFTZ and on the correlation, if any, between large-scale crustal structure and thermal history before a tectonic model can be developed for the entire northwest margin of the Grenville orogen.

Conclusions

1. High-grade metamorphic assemblages at the western end of the GFTZ pre-date the Grenvillian orogeny and were exhumed, with little or no metamorphic overprinting, by Grenvillian deformation.

2. The U-Pb titanite data from the transect form a discordant array with an upper intercept of 1454 ± 8 Ma and a lower intercept of 978 ± 13 Ma. These data are best interpreted in terms of a short-lived thermal event that increased in intensity from west to east away from the Grenville Front. The data preclude any event producing temperatures above 500°C between 1450 and ~980 Ma.

3. $^{40}Ar/^{39}Ar$ data from hornblende at the eastern end of the transect indicate cooling through ~450°C at 993–979 Ma, although the older ages in this range may not be true cooling ages. K-feldspar models indicate cooling through 365–340°C at 990–960 Ma. Muscovite data indicate cooling through ~320°C at 930 Ma, but biotite data are not easily interpreted owing to the effects of partial resetting and/or excess ^{40}Ar.

4. The thermal history, which suggests a short-lived thermal event followed by very rapid cooling, can be explained in terms of exhumation in the hangingwall of a crustal-scale step-up shear zone during a ~980 Ma episode of convergence.

5. This study shows the usefulness of $^{40}Ar/^{39}Ar$ K-feldspar data, interpreted using multi-domain diffusion theory, in delineating thermal histories, especially in cases where the effects of cooling, partial resetting, and excess ^{40}Ar on $^{40}Ar/^{39}Ar$ age spectra are not easily distinguished.

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