Stabilization and reactivation of cratonic lithosphere from the lower crustal record in the western Canadian shield

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Abstract New U–Pb geochronology for an extensive exposure of high-pressure granulites in the East Lake Athabasca region of the western Canadian shield is consistent with a history characterized by 2.55 Ga stabilization of cratonic lithosphere, 650 million years of lower crustal residence and cratonic stability, and 1.9 Ga reactivation of the craton during lithospheric attenuation and asthenospheric upwelling. High precision single-grain and fragment zircon data define distinctive discordia arrays between 2.55 and 1.9 Ga. U–Pb ion microprobe spot analyses yield a similar range of U–Pb dates with no obvious correlation between date and cathodoluminescence zonation. We attribute the complex U–Pb zircon systematics to growth of the primary populations during a 2.55 Ga high-pressure granulite facies event (~ 1.3 GPa, 850°C) recorded by the dominant mineral assemblage of the mafic granulite gneisses, with subsequent zircon recrystallization and minor secondary zircon growth

during a second high-pressure granulite facies event (1.0 GPa, $\sim 800^{\circ}\text{C}$) at 1.9 Ga. The occurrence of two discrete granulite facies metamorphic events in the lower crust, separated by an interval of 650 million years that included isobaric cooling for at least some of this time, suggests that the rocks resided at lower crustal depths until 1.9 Ga. We infer that this phase of lower crustal residence and little tectonic activity is coincident with an extended period of cratonic stability. Detailed structural and thermochronological datasets indicate that multistage unroofing of the lower crustal rocks occurred in the following 200 million years. Extended lower crustal residence would logically be the history inferred for lower crust in most cratonic regions, but the unusual aspect of the history in the East Lake Athabasca region is the subsequent lithospheric reactivation that initiated transport of the lower crust to the surface. We suggest that a weakened strength profile related to the 1.9 Ga heating left the lithosphere susceptible to far-field tectonic stresses from bounding orogens that drove the lower crustal exhumation. An ultimate return to cratonic stability is responsible for the preservation of this extensive lower crustal exposure since 1.7 Ga.

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Introduction

Understanding the processes responsible for the survival or destruction of continents is a fundamental problem in lithospheric evolution. There is increasing consensus that significant crustal recycling occurred throughout Earth's history, such that the volume of the Earth's continents may have remained roughly constant since early in the Archean (e.g., Armstrong 1981; Bowring and Housh 1995; Harrison et al. 2005). The recognition that Earth's preserved continents are fragmentary remnants of much larger volumes of

past crust raises important questions regarding both the mechanisms that disrupt and recycle continental lithosphere, as well as the characteristics conducive to its long-term survival. Archean cratons are stabilized by anomalously thick and depleted mantle roots that help insulate them from subsequent tectonism (Jordan 1978). However, even these relatively stable regions commonly preserve evidence for subsequent thermal perturbations and magmatic episodes in response to plate margin processes and asthenospheric upwelling events, as recorded in the lower crustal xenolith record (e.g., Bell et al. 2003; Farmer et al. 2005).

Large expanses of high-pressure granulites exposed in the western Canadian shield provide an unusual opportunity to directly examine a lower crustal record that may preserve a signature of craton stabilization and reactivation. The Chipman domain in the East Lake Athabasca region records two high-pressure granulite facies metamorphic events (Mahan et al. 2008). Here we present new geochronological data that constrain 650 million years of lower crustal evolution prior to exhumation of these rocks during final stabilization of the Hearne and Rae cratons. We then consider the implications for a cycle of stabilization, reactivation, and re-stabilization of cratonic lithosphere in this portion of the western Canadian shield.

Geological setting

The western Canadian shield is a collage of Archean cratons sutured by Paleoproterozoic orogens. The Churchill Province within this region is bounded by the Thelon orogen (2.02–1.91 Ga) to the northwest and the Trans-Hudson orogen (1.88–1.78 Ga) to the southeast (Fig. 1). The Snowbird tectonic zone, extending ~2,800 km from the coast of Hudson Bay to the Canadian Cordillera, is a dramatic, northeast-trending linear gravity anomaly that separates the Rae and Hearne crustal domains of the Churchill Province (Goodacre et al. 1987; Hoffman 1988). The Snowbird tectonic zone has been interpreted as a Paleoproterozoic intercontinental suture (Hoffman 1988; Ross et al. 1995, 2000), a Paleoproterozoic intracontinental transform (Lewry and Sibbald 1980) and an Archean intracontinental strike-slip shear zone (Hanmer 1994, 1997; Hanmer et al. 1995). Recent study attributed the distinctive geometry of the central Snowbird zone to the interaction of major (100 s km in length) Paleoproterozoic intracontinental thrust and strike-slip shear zones (Mahan and Williams 2005). Other recent work suggested that the distinctive geophysical characteristics and ca. 1.9 Ga granulite facies metamorphism along the central Snowbird zone were caused by significant Paleoproterozoic mafic intrusives emplaced along >1,200 of strike length of the feature (Flowers et al. 2006b).

The East Lake Athabasca region spans the central segment of the Snowbird tectonic zone, and includes an extensive tract of granulite facies rocks with peak metamorphic conditions from 1.0 to >1.5 GPa, >750°C (Fig. 2). This region contains high-pressure (high-P) granulite facies rocks within the East Athabasca mylonite triangle along the trace of the Snowbird tectonic zone, high-P granulite facies rocks in the Rae domain to the west, as well as shallower level amphibolite facies rocks within the Hearne domain to the east. To the south, metamorphic basement is overlain by sedimentary rocks of the ca. 1.7 Ga Athabasca basin. The East Athabasca mylonite triangle has been subdivided into three distinct tectonically bounded domains. The Chipman domain is dominated by heterogeneous >3.0 Ga Chipman tonalitic gneisses, contains lenses of mafic and felsic granulites, and preserves the Chipman mafic dike swarm that records metamorphic conditions of 1.0 GPa, 750–850°C (Williams et al. 1995) (Figs. 2, 3). The northwestern domain is composed of ca. 2.6 Ga mafic to felsic plutonic rocks, emplaced and metamorphosed at ~1.0 GPa, 900°C (Williams et al. 2000). The southern domain contains felsic and mafic granulites with minor eclogite, recording peak conditions ≥ 1.5 GPa, ~900°C (Snoeyenbos et al. 1995; Kopf 1999; Baldwin et al. 2003, 2004, 2006). The East Athabasca mylonite triangle is bounded by the contractional Legs Lake shear zone to the east, and by the dextral strike-slip Grease River shear zone to the northwest. Multistage contractional and extensional processes exhumed the region between ca. 1.9 and 1.7 Ga (Mahan et al. 2003, 2006b; Flowers et al. 2006c).

The challenge of distinguishing the signatures of Archean and Paleoproterozoic metamorphism and deformation along the Snowbird tectonic zone is largely responsible for the ongoing controversy over the tectonic significance of this fundamental structure. In the Chipman domain, high-P granulite facies metamorphism (1.0 GPa, 750–850°C) was dated at 1896.2 ± 0.3 Ma using high-precision U–Pb analysis of metamorphic zircon grains from migmatitic mafic granulite dikes (Flowers et al. 2006b). Dike emplacement is also interpreted to be ca. 1.9 Ga based on field relationships that suggest synmetamorphic dike intrusion, and therefore the dikes do not preserve a record of earlier tectonism. The timing of peak conditions in the southern domain was interpreted at 1904.0 ± 0.3 Ma based on high-precision U–Pb dates on metamorphic zircon grains in eclogite, although these same rocks also contain a record of earlier metamorphic zircon growth at ca. 2.55 Ga (Baldwin et al. 2004). Felsic granulites in the Chipman and southern domains contain abundant garnet with inclusions of >2.55 Ga monazite that indicate garnet growth in the Archean. Younger monazite populations are distributed in the felsic granulite matrix (Baldwin et al. 2006; Mahan et al. 2006a). A record of ca. 2.55 and 1.9 Ga metamorphism was

Fig. 1 Geological map of the western Canadian Shield showing major tectonic features. The rectangle marks the East Lake Athabasca region enlarged in Fig. 2. Dark gray shading shows locations of high-P granulites along the Snowbird tectonic zone. Medium gray shading immediately to west of East Lake Athabasca region depicts known extent of granulites in the Rae domain

also documented along the northern trace of the Snowbird tectonic zone, interpreted to date conditions of 0.5–0.6 GPa, 650–750°C, and 0.9–1.1 GPa, 650–750°C, respectively (Stern and Berman 2000; Berman et al. 2002). Based primarily on field relationships, prior workers suggested that the isobarically cooled granulites of the East Lake Athabasca region resided in the deep crust for an extensive portion of their history (Williams and Hanmer 2005). However, the polymetamorphic complexity of these rocks causes their metamorphic history to continue to be debated.

Chipman domain mafic and felsic granulite gneisses

Garnetiferous mafic and felsic granulites occur in all three lithotectonic domains of the East Athabasca mylonite triangle. These lithologies are most abundant in the southern and northwestern domains, but also occur as minor components in the western part of the Chipman domain. The Chipman domain mafic granulites crop out as deformed lenses meters to tens of meters wide, and up to hundreds of meters long, within >3.0 Ga heterogeneous Chipman tonalitic gneisses (Fig. 3). The western Chipman domain coincides with a dramatic magnetic anomaly visible in the total field magnetic map, suggesting that the mafic lithologies may increase in abundance at depth. The restriction of the mafic granulite lenses and associated magnetic anomaly to the western part of the domain, and the less

common occurrence of anorthositic, gabbroic and ultramafic inclusions that are prevalent within Chipman tonalite further east, may reflect a crustal boundary between the eastern and western portions of the domain. Chipman mafic dikes that preserve one granulite facies metamorphic assemblage dated at 1.9 Ga, intrude the entire domain (Williams et al. 1995; Flowers et al. 2006b). The primary chemical and mineralogical distinctions between the mafic granulites of this study and the Chipman mafic dikes are a distinctly lower MgO content and lack of orthopyroxene (either igneous or metamorphic) in the mafic dikes.

Chipman domain mafic granulites include both $\text{Grt} + \text{Cpx} \pm \text{Opx} \pm \text{Hbl} \pm \text{Pl} + \text{Qtz}$ units (mineral abbreviations follow Kretz 1983), and medium to very coarse-grained $\text{Grt-absent Opx} + \text{Pl}$ units. This study is primarily focused on the former, because it is possible to extract more information about the metamorphic history from the garnet-bearing unit. Accessory minerals include ilmenite, titanite and apatite, with zircon and rutile in some samples (Fig. 4a). Petrologic analysis and thermobarometric study of Chipman domain mafic granulite samples by Mahan et al. (2008) documented: (1) an early assemblage of $\text{Grt}_1 + \text{Cpx}_1 \pm \text{Pl}_1 + \text{Qtz}$ that preserves conditions of ~ 1.3 GPa, 850°C that is interpreted to have developed from an igneous Opx -bearing protolith (Fig. 4a–c), (2) retrogression to a $\text{Hbl}_2 + \text{Grt}_2 + \text{Pl}_2$ bearing matrix assemblage (~ 1.0 GPa, 700–750°C) during rehydration (Fig. 4a, d), and (3) a second less pervasive granulite facies assemblage

Fig. 2 **a** Geological map of the East Lake Athabasca region, northern Saskatchewan. References for P–T data: (1) Baldwin et al. 2003; (2) Baldwin et al. 2004; (3) Kopf 1999; (4) Krikorian 2002; (5) Mahan et al. 2003; (6) Mahan and Williams 2005; (7) K. H. Mahan (unpublished data); (8) K. H. Mahan, K. E. Ashton, M. L. Williams

(unpublished data); (9) Snoeyenbos et al. 1995; (10) Williams et al. 1995; (11) Williams et al. 2000; (12) Williams and Hanmer 2005; (13) M. L. Williams (unpublished data). **b** Key to major regional lithotectonic domains. **c** Schematic NW–SE cross-section from A to A' illustrating shear zones that form major domain boundaries

of $\text{Opx}_3 + \text{Cpx}_3 + \text{Pl}_3$ developed at the expense of Hbl_2 that records conditions of ~ 1.0 GPa, 800–850°C (Fig. 4a, d; in 4D note Cpx + Opx surrounding Hbl).

The petrographic setting of zircon grains was documented within mafic granulite sample 02M133A, the same sample on which much petrologic and thermobarometric analysis, high-precision U–Pb dating, and ion microprobe study also focused. Commonly 10–50 μm zircon crystals occur in pod-shaped domains dominated by the $\text{Grt}_1 + \text{Cpx}_1$ assemblage. Zircon crystals also occurs as inclusions in matrix garnet, clinopyroxene, hornblende and plagioclase (Fig. 4b). In situ cathodoluminescence (CL) images of zircon crystals in the M1 assemblage reveal darker unzoned or weakly zoned cores with bright overgrowths, and less commonly, grains with dark overgrowths or homogeneous zircon grains lacking obvious overgrowths. Matrix zircons are characterized by dark cores and bright overgrowths.

Less common garnetiferous felsic granulite gneisses in the Chipman domain are spatially associated with the mafic granulite gneisses. The felsic granulites contain a $\text{Grt} + \text{Sil} + \text{Kfs} + \text{Pl} + \text{Qtz}$ anhydrous peak assemblage, with abundant accessory zircon, monazite and rutile. Mahan et al. (2006a) report more detailed descriptions of mineral assemblages, petrological analysis, and documentation of multiple monazite populations using in situ electron microprobe monazite imaging and dating techniques.

Whole rock geochemistry and Sm–Nd isotopic data

Major and trace element concentrations were determined for five Grt-bearing mafic granulites, five orthopyroxenites, and three Grt-rich felsic granulites from the Chipman

Fig. 3 Geological map of the Chipman domain with locations of geochemistry and geochronology samples. Sample names are provided for the geochronology samples

domain by XRF and ICPMS at the University of Massachusetts, Amherst and Activation Laboratories, respectively (Table 1). The Grt-bearing mafic granulites contain SiO_2 ranging from 47.9 to 52.9%, MgO from 7.0 to 10.6%, and Al_2O_3 from 11.0 to 16.8%. The orthopyroxenites contain overlapping SiO_2 from 48.9 to 53.8%, higher MgO from 13.7 to 26.0%, and lower Al_2O_3 from 3.7 to 7.8%. Bulk compositions of felsic granulites in the Chipman domain are characterized by SiO_2 from 61.0 to 69.8%, MgO from 3.2 to 5.7%, and Al_2O_3 from 13.4 to 17.1%, similar to felsic granulites in the southern domain (Baldwin et al. 2006).

Rare earth element (REE) patterns for the Grt-bearing mafic granulite samples include three that are LREE enriched, one that is flat, and one that is LREE depleted (Fig. 5a). REE patterns for the orthopyroxenites are LREE enriched, either suggesting the presence of garnet in the source or that the precursor lithology was contaminated by a subduction component (Fig. 5a). Most samples do not contain pronounced positive or negative Eu anomalies that would imply the role of plagioclase fractionation during the generation or emplacement of the mafic protolith. The felsic granulite samples are LREE enriched, are characterized by flat HREE patterns, with two of three

samples displaying positive Eu anomalies (Fig. 5b). These patterns are similar to those reported for southern domain felsic granulites (Baldwin et al. 2006).

Whole rock Sm–Nd isotopic data were previously acquired for the same ten Grt-bearing and Grt-absent mafic granulite samples for which major and trace element data are reported in this study (Flowers et al. 2006b). Nine of the samples were characterized by time-integrated LREE enrichment, ϵ_{Nd} values at 2.6 Ga ranged from -7.29 to 2.77 , and depleted mantle model dates from 3.91 to 2.91 Ga. Sm–Nd data for one Grt-bearing granulite was characterized by time-integrated LREE depletion, consistent with its REE pattern. The results display a correlation between SiO_2 content and the ϵ_{Nd} values at 2.6 Ga, consistent with contamination of the mafic granulite protolith by felsic components during its generation and emplacement.

U–Pb geochronology

Analytical methods

U–Pb ID-TIMS analytical methods

Zircon and monazite grains for ID-TIMS (isotope-dilution thermal ionization mass spectrometry) were isolated from the samples by standard crushing, Wilfley table, heavy liquid, and magnetic separation techniques. All analyzed zircon and monazite grains were first mounted in epoxy, polished to expose grain interiors, and imaged to characterize internal structures in order to optimize crystal selection. Zircon grains were imaged with cathodoluminescence (CL) and/or backscattered electron (BSE) imaging techniques. Monazite grains were imaged with BSE methods, and electron microprobe Th, Ca, Y and U compositional maps were acquired for a subset of monazite crystals.

Zircon and monazite grains were extracted from the grain-mount, photographed, and measured. Sizeable ($>100 \mu\text{m}$) zircon and monazite crystals were broken into fragments for analysis, while smaller grains were analyzed in their entirety. Most zircon grains and fragments were air-abraded with pyrite after the method of Krogh (1982), and rinsed and ultrasonicated in 3 M HNO_3 . Monazite grains were rinsed and ultrasonicated in high-purity water. A subset of zircon grains were subjected to a version of the chemical abrasion (CA-TIMS) technique (Mattinson 2005), in which zircon crystals were placed in quartz crucibles in a muffle furnace at $\sim 900^\circ\text{C}$ for ~ 60 h, loaded into Teflon FEP microcapsules and leached in 29 M HF at $\sim 180^\circ\text{C}$ for 12 h, then transferred to savillex beakers and prepared for dissolution according to standard procedures. All fractions were loaded into savillex beakers, washed and

Fig. 4 **a** Mg map of thin section from mafic granulite sample 02M133A. Brighter colors represent higher Mg concentrations. Red is primarily garnet and yellow is clinopyroxene or hornblende. Blue dots represent locations of zircon crystals. Thermobarometric estimates for the M1 assemblage in the coarse-grained Grt + Cpx boudin, and in the Grt + Hbl + Cpx + Opx matrix are from Mahan et al. (2008). BE images show petrographic setting of zircon, with associated CL images of the in situ zircon, in **b** and **c** M1 assemblage, and **d** matrix assemblage

ultrasonicated again in 3 M HNO₃ (for zircon) or high-purity water (for monazite), rinsed in distilled acetone and high-purity water, and loaded into Teflon FEP microcapsules in high-purity water. Samples were spiked with a mixed ²⁰⁵Pb–²³³U–²³⁵U tracer. Zircon grains were dissolved in 29 M HF at 220°C for 48–96 h, and converted to 6 M HCl at 180°C for 12–24 h. Monazite crystals were dissolved in 12 M HCl at 180°C for 48 h. Pb and U were chemically separated using HCl anion exchange chemistry modified after Krogh (1973). Pb and U were loaded on single Re filaments with a silica gel 0.1 M H₃PO₄ emitter solution and analyzed on the Massachusetts Institute of Technology VG Sector 54 mass spectrometer. Pb isotopic ratios were measured either by peak jumping using an axial counting Daly detector, or dynamically with Faraday cups and the Daly detector, peak-jumping ²⁰⁵Pb into the axial position to obtain a real-time Faraday-Daly gain calibration. U was measured as an oxide in static mode on three

Faraday cups. U–Pb data and details regarding isotopic ratio corrections are provided in Table 2. Decay constant and U–Pb tracer calibration uncertainties are systematic errors not included in the cited dates, as these are unnecessary for comparison of data acquired within a single lab. Concordia diagrams for all zircon and monazite grains analyzed by TIMS are shown in Figs. 6, 7, 8, 9 and 10.

U–Pb ion microprobe analytical methods

Zircon crystals for SHRIMP (sensitive high-resolution ion microprobe) U–Pb analysis from one sample previously analyzed by ID-TIMS were prepared in a 1-inch diameter, Au-coated, polished grain-mount. Spots within the zircon grains were selected using optical and CL imaging, and were analyzed on the SHRIMP II at the Australian National University using a 25-µm diameter, 3.5 nA, 10 kV, negative O₂ primary ion beam. Positive secondary ions were

Table 1 Major and trace element data for Chipman domain mafic and felsic granulite gneisses

Sample Lake	Mafic granulites (Gt-cpx-hb-opx)					Orthopyroxenites				Felsic granulites			
	03-6B	03-88A	03-159A	03-168E	03-228A	03-92B	03-125A	03-158A	03-189A	03-20C	03-217A	03-182A	03-175A
	Merret	Bompas	Lytle	Cora	Wasend	Bompas	Lytle	Lytle	Cora	Merret	Wasend	Cora	Cora
SiO ₂	47.91	49.56	52.94	50.92	48.37	50.26	53.81	53.77	50.07	69.75	49.23	61.01	62.85
TiO ₂	0.91	1.01	0.38	0.62	0.48	0.67	0.37	0.39	0.68	0.81	0.79	0.92	0.73
Al ₂ O ₃	14.87	13.78	16.84	15.80	11.02	4.69	3.73	7.64	7.83	13.42	3.91	17.13	17.05
Fe ₂ O ₃	15.16	14.23	8.73	14.02	12.81	13.13	10.55	10.46	13.24	6.02	13.49	8.11	8.49
MnO	0.28	0.26	0.15	0.15	0.25	0.23	0.19	0.19	0.23	0.08	0.22	0.09	0.13
MgO	7.50	6.98	6.96	7.16	10.58	18.03	25.99	18.12	13.71	3.15	22.41	5.71	3.79
CaO	10.72	11.48	10.26	7.78	15.51	10.94	3.80	6.96	12.14	1.42	7.53	2.76	2.77
Na ₂ O	2.00	2.39	3.23	2.72	0.74	1.03	0.79	1.39	1.54	2.18	1.60	2.41	2.66
K ₂ O	0.35	0.22	0.67	0.41	0.02	0.35	0.13	0.57	0.38	3.06	0.28	1.97	1.85
P ₂ O ₅	0.08	0.09	0.04	0.32	0.02	0.08	0.03	0.04	0.06	0.02	0.08	0.05	0.02
Total	99.78	100.00	100.20	99.90	99.80	99.41	99.39	99.53	99.88	99.91	99.54	100.16	100.34
V	241	248	128	195	250	158	97	139	246	149	110	134	134
Cr	170	120	60	240	780	2190	4380	2770	1340	190	4100	640	550
Co	30	47	41	38	46	55	85	60	56	23	75	28	26
Ni	90	110	150	40	140	410	1030	610	250	70	690	170	150
Cu	10	250	90	0	60	90	40	40	120	20	40	20	60
Zn	50	50	60	130	40	0	130	40	0	60	0	30	70
Ga	15	15	15	22	10	7	6	9	11	20	7	18	19
Ge	0	1	1	1	1	0	1	1	1	1	<1	0	1
As	0	0	0	0	0	0	0	0	0	0	0	0	6
Rb	2	0	17	2	0	2	0	13	3	72	6	91	109
Sr	61	55	221	210	80	79	59	79	79	284	172	208	120
Y	21	22	10	10	12	14	4	10	13	35	9	33	32
Zr	41	65	46	49	15	42	20	36	33	178	57	150	281
Nb	3	3	2	3	0	3	1	3	2	9	7	9	8
Mo	0	0	0	0	0	0	0	0	0	0	0	0	0
Ag	0	0	0	0	0	0	0	0	0	0	0	0	0
In	0	0	0	0	0	0	0	0	0	0	0	0	0
Sn	0	0	0	3	0	0	0	0	0	0	0	0	0
Sb	0	0	0	0	0	0	0	0	0	0	0	0	0
Cs	0	0	0.6	0	0	0	0	0	0	0.5	0	1.2	0.8
Ba	37	11	154	123	0	28	17	87	22	510	88	335	541
La	6	3.7	6.8	8.1	0.4	7.3	2.2	7.8	4.8	19.4	9.2	71.9	21.3
Ce	15.5	9.2	13	18.2	1.7	19.3	4.9	13.5	12.1	31.3	20.8	133	33.5

Table 1 continued

Mafic granulites (Gt-cpx-hb-opx)														Orthopyroxenites				Felsic granulites									
Sample Lake	03-6B		03-88A		03-159A		03-168E		03-228A		03-92B		03-125A		03-158A		03-189A		03-217A		03-20C		03-182A		03-175A		
	Merret	Bompas	Bompas	Lytle	Lytle	Wasend	Wasend	Bompas	Lytle	Lytle	Lytle	Lytle	Lytle	Lytle	Lytle	Lytle	Lytle	Cora	Cora	Wasend	Wasend	Merret	Merret	Cora	Cora	Cora	Cora
Pr	2.21	1.36	1.36	1.53	1.53	0.33	0.33	2.76	0.68	1.52	1.52	1.65	2.79	3.14	14.6	3.27											
Nd	9.8	6.8	6.8	6.3	6.3	2.2	2.2	12.7	3.2	5.8	5.8	7.3	12.2	10.8	52.8	10.8											
Sm	2.5	2.1	2.1	1.4	1.4	0.8	0.8	3	0.8	1.3	1.3	1.8	2.6	1.7	8.6	1.9											
Eu	0.74	0.89	0.89	0.56	0.56	0.38	0.38	0.71	0.32	0.44	0.44	0.66	0.9	1.51	1.34	1.09											
Gd	2.8	3	3	1.7	1.7	1.3	1.3	3.1	1	1.5	1.5	2.2	2.5	2.5	6.6	2.6											
Tb	0.5	0.6	0.6	0.3	0.3	0.3	0.3	0.5	0.2	0.3	0.3	0.4	0.4	0.6	0.9	0.6											
Dy	3.5	3.8	3.8	1.8	1.8	1.9	1.9	2.8	0.9	1.7	1.7	2.4	2	5.1	5.2	4.7											
Ho	0.8	0.8	0.8	0.4	0.4	0.4	0.4	0.5	0.2	0.3	0.3	0.5	0.4	1.3	1.1	1.1											
Er	2.5	2.6	2.6	1.1	1.1	1.3	1.3	1.5	0.5	1	1	1.4	1	4.1	3.6	3.5											
Tm	0.39	0.38	0.38	0.17	0.17	0.21	0.21	0.21	0.07	0.15	0.15	0.19	0.13	0.59	0.54	0.53											
Yb	2.5	2.4	2.4	1.1	1.1	1.3	1.3	1.2	0.4	1	1	1.2	0.8	3.6	3.3	3.5											
Lu	0.4	0.37	0.37	0.15	0.15	0.2	0.2	0.17	0.06	0.15	0.15	0.17	0.1	0.53	0.5	0.51											
Hf	1.3	1.8	1.8	1.3	1.3	0.5	0.5	1.2	0.6	1	1	1.1	1.5	5.3	4.1	8.3											
Ta	0.2	0.2	0.2	0.1	0.1	0	0	0.1	0	0.1	0.1	0.1	0.5	0.4	0.5	0.4											
W	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0											
Tl	0	0	0	0.1	0.1	0	0	0	0	0	0	0	0	0.5	0.3	0.6											
Pb	0	0	0	0	0	0	0	0	0	0	0	0	0	9	5	10											
Bi	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0											
Th	0.4	0.5	0.5	1.4	1.4	0.3	0	0	0.2	1.1	1.1	0.1	0.7	2.5	22.4	3.2											
U	0.1	0.2	0.2	0.4	0.4	0	0	0	0	0.3	0.3	0	0.2	0.5	2.5	0.7											
Recalculation of composition of protolith																											
Qtz	0.045	0.044	0.044	0.044	0.044	0.046	0.046	0.083	0.244	0.165	0.165	0.049	0.071														
Plag	0.524	0.511	0.511	0.602	0.602	0.361	0.361	0.177	0.125	0.265	0.265	0.293	0.177														
Oliv	0.214	0.172	0.172	0.144	0.144	0.178	0.178	0.328	0.532	0.362	0.362	0.248	0.449														
Cpx	0.179	0.243	0.243	0.163	0.163	0.407	0.407	0.378	0.085	0.167	0.167	0.376	0.272														
Or	0.022	0.014	0.014	0.041	0.041	0.001	0.001	0.022	0.008	0.035	0.035	0.024	0.017														
Sp	0.013	0.014	0.014	0.005	0.005	0.007	0.007	0.009	0.005	0.005	0.005	0.010	0.011														
Ap	0.002	0.002	0.002	0.001	0.001	0.000	0.000	0.002	0.001	0.001	0.001	0.001	0.002														
Ternary Normalization																											
Cpx	0.20	0.26	0.26	0.18	0.18	0.43	0.43	0.43	0.11	0.21	0.21	0.41	0.30														
Pl	0.57	0.55	0.55	0.66	0.66	0.38	0.38	0.20	0.17	0.33	0.33	0.32	0.20														
Ol	0.23	0.19	0.19	0.16	0.16	0.19	0.19	0.37	0.72	0.46	0.46	0.27	0.50														

Samples analyzed at the University of Massachusetts, Amherst by XRF for major elements and at Activation Laboratories by ICPMS for trace elements

Fig. 5 REE diagrams for Chipman domain **a** mafic granulites, and **b** felsic granulites. Data for southern domain felsic granulites from Baldwin et al. (2006) are shown in b for reference. Chondrite values from Sun and McDonough (1989)

extracted at 10 kV and analyzed at mass resolution 5000 using a single ETP electron multiplier and magnetic field switching. Each analysis consisted of five scans through the isotopes of interest and took 8 min. The Pb isotopic compositions were measured directly. Pb/U and Pb/Th were determined relative to chips of zircon standards SL13 ($U = 238$ ppm) and FC1 (radiogenic $^{206}\text{Pb}/^{238}\text{U} = 0.1859$) using a power-law relationship between Pb^+/U^+ and UO^+/U^+ (Claoué-Long et al. 1995) and an exponent of 1.55. The standard also provided an estimate of Pb isotopic fractionation ($2.5 \pm 2.0\text{‰/amu}$). Common Pb was corrected using ^{204}Pb assuming the composition of 2.5 Ga average crustal Pb (Cumming and Richards 1975). Refer to Table 3 for analytical results and to Fig. 11 for the ion microprobe zircon concordia diagram.

U–Pb results

U–Pb ID-TIMS results

U–Pb ID-TIMS data were acquired for zircon grains from three representative samples of Grt-bearing mafic granulites and for zircon and monazite crystals from one sample of felsic granulite (Fig. 3). Mafic granulite sample 02M133A was targeted due to its well-preserved reaction textures that also permitted detailed petrological and kinematic study (Mahan et al. 2008). Felsic granulite sample 03-175A was collected from a lens associated with mafic granulite sample 02M133A. The other two mafic granulite samples were selected to obtain a wider spatial distribution of data over the Chipman domain.

Zircon grains from Grt-bearing mafic granulite 02M133A range in size from tens to hundreds of microns. The dominant zircon population is characterized by simple core-overgrowth relationships (Fig. 6). A secondary zircon population is homogeneous and bright in CL (Fig. 11c, z4). Within the dominant population, three general groups can be distinguished based on differences in external morphology and CL-imaged internal structures. The first group is spherical to ovoid, is dominated by zircon grains that dimly luminesce in CL, and has thin bright CL-imaged rims (e.g., Fig. 6, z3). The second group contains rounded to tabular ovoid grains, sector and oscillatory-zoned cores that appear rounded and resorbed, and bright CL-imaged overgrowths of variable thicknesses (up to 100 μm) (e.g., Fig. 6, z5). The third group is characterized by faceted spherical to ovoid grains with sector and oscillatory-zoned cores of variable resorption, and similar sector zoned overgrowths that are not obviously resorbed (up to 200 μm) (e.g., Fig. 6, z6). The three groups of this dominant zircon population are characterized by Th/U of 0.21–0.31 and U concentrations from 55 to 280 ppm. Ten abraded, one unabraded, and one chemically abraded zircon fragments plotted together yielded a linear array with an upper intercept of 2554.5 ± 4.3 Ma and a lower intercept of 1896.0 ± 18 Ma (MSWD = 3.2) (Fig. 6). The three fragments analyzed from the first group of zircon grains, characterized by grains that dimly luminesce in CL and the thinnest rims, yielded the oldest and most concordant analyses. Two zircon grains from the homogeneous, CL-bright secondary population were analyzed by both ion microprobe (results discussed further below) and CA-TIMS; and yield ca. 1.9 Ga dates.

Sample 03-88A is from the interior of a medium-grained, banded Grt-bearing mafic granulite body at Bompas Lake. Zircon grains are spherical to ovoid, are typically <150 μm in diameter, and compose two populations characterized by distinct internal structures visible in CL (Fig. 7). The first population is dominated by

Table 2 U-Pb TMS isotopic data for zircon

Composition				Isotopic ratios						Dates (Ma)										
Wt (μg) ^b	U (ppm)	Pb (ppm)	Th ^c /U	Pb ⁸⁴ / Pb ^c	Pb ^c / (pg)	²⁰⁶ Pb/ ²⁰⁴ Pb	²⁰⁸ Pb/ ²⁰⁶ Pb	²⁰⁶ Pb/ ²³⁸ U	% err ^g	²⁰⁷ Pb/ ²³⁵ U	% err ^g	²⁰⁷ Pb/ ²⁰⁶ Pb	% err ^g	²⁰⁶ Pb/ ²³⁸ U	²⁰⁷ Pb/ ²³⁵ U	²⁰⁷ Pb/ ²⁰⁶ Pb	±	corr. coef.	% disc	
02M133A—mafic granulite																				
z2b	4.7	69.2	35.5	0.270	67.6	2.4	3715.1	0.076	0.475766	(0.10)	10.9369	(0.11)	0.16672	(0.05)	2508.8	2517.8	2525.0	0.8	0.886	0.8
z2c	1.6	56	29	0.280	30.1	1.5	1725.0	0.079	0.472025	(0.53)	10.7924	(0.56)	0.16583	(0.18)	2492.4	2505.4	2515.9	3.0	0.947	1.1
z3a	0.8	281	146	0.302	68.9	1.7	3867.1	0.085	0.478581	(0.11)	11.0740	(0.12)	0.16782	(0.05)	2521.1	2529.4	2536.0	0.8	0.916	0.7
z3b	1.3	175	94	0.309	23.6	5.0	1255.7	0.087	0.482179	(0.12)	11.1899	(0.13)	0.16831	(0.06)	2536.7	2539.1	2540.9	1.0	0.890	0.2
z5b	0.8	194	95	0.255	94.5	0.8	5735.8	0.073	0.463069	(0.14)	10.4225	(0.16)	0.16324	(0.07)	2453.1	2473.1	2489.5	1.1	0.911	1.8
z6a	2.7	107	47	0.266	241.9	0.5	14757.9	0.078	0.415933	(0.06)	8.5012	(0.07)	0.14824	(0.05)	2242.0	2286.1	2325.7	0.8	0.779	4.3
z6b	2.8	154	74	0.287	305.0	0.7	18374.6	0.083	0.448034	(0.06)	9.8231	(0.08)	0.15901	(0.05)	2386.5	2418.3	2445.2	0.8	0.788	2.9
z7a	1.5	165	76	0.272	161.1	0.7	9785.4	0.080	0.431816	(0.06)	9.1418	(0.08)	0.15354	(0.06)	2313.9	2352.3	2385.8	0.9	0.754	3.6
z8a	2.6	141	74	0.311	337.0	0.6	20064.2	0.087	0.483239	(0.08)	11.2654	(0.09)	0.16908	(0.05)	2541.4	2545.3	2548.5	0.9	0.834	0.3
z9a	5.0	55	25	0.214	158.7	0.8	9776.9	0.063	0.434241	(0.08)	9.2587	(0.09)	0.15464	(0.05)	2324.8	2363.9	2397.9	0.8	0.852	3.6
z10a.un	2.7	74	34	0.225	181.2	0.5	11123.9	0.066	0.431690	(0.16)	9.1286	(0.17)	0.15337	(0.07)	2313.3	2351.0	2383.8	1.2	0.912	3.5
z5c.ca	0.9	—	—	—	32.9	1.3	1926.0	0.075	0.475616	(0.22)	10.9456	(0.26)	0.16691	(0.13)	2508.1	2518.5	2526.9	2.2	0.863	0.9
z7.ip.ca	1.5	—	—	—	4.7	1.8	296.9	0.045	0.344838	(0.94)	5.5442	(1.07)	0.11661	(0.43)	1909.9	1907.5	1904.8	7.7	0.917	−0.3
z17.ip.ca	1.0	—	—	—	8.1	0.6	546.1	0.037	0.345826	(0.40)	5.5933	(0.46)	0.11730	(0.22)	1914.6	1915.1	1915.5	3.9	0.884	0.1
03-88A—mafic granulite																				
z1	3.0	152	72	0.24	320.5	0.7	19540.1	0.068	0.448281	(0.06)	9.8189	(0.07)	0.15886	(0.04)	2387.6	2417.9	2443.6	0.7	0.803	2.7
z2	1.0	196	84	0.10	108.7	0.8	6899.8	0.031	0.423106	(0.08)	8.8137	(0.11)	0.15108	(0.07)	2274.6	2318.9	2358.2	1.2	0.763	4.2
z3	3.7	35	17	0.24	112.3	0.6	6830.3	0.068	0.463900	(0.22)	10.5086	(0.24)	0.16429	(0.07)	2456.8	2480.7	2500.3	1.2	0.954	2.1
z4	1.7	105	50	0.39	57.9	1.5	3262.9	0.113	0.433846	(0.13)	9.2026	(0.14)	0.15384	(0.07)	2323.0	2358.4	2389.1	1.1	0.892	3.3
z5	2.1	32	15	0.18	76.7	0.4	4752.1	0.051	0.445591	(0.14)	9.7934	(0.16)	0.15940	(0.07)	2375.6	2415.5	2449.3	1.2	0.902	3.6
z6.un	2.0	61	26	0.35	70.1	0.7	4211.8	0.105	0.399378	(0.09)	7.9212	(0.11)	0.14385	(0.06)	2166.2	2222.1	2274.1	1.0	0.849	5.6
z7.un	1.3	235	109	0.14	145.8	1.0	9107.8	0.041	0.447124	(0.12)	9.8680	(0.14)	0.16007	(0.07)	2382.5	2422.5	2456.4	1.2	0.873	3.6
z8a.un	1.8	198	88	0.08	46.4	3.4	2647.4	0.023	0.431747	(0.08)	9.1651	(0.10)	0.15396	(0.05)	2313.6	2354.6	2390.4	0.8	0.872	3.8
z8c.ca	1.1	—	—	—	40.5	3.7	2284.2	0.025	0.450704	(0.14)	9.9438	(0.15)	0.16001	(0.06)	2398.4	2429.6	2455.8	1.0	0.919	2.8
03-159A—mafic granulite																				
z1a	3.1	662	337	0.25	1473.2	0.7	88957.9	0.071	0.478610	(0.07)	11.0621	(0.09)	0.16763	(0.04)	2521.2	2528.4	2534.1	0.7	0.876	0.6
z1b	1.6	585	303	0.32	354.5	1.4	20191.2	0.089	0.478078	(0.07)	11.0516	(0.08)	0.16766	(0.04)	2518.9	2527.5	2534.4	0.7	0.841	0.7
z2a	2.9	721	362	0.25	1629.9	0.6	98514.2	0.071	0.472077	(0.05)	10.7982	(0.06)	0.16590	(0.04)	2492.7	2505.9	2516.7	0.7	0.771	1.2
z3	0.9	404	201	0.20	438.6	0.4	26799.1	0.058	0.471242	(0.06)	10.7828	(0.08)	0.16595	(0.05)	2489.0	2504.6	2517.2	0.9	0.724	1.4
z4	1.0	446	227	0.28	148.1	1.5	8410.6	0.079	0.473970	(0.16)	10.9013	(0.17)	0.16681	(0.05)	2500.9	2514.7	2525.9	0.9	0.949	1.2
z5.un	0.0	351	173	0.23	275.2	0.6	16708.1	0.067	0.465182	(0.06)	10.6276	(0.08)	0.16570	(0.05)	2462.4	2491.1	2514.6	0.8	0.775	2.5
z6.un	0.7	413	202	0.22	155.0	0.9	9435.6	0.065	0.463449	(0.10)	10.6056	(0.11)	0.16597	(0.05)	2454.8	2489.2	2517.4	0.9	0.867	3.0
z7.un	0.5	395	196	0.22	70.5	1.4	4122.3	0.063	0.467062	(0.11)	10.6150	(0.15)	0.16483	(0.10)	2470.7	2490.0	2505.9	1.7	0.733	1.7

Table 2 continued

Fr ^a	Composition			Isotopic ratios							Dates (Ma)									
	Wt (μg) ^b	U (ppm)	Pb (ppm)	Th ^c /U	Pb ^{*d} /Pbc	Pbc (pg)	²⁰⁶ Pb ^e / ²⁰⁴ Pb	²⁰⁸ Pb ^f / ²⁰⁶ Pb	²⁰⁶ Pb ^g / ²³⁸ U	% err ^g	²⁰⁷ Pb ^h / ²³⁵ U	% err ^g	²⁰⁷ Pb ⁱ / ²⁰⁶ Pb	% err ^g	²⁰⁶ Pb ^h / ²³⁸ U	²⁰⁷ Pb ^h / ²³⁵ U	²⁰⁷ Pb ^h / ²⁰⁶ Pb	±	corr. coef.	% disc
z8.un	1.2	353	184	0.32	350.3	0.6	20668.8	0.092	0.478831	(0.08)	11.6078	(0.09)	0.17582	(0.04)	2522.2	2573.3	2613.8	0.7	0.866	4.2
03-175A—felsic granulite																				
z1	1.7	295	141	0.10	324.6	0.7	20376.6	0.029	0.465360	(0.06)	10.5381	(0.09)	0.16424	(0.07)	2463.2	2483.3	2499.8	1.1	0.701	1.8
z2	3.7	457	222	0.14	261.2	3.1	14540.2	0.039	0.467845	(0.05)	10.7121	(0.07)	0.16606	(0.04)	2474.1	2498.5	2518.3	0.7	0.794	2.1
z3a	1.6	385	187	0.13	91.6	3.2	5109.4	0.036	0.465739	(0.07)	10.6915	(0.08)	0.16649	(0.05)	2464.8	2496.7	2522.7	0.8	0.832	2.8
z3b	1.0	347	169	0.14	108.4	1.6	6363.4	0.039	0.467653	(0.08)	10.6694	(0.10)	0.16547	(0.05)	2473.3	2494.8	2512.3	0.8	0.864	1.9
z4	1.4	402	197	0.13	192.5	1.4	11392.3	0.037	0.472672	(0.08)	10.8537	(0.09)	0.16654	(0.05)	2495.3	2510.7	2523.2	0.8	0.851	1.3
z5.un	1.6	401	188	0.10	289.8	1.0	18105.3	0.030	0.456045	(0.06)	10.2039	(0.07)	0.16228	(0.04)	2422.1	2453.4	2479.5	0.7	0.796	2.8
z6.un	1.3	274	129	0.09	119.6	1.4	7184.8	0.026	0.461441	(0.22)	10.4195	(0.23)	0.16377	(0.06)	2445.9	2472.8	2494.9	0.9	0.969	2.4
z7.un	1.4	286	136	0.13	85.3	2.2	4873.9	0.036	0.456323	(0.08)	10.3088	(0.10)	0.16385	(0.07)	2423.3	2462.9	2495.7	1.2	0.737	3.5
m1b	3.3	125	1,149	75.03	967.8	3.9	2865.8	21.088	0.476742	(0.13)	10.9549	(0.15)	0.16666	(0.08)	2513.1	2519.3	2524.3	1.3	0.857	0.5
m2a	1.5	1,206	4,225	25.54	1794.0	3.5	14259.6	7.129	0.486962	(0.05)	11.4200	(0.07)	0.17009	(0.04)	2557.5	2558.1	2558.5	0.7	0.791	0.0
m2b	3.6	1,104	3,744	24.59	3121.2	4.3	25415.2	6.862	0.487153	(0.12)	11.4255	(0.13)	0.17010	(0.04)	2558.3	2558.5	2558.6	0.7	0.939	0.0
m3b	0.0	357	169	0.51	64.4	2.3	3474.2	0.144	0.422905	(0.13)	8.4017	(0.16)	0.14409	(0.09)	2273.7	2275.4	2276.9	1.6	0.817	0.2
m4a	0.9	3,185	6,998	14.56	1322.9	4.8	16542.8	4.084	0.482881	(0.08)	11.2629	(0.09)	0.16916	(0.05)	2539.8	2545.1	2549.4	0.8	0.833	0.5
m5a	2.0	1067	624	1.57	253.7	4.9	10391.1	0.461	0.420338	(0.06)	8.6215	(0.07)	0.14876	(0.04)	2262.0	2298.8	2331.7	0.7	0.805	3.5
m5b	1.1	2744	1,583	1.42	440.6	3.9	18735.9	0.411	0.427842	(0.13)	8.8702	(0.14)	0.15037	(0.04)	2296.0	2324.7	2350.1	0.7	0.952	2.7
m6a	1.8	2228	2,284	5.21	1633.3	2.5	42030.1	1.474	0.451027	(0.05)	9.6772	(0.06)	0.15561	(0.04)	2399.8	2404.5	2408.5	0.7	0.768	0.4

^a All fractions are single grains or grain fragments. Fragments are denoted with a letter after the fraction number. All zircon fractions were physically abraded, except for those marked *un* unabraded, or *ca* chemical abrasion. *ip* indicates grains first analyzed by ion probe prior to TIMS analysis. No monazite fractions were physically abraded

^b Sample weights were estimated to within 40% using measured grain dimensions and a nominal density of 4.5 g/cm³ for zircon and 5.0 g/cm³ for monazite

^c Th contents calculated from radiogenic ²⁰⁸Pb and the ²⁰⁷Pb/²⁰⁶Pb date of the sample, assuming concordance between U–Th and Pb systems

^d Pb* and Pbc represent radiogenic Pb and common Pb, respectively

^e Measured ratio corrected for fractionation and spike contribution; Pb fractionation was 0.07 ± 0.04‰/a.m.u. for Faraday detector or 0.25 ± 0.04‰/a.m.u. for Daly detector analysis, based on daily analysis of NBS-981

^f Measured ratios corrected for fractionation, spike, blank, and initial common Pb; nominal U blank = 0.1 pg ± 50% (2 s); nominal Pb blank = 1.0 pg ± 50% (2 s); initial Pb composition estimated using the model for terrestrial Pb evolution of Stacey and Kramers (1975) at the nominal age of the fraction (i.e., 2.55 Ga)

^g Numbers in parentheses are the % errors reported at the 2 sigma confidence interval, propagated using the algorithms of Ludwig (1980)

^h Isotopic ages calculated using the decay constants of Jaffey et al. (1971): $\lambda(^{235}\text{U}) = 9.8485 \times 10^{-10} \text{ year}^{-1}$ and $\lambda(^{238}\text{U}) = 1.55125 \times 10^{-10} \text{ year}^{-1}$; uncertainty in ²⁰⁷Pb/²⁰⁶Pb date reported at the 2 sigma confidence interval

Fig. 6 U–Pb concordia diagram for zircon analyzed by TIMS from mafic granulite sample 02M133A. CL images of analyzed zircon crystals are shown, except for analyses z7.ip.ca and z17.ip.ca

zircon grains that are homogeneous and bright in CL, some with 5–20 μm darker CL overgrowths (e.g., Fig. 7, z7). The second, more abundant type of zircon grain is characterized by an oscillatory to irregularly zoned internal domain, irregular embayments on some grains suggesting possible resorption, and overgrowths of variable thickness (e.g., Fig. 7, z4). All analyzed zircon grains have U concentrations from 30 to 235 ppm. No dramatic difference in U–Pb systematics was detected between the two populations distinguished in CL. Five abraded single zircon grains, two unabraded single zircon grains, one unabraded zircon fragment, and one chemically abraded zircon fragment yielded a linear array with an upper intercept of ca. 2.56 Ga and a lower intercept of ca. 1.86 Ga (Fig. 7).

Fig. 7 U–Pb concordia diagram for zircon analyzed by TIMS from mafic granulite sample 03-88A. CL images of analyzed zircon crystals are shown

Sample 03-159A was collected from a ~50-m wide layer of Grt-bearing mafic granulite at Lytle Lake. Zircon grains are spherical to ovoid, range in size from tens to hundreds of microns, do not luminesce in CL, and have U concentrations of 340–690 ppm (Fig. 8). Some grains display faint internal sector zoning in BE, and possible overgrowths. Three abraded zircon fragments, two abraded single zircon grains, and four unabraded single zircon grains were analyzed. The five abraded fractions and one unabraded fraction yielded a poorly defined array with a ca. 2.55 Ga upper intercept and a Proterozoic lower intercept (Fig. 8). One significantly older analysis is interpreted to contain a minor inherited component. Although the unabraded zircon fractions were intended to better define the

Fig. 8 U–Pb concordia diagram for zircon analyzed by TIMS from mafic granulite sample 03-159A. BE images of some analyzed zircon crystals are shown. Shown for reference is the discordia line from 02M133A

Fig. 9 U–Pb concordia diagram for zircon analyzed by TIMS from felsic granulite sample 03-175A. BE images of analyzed zircon crystals are shown. Shown for reference is the discordia line from 02M133A

lower intercept of the array by retaining the zircon overgrowths, these fractions show greater discordance. This increased discordance is likely caused by post -1.9 Ga, low temperature Pb loss in the radiation damaged rims.

Sample 03-175A is a partially retrogressed felsic granulite from Cora Lake containing cm-scale garnet. Zircon grains are abundant, spherical to ovoid, typically <100 μm (up to 150 μm in length), and do not luminesce in CL (Fig. 9). In BE, zircon grains show faint internal concentric zoning, with possible thin <10 μm rims on some crystals. Two abraded fragments, three abraded single zircon grains, and three unabraded single zircon grains were analyzed. Five fractions yield a poorly defined array with an upper intercept at ca. 2.5 Ga and a Proterozoic lower intercept (Fig. 9).

Felsic granulite sample 03-175A also contains abundant monazite. Eight monazite fragments were analyzed from

six different monazite crystals (Fig. 10). Imaged monazite grains show broad compositional variation, consistent with previous in situ imaging of monazite from Chipman domain felsic granulites (Mahan et al. 2006a). Populations include grains that display concentric, patchy and homogeneous zonation. Monazite grains vary widely in Th, Ca, Y and U concentrations, and are characterized by a broad range (0.5–75) of Th/U. Analyses ranged from concordant to several percent discordant, with $^{207}\text{Pb}/^{206}\text{Pb}$ dates from 2.28 to 2.56 Ga, indicating multiple age populations (Fig. 10). The discordance of some fractions is likely due to analysis of multiple age domains in some fragments.

U–Pb ion microprobe analytical results

Ion microprobe spot data were acquired to help decipher the significance of the mixing lines defined by the zircon

Fig. 10 U–Pb concordia diagram for monazite analyzed by TIMS from felsic granulite sample 03-175A. Associated Th and Y compositional maps are shown for several analyzed crystals. Higher concentrations of each element are in yellow, and lower concentrations are in red.

TIMS analyses. We targeted lower U zircon grains from mafic granulite sample 02M133A because these crystals yielded the best-defined TIMS mixing array with little scatter, suggesting minimal secondary effects from post-1.9 Ga low-temperature Pb loss. In addition, detailed thermobarometric study, petrogenetic modeling and kinematic analysis of this same sample (Mahan et al. 2008), provided a particularly well-documented context for interpretation of the geochronological results. Thirty analyses on 26 zircon crystals from the dominant population yielded a range of apparent $^{207}\text{Pb}/^{206}\text{Pb}$ dates, including thirteen analyses on zircon cores from $2,674 \pm 15$ to $2,065 \pm 53$ Ma, twelve analyses on zircon overgrowths from $2,710 \pm 27$ to $1,911 \pm 40$ Ma, and five analyses on grains without overgrowths from $2,600 \pm 19$ to $2,161 \pm 45$ Ma (Fig. 11a). Zircon cores and overgrowths

show a similar range of Th/U from 0.15 to 0.42. Eight analyses on four unzoned to weakly zoned zircon grains that are CL-bright yielded dates from $1,805 \pm 88$ to $1,949 \pm 70$ Ma. This subset of ca. 1.9 Ga homogeneous grains displays the lowest Th/U (0.08–0.14) and the lowest U and Th concentrations of the dataset.

Discussion

Zircon growth and recrystallization

U–Pb TIMS results for lower U (<300 ppm) mafic granulite zircons yielded discordia arrays with intercepts at ca. 2.55 and 1.9–1.8 Ga. These lower U zircons luminesced in CL and were characterized by sector zoning with isometric overgrowths typical of crystallization at amphibolite through granulite facies conditions (Vavra et al. 1996, 1999; Corfu et al. 2003) (Figs. 6, 7). Higher U zircon crystals (>350 ppm, 03-175A, 03-159A) did not luminesce in CL and yielded more scattered discordia arrays than the low U grains (Figs. 8, 9). Suppression of zircon CL is attributed to elevated levels of accumulated radiation damage in the higher U grains (e.g., Nasdala et al. 2002). The greater data scatter is presumably caused by low temperature Pb loss in more highly radiation damaged zircon crystals as widely documented in prior studies (e.g., Silver and Deutsch 1963; Cherniak et al. 1991). We infer this radiation damage accumulated post-1.9 Ga, because radiation damage in zircon is annealed at elevated temperatures (e.g., Mattinson 2005). Therefore, the radiation damage in the mafic granulite zircons would have been continuously annealed at the temperature conditions prevailing in the lower crust prior to subsequent cooling and unroofing.

Ion microprobe dates for zircon cores and overgrowths of the dominant population in mafic granulite sample 02M133A are inconsistent with simple mixing of cores and overgrowths of different age. No correlation of date with CL zonation is observed. Cores and overgrowths are characterized by relatively uniform Th/U. The analyses yielded $^{207}\text{Pb}/^{206}\text{Pb}$ dates from 2.71 to 1.91 Ga. The dataset includes a group of concordant U–Pb ca. 2.55 Ga dates, as well as reversely discordant analyses. It is unclear whether the reversely discordant dates represent unsupported radiogenic Pb as reported in rare cases due to local Pb gain (e.g., Williams et al. 1984), or are analytical artifacts (e.g., Wiedenbeck 1995). In contrast to the complex U–Pb zircon systematics of this dominant population, the secondary population of bright CL zircon is characterized by distinctly lower Th/U, lower U, and consistent ca. 1.9 Ga dates.

The secondary population of low U zircon crystals is best explained by neoblastic growth at ca. 1.9 Ga. Any

Grain ^a			Composition			Apparent ages (Ma)																
Spot	Type ^b	Pb ^c (ppm)	U (ppm)	Th (ppm)	Th/U (ppm)	²⁰⁴ Pb/ ²⁰⁶ Pb	%±	F206%	²⁰⁶ Pb/ ²³⁸ U	%±	²⁰⁷ Pb/ ²³⁵ U	%±	²⁰⁷ Pb/ ²⁰⁶ Pb	%±	208/232	±	206/238	±	207/235	±	207/206	±
Zircon cores																						
28.2	MWzC	26	45	9.3	0.21	2.71E-05	83	0.038	0.5500	2.1	13.83	2.4	0.1823	0.9	3002	127	2825	49	2738	23	2674	15
5.2	DÜzC	116	204	60	0.30	4.35E-06	98	0.006	0.5236	1.1	12.91	1.3	0.1789	0.5	2627	43	2714	25	2673	12	2642	8
26.1	DÜzC	700	1370	254	0.19	4.31E-06	36	0.006	0.4861	0.4	11.56	0.6	0.1724	0.3	2473	21	2554	9	2569	5	2581	6
21.1	MZC	15	27	5.8	0.21	6.66E-05	58	0.094	0.5174	1.8	12.22	2.3	0.1713	1.2	2539	121	2688	40	2622	22	2571	20
15.2	DÜzC	97	188	55	0.30	1.58E-05	58	0.022	0.4806	1.4	11.22	1.6	0.1693	0.7	2446	50	2530	29	2542	15	2551	12
30.1	DÜzC	91	174	46	0.26	2.67E-05	39	0.038	0.4919	1.2	11.47	1.6	0.1691	0.9	2576	52	2579	25	2562	15	2549	16
20.1	LWzC	9.0	17	4.9	0.30	4.61E-05	90	0.065	0.4877	2.3	11.32	3.1	0.1684	1.8	2605	139	2561	49	2550	30	2542	31
8.1	DZC	62	118	49	0.42	3.26E-05	63	0.046	0.4739	0.9	10.85	1.2	0.1660	0.7	2449	55	2501	19	2510	11	2517	11
9.1	DÜz	81	168	38	0.23	1.21E-05	65	0.017	0.4568	0.8	10.12	1.1	0.1607	0.6	2397	41	2426	17	2446	10	2463	10
6.2	DWzC	40	89	21	0.24	3.00E-06	146	0.004	0.4342	1.2	9.17	1.8	0.1531	1.2	2303	60	2325	24	2355	17	2381	20
12.1	LÜzC	12	28	6.3	0.23	3.88E-05	116	0.055	0.4175	3.4	8.28	4.3	0.1439	2.3	2264	178	2249	65	2262	40	2274	39
14.1	MWzC	18	46	10	0.23	1.21E-05	91	0.017	0.3913	1.9	7.62	3.1	0.1412	2.2	2040	88	2129	34	2187	28	2242	39
13.1	LÜzC	8.0	21	5.1	0.25	2.35E-04	45	0.360	0.3745	4.1	6.59	5.4	0.1276	3.0	1912	190	2051	73	2058	49	2065	53
Zircon overgrowths																						
25.1	LWzO	20	33	6.1	0.19	3.16E-05	141	0.045	0.5707	1.8	14.67	2.6	0.1864	1.6	3040	136	2911	42	2794	25	2710	27
15.1	LWzO	16	26	5.6	0.22	1.29E-04	55	0.181	0.5651	2.4	14.39	3.9	0.1847	2.8	2704	201	2888	56	2776	38	2696	47
22.1	LWzO	23	43	11	0.26	2.40E-05	156	0.034	0.5139	2.6	12.13	3.3	0.1711	1.7	2688	146	2673	58	2614	32	2569	29
2.1	MÜzO	35	67	14	0.21	5.02E-06	92	0.007	0.5028	1.3	11.79	2.0	0.1701	1.3	2653	67	2626	29	2588	19	2558	22
24.1	LÜzO	19	37	9.0	0.24	9.85E-06	132	0.014	0.4861	2.9	11.25	3.5	0.1678	1.7	2504	108	2554	61	2544	33	2536	29
23.1	DÜzO	296	554	129	0.23	9.06E-06	73	0.013	0.5038	1.1	11.64	1.3										

Table 3 continued

Grain ^a spot	Composition		Apparent ages (Ma)		Apparent ages (Ma)		Apparent ages (Ma)		Apparent ages (Ma)		Apparent ages (Ma)	
	Type ^b	Pb ^c (ppm)	U (ppm)	Th (ppm)	Th/U	$^{204}\text{Pb}/^{206}\text{Pb}$	% $\pm$	F206%	$^{206}\text{Pb}/^{238}\text{U}$	% $\pm$	$^{207}\text{Pb}/^{235}\text{U}$	% $\pm$
Zircon grains bright in CL												
18.1	LWz	3.0	7.6	0.6	0.08	2.35E-04	74	0.361	0.3352	4.7	5.52	6.4
												0.1195
18.2	LWz	3.0	8.7	0.7	0.08	1.38E-04	87	0.212	0.3428	4.5	5.33	7.3
												0.1128
7.1	LWz	4.0	12	1.5	0.12	1.30E-04	93	0.200	0.3537	4.3	5.52	5.8
												0.1132
7.2	LWz	4.0	11	1.3	0.12	2.45E-05	101	0.038	0.3408	5.5	5.23	6.9
												0.1113
4.1	LWz	4.0	12	1.1	0.09	6.35E-04	39	0.973	0.3456	3.3	5.26	6.0
												0.1103
4.2	LWz	3.0	10	1.0	0.09	2.78E-04	56	0.426	0.3371	4.2	5.26	6.8
												0.1132
17.1	LUz	4.0	11	1.6	0.14	2.25E-04	72	0.345	0.3422	4.5	5.28	6.1
												0.1119
17.2	LUz	4.0	11	1.4	0.12	7.16E-05	100	0.110	0.3345	4.8	5.24	5.8
												0.1137

Errors are given at the 1-sigma level

^a Refers to individual grain followed by analysis number^b L, M, D CL (brightness): light, medium, dark; Uz, Wz, Z (zoning: unzoned), weakly zoned, well zoned; C, O (location): core, overgrowth^c Pb* Radiogenic lead

Fig. 11 **a** U–Pb concordia diagram for zircon analyzed by ion microprobe from mafic granulite sample 02M133A with data plotted as 1-sigma error ellipses. **b** U–Pb concordia diagram showing both TIMS data plotted as 2-sigma error ellipses with discordia line, and the ion microprobe data plotted as 1-sigma error ellipses. **c** CL images of representative zircon crystals with locations of ion microprobe spots are shown. The $^{207}\text{Pb}/^{206}\text{Pb}$ date and 1-sigma error for each spot analysis is indicated

explanation for the U–Pb systematics of the dominant population, however, must account for the absence of correlation between date and CL zonation. Episodic zircon growth between 2.55 and 1.9 Ga does not appear to be a viable interpretation, because a similar range of dates and Th/U are documented for cores and overgrowths. Simple diffusional Pb loss at 1.9 Ga appears improbable because of the extreme sluggishness of Pb diffusion in zircon documented experimentally (Lee et al. 1997; Cherniak and Watson 2000), and because this mechanism cannot easily explain the heterogeneous spatial distribution of dates unless Pb loss primarily occurred via microfractures that were subsequently annealed. Our preferred interpretation is that both cores and overgrowths of the dominant population grew at ca. 2.55 Ga with homogeneous Th/U, and were heterogeneously and incompletely recrystallized at ca. 1.9 Ga. The well-defined TIMS discordia array with a lower intercept at 1.9 Ga supports a single phase of zircon recrystallization at this time, rather than multiple recrystallization episodes between 2.55 and 1.9 Ga. Datasets similar to ours are rare in the published literature. Two other zircon studies in different granulite localities reported a range of ion probe dates that do not correlate with CL zonation (Percival et al. 2003; McFarlane et al. 2004), analogous to what we observe in our dataset. In contrast, many magmatic and metamorphic zircon crystals that underwent isotopic resetting by recrystallization are characterized by unusual patterns of dates that coincide with bleached zones and transgressive replacement fronts visible in CL, changed Th/U, and modified trace element compositions. These characteristics are generally attributed to high temperature metamorphic processes, metasomatic alteration, and/or the instability of impurity-rich zircon domains (e.g., Williams et al. 1984; Pidgeon 1992; Pidgeon et al. 1998; Schaltegger et al. 1999; Hoskin and Black 2000; Carson et al. 2002). Our results deviate from these typical zircon recrystallization patterns because our metamorphic zircon crystals show no textural recrystallization in CL, and are characterized by consistent Th/U.

We suggest that our pattern of discordant zircon that define a mixing array between 2.55 and 1.9 Ga may be the signature of the two granulite facies metamorphic events in the East Lake Athabasca region. A similar TIMS array was observed in mafic granulite zircon from the southern domain (Baldwin et al. 2003), and we also anticipate a similar pattern of ion microprobe data for these crystals. The conventional TIMS and ion microprobe data offer complementary information in our study. The ion microprobe analyses provide the spatial resolution to document the lack of correlation between age, composition and location within individual crystals. However, the large errors associated with individual analyses do not allow discrimination between concordant and discordant points,

and could be interpreted to indicate either continuous zircon growth or two phase mixing. The more precise TIMS results clearly define a discordia array with two intercepts, and result in our explanation of the data by the two component mixing model (Fig. 11b).

Monazite growth and recrystallization

Imaging, compositional mapping, and U–Pb TIMS data for monazite grains separated from the felsic granulite indicate multiple populations that yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 2.56 and 2.33 Ga. Detailed textural observations, petrological analysis, in situ monazite compositional maps and electron microprobe monazite geochronology on anhydrous felsic granulites from the Chipman domain documented four compositionally distinct monazite populations (Mahan et al. 2006a). These authors interpreted their dataset to record metamorphic growth of two relatively high Y monazite populations from 2.55 to 2.5 Ga during what was likely a single metamorphic cycle, episodic growth of low Y monazite between 2.5 and 2.15 Ga, and crystallization of low Y monazite during 1.9 Ga granulite facies metamorphism. The composition and dates for monazite analyzed by TIMS in our study correlate well with the previously identified monazite populations. For example, M2 and M4 have high Y and Th (brighter in the simultaneously processed maps in Fig. 10b), high Th/U, and give concordant or near concordant $^{207}\text{Pb}/^{206}\text{Pb}$ dates of 2,549–2,558 Ma, all of which are consistent with population 1 of Mahan et al. (2006a). The monazite grains that yielded Paleoproterozoic TIMS dates likely grew by fluid-mediated or recrystallization processes, and belong to the third population described above. The propensity of monazite to grow significantly below its closure temperature to Pb diffusion is a well-known characteristic of this mineral (e.g., Rubatto et al. 2001).

Repeated granulite facies metamorphic events in the lower crust

Mafic and felsic granulites underwent zircon growth at 2.55 Ga (dominant population) and 1.9 Ga (secondary population), with recrystallization of zircon inferred at 1.9 Ga to explain the unusual U–Pb systematics of the dominant population in sample 02M133A. A simultaneous metamorphic study of this sample identified 1) a M_1 $\text{Grt}_1 + \text{Cpx}_1 + \text{Qtz}$ assemblage (~ 1.3 GPa, 850°C) inferred to have developed from an Opx-bearing igneous assemblage, 2) a $\text{Grt}_2 + \text{Hbl}_2$ -bearing M_2 matrix assemblage interpreted to represent subsequent retrogression of the M_1 assemblage, and 3) a second prograde M_3 assemblage of $\text{Opx}_3 + \text{Cpx}_3$ (~ 1.0 GPa, $>800^\circ\text{C}$) (Mahan et al. 2008). The geochronologic and metamorphic data appear

best explained by growth of the dominant population of ca. 2.55 Ga metamorphic zircon during the first high-P event preserved in the mafic granulite, followed by growth and recrystallization of metamorphic zircon at 1.9 Ga during development of the second high-P assemblage.

These data support prior studies that indicated two phases of high-grade metamorphism in the Chipman domain. Archean metamorphism was interpreted from the observation that garnet in felsic granulite exclusively contains ca. 2.55 Ga monazite, indicating Archean growth of garnet (Mahan et al. 2006a). Until recently, however quantitative P–T estimates for this metamorphic event were elusive (Mahan et al. 2008). The timing of the second high-grade metamorphic phase in the Chipman domain was previously constrained at 1,896 Ma, based on U–Pb dates for metamorphic zircon grains from migmatized Chipman mafic granulite dikes that record conditions of 1.0 GPa, >750°C (Williams et al. 1995; Flowers et al. 2006b). These P–T estimates are consistent with those for the second prograde assemblage in the mafic granulite gneiss (1.0 GPa, >800°C). Titanite grains from mafic granulite sample 02M133A yielded dates of ca. 1.89 Ga, compatible with ca. 1.88–1.90 Ga titanite dates for five other samples across the Chipman domain. From petrologic modeling, Mahan et al. (2008) constrain titanite growth to have occurred during the M₂ retrograde period prior to reheating to peak M₃ conditions. Therefore, the titanite dates are interpreted to reflect diffusional Pb loss during attainment of 1.9 Ga temperatures >550–650°C (above the U–Pb titanite closure temperature, e.g., Hanson et al. 1971; Cherniak 1993; Flowers et al. 2006c). Collectively, these data support two episodes of high-P granulite facies metamorphism separated by 650 million years.

Long-term lower crustal residence

The two discrete episodes of lower crustal metamorphism raise the question of the history in the intervening 650 m.y. and would suggest either that the rocks 1) resided and cooled in the lower crust for 650 million years between the Archean and Proterozoic high-P granulite events, or 2) underwent post-2.55 Ga exhumation, were transported a second time to lower crustal depths for the 1.9 Ga granulite event, and then underwent final unroofing. The former history is simpler than one requiring a second independent cycle of exhumation and emplacement/transport to lower crustal depths, while still preserving the Archean peak metamorphic assemblage.

Two lines of evidence in the Chipman domain support the protracted lower crustal residence hypothesis. First, there is little signature of activity in the 2.50–1.9 Ga time interval, with the exception of fluid-mediated monazite growth in the felsic granulite. Deformation and the

production of Hbl in the M₂ assemblage of the mafic granulite is interpreted to be synchronous with decompression from ~1.3 GPa (M₁) to ~1.0 GPa (M₂) in the 2.55 to 2.50 Ga interval (Mahan et al. 2008). Second, the limited melting of Chipman domain felsic tonalite gneisses during 1.9 Ga high-temperature metamorphism is consistent with an already dehydrated state at this time (Flowers et al. 2006b), implying little post-2.55 Ga rehydration that presumably would be induced by significant unroofing and reburial. In addition, in the adjacent northwestern domain, isobaric cooling of the 2.6 Ga Mary granite following emplacement and metamorphism has been used to support crustal residence (Williams et al. 2000; Williams and Hanmer 2005), although the duration of this isobaric cooling phase is not constrained. We also note that two episodes of 2.55 and 1.9 Ga metamorphic zircon growth are reported for mafic rocks in the southern domain (Baldwin et al. 2004), which we similarly suggest may bracket an interval during which these rocks remained deeply buried.

Such a history of extended lower crustal residence is common; it is similar to that inferred for most lower crustal xenoliths erupted within stable continental interiors, and logically is the history of the lower crust in most cratonic regions. Calculation of cratonic conductive geotherms indicates that the temperatures at the base of the crust may be as cool as 400°C during periods of cratonic stability. Consistent with this idea, preservation of Archean metamorphic assemblages in lower crustal xenoliths erupted in Phanerozoic kimberlites demonstrates that these steady-state temperatures are typically insufficient to re-equilibrate the peak mineral assemblages. Furthermore, this would be compatible with dry conditions, like those inferred for the Chipman domain, because rehydration leads to mineral growth and recrystallization (e.g., Schaltegger et al. 1999). In the Chipman domain, the apparent preservation of Archean metamorphic assemblages despite subjection to high-temperatures (>750°C) at 1.9 Ga supports a relatively brief interval at the metamorphic peak. This is consistent with U–Pb zircon constraints for the Chipman dikes that suggest the time span for dike migmatization at exposed crustal levels was ≤2.5 million years, and with cross-cutting pegmatitic veins that indicate significant deformation terminated within ~3 million years of mafic dike anatexis (Flowers et al. 2006b).

Stabilization and reactivation of cratonic lithosphere

The unusual signature of two lower crustal high-temperature metamorphic events, widely separated in time, is consistent with the stabilization and subsequent reactivation of cratonic lithosphere in this portion of the western Canadian shield. The 2.55-Ga phase of lower crustal granulite facies metamorphism documented in this study

followed widespread 2.6 Ga magmatism, and suggests initial stabilization of this portion of the craton at this time. We infer that the deep crustal isobaric cooling and hiatus in significant lower crustal tectonism indicate cooling to a cratonic geotherm and the presence of a protective cratonic root during the following 650 million years. The 1.9-Ga high-P granulite facies event would then record the reactivation of the craton during the amalgamation of the Laurentian supercontinent ca. 2.02–1.78 Ga. In other Archean cratons, lower crustal xenoliths commonly record protracted histories of thermal and magmatic perturbations, but the dramatic aspect of the record in the East Lake Athabasca region is that this subsequent episode of lithospheric disruption was sufficiently severe to initiate exhumation of the lower crust to the surface.

The reactivation of cratonic lithosphere in this portion of the western Canadian shield was likely driven by its position between two major convergent orogens; the 1.91–2.02-Ga Taltson-Thelon orogen to the northwest and the 1.78–1.88-Ga Trans-Hudson orogen to the southeast. Significant activity in the southern Taltson orogen included intrusion of voluminous magmatic rocks from 1.99 to 1.92 Ga, granulite to upper amphibolite facies shearing from 1.938 to 1.934 Ga, and subsequent widespread cooling and amphibolite to greenschist grade shearing between 1.9 and 1.8 Ma (McDonough et al. 2000). Recent models for the evolution of the Trans-Hudson orogenic system propose that the Hearne still retained a passive margin during initial ocean closure and arc formation from 1.92 to 1.88 Ga, followed by significant accretionary activity from 1.88 to 1.835 Ga and the terminal collisional phase from 1.835 to 1.78 Ga (Corrigan et al. 2005). The 1.9-Ga metamorphic peak in the Chipman and adjacent domains of the Athabasca mylonite triangle temporally falls between the peak activity of the bounding orogens, reflecting an overall progression of west to east tectonic activity.

Flowers et al. (2006b) suggested that a continent-wide corridor (>1,200 km of strike-length) of short-lived asthenospheric upwelling and incipient rifting were associated with the 1.9 Ga event, with upwelling induced as a convective response to the downgoing slabs in the adjacent orogens, or by hinterland transtension due to the strike-slip dominated regime of the Taltson-Thelon system. The ca. 1.9-Ga magmatism was manifested locally in the central segment as the Chipman mafic dike swarm, further north as the emplacement of the Kazan dikes in the Tulemalu fault zone, and in the northern segment as the emplacement of the enormous Kramanituvar, Uvauk, and Daly Bay mafic complexes (Fahrig et al. 1984; Mills et al. 2000; Sanborn-Barrie et al. 2001; Hanmer and Williams 2001). In addition, the southern extension of the Snowbird zone beneath Phanerozoic cover has been interpreted to record the opening and subsequent closure of a marginal Proterozoic

ocean basin (Ross et al. 1995; 2000). Linkage of these southern observations with the nearly synchronous 1.9 Ga mafic magmatism along much of the remainder of the Snowbird zone would extend the correlation of extensional tectonism and magmatism along the entire 2,800 km length of the geophysical feature. Accretionary activity in the Trans-Hudson orogen may have driven a later transition to hinterland contraction, terminating mafic magmatism and driving basin closure and collision to the south. We speculate that this sequence of events occurred without wholesale removal of the Archean cratonic root, but mantle xenolith studies to constrain the age of lithosphere beneath this region would be required to confirm this hypothesis. Others have argued that the Snowbird tectonic zone may represent a 1.9-Ga collisional orogen between the Rae and Hearne cratons, possibly along its entire length (Hoffman 1988; Baldwin et al. 2003; Berman et al. 2007). We note that whatever model is used to explain this fundamental feature must also explain the distribution of 1.9 Ga mafic magmas spanning >1,200 km of its extent.

Detailed structural, metamorphic and thermochronologic work reconstructed a protracted multiphase unroofing history in the 200-million years following the 1.9 Ga phase of granulite facies metamorphism (Flowers et al. 2006a, 2006c; Mahan et al. 2006a, 2006b). These studies deciphered extensional unroofing associated with metamorphism and mafic magmatism from ~1.0 to 0.8 GPa at ca. 1.90 to 1.89 Ga, subsequent regional unroofing due to contractional uplift along the Legs Lake shear zone from ~0.8 to 0.5 GPa at ca. 1.85 Ga, and a final interval of extensional unroofing that transported the exposures to near-surface conditions by ca. 1.7 Ga. The multistage unroofing history was attributed to the changing contractional and extensional regimes across the Laurentian supercontinent during the terminal phases of collision along the Taltson-Thelon and Trans-Hudson orogenic systems (Flowers et al. 2006c; Mahan et al. 2006b).

The post-1.9 Ga history is somewhat paradoxical with evidence for a short-lived event that perhaps changed the strength profile of the lithosphere sufficiently to predisposed it for deformation driven largely by far field effects from adjacent orogens. This is consistent with ideas regarding activity along intracratonic deformation zones due to thermal weakening (e.g., Celerier et al. 2005). The preservation of the exhumed lower crust of the East Lake Athabasca region indicates that this region ultimately re-attained a stable lithospheric configuration, and has persisted as a stable craton for the last 1.7 billion years.

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