GEOLOGICAL SURVEY OF CANADA
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Metamorphic map of the western Churchill Province, Canada

R. G. Berman

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This report contains marginal notes to accompany 3 map sheets.

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INTRODUCTION

1. Overview:

The western Churchill Province represents a large region of dominantly Meso- to Neo-Archean continental crust comprised of at least four major crustal blocks: Rae craton, Hearne craton, Chesterfield block, and Meta Incognita microcontinent (Fig. 1). Recent tectonic interpretation (Berman et al., 2007) suggests that the Rae and Chesterfield blocks were amalgamated at ca. 2.65 Ga (prior to the emplacement of ca. 2.6 Ga stiching plutons), whereas the Hearne craton accreted to the composite Rae-Chesterfield block at ca. 1.9 Ga, prior to the ca. 1.88 – 1.865 Ga collision of Meta Incognita microcontinent with the Rae craton (St-Onge et al., 2002).

The western Churchill Province has a complex metamorphic history with some regions affected by more than eight thermotectonic events between ca. 2.8 – 1.7 Ga. Not surprisingly, the most recent of these events is the most extensively preserved and the most reliably documented. A key tool in deciphering earlier metamorphic events has been *in situ* geochronology of metasedimentary rocks (e.g. Williams et al., 1999; Stern and Berman, 2000; Berman et al., 2005). Application of this technology both as an integral part of regional mapping studies and via reconnaissance studies has been successful in helping to define distinct regional events at ca. 2.56 – 2.50 Ga (MacQuoid orogeny; Stern and Berman, 2000), 2.5 – 2.3 Ga (Arrowsmith orogeny; Berman et al., 2005; Schultz et al., 2007; Hartlaub et al., 2007), 2.0 – 1.92 Ga (Thelon - Taltson orogeny; Hoffman, 1988; Bostock and van Breemen, 1992; Thériault, 1992; Henderson et al., 1999; De et al., 2000; McNicoll et al., 2000), 1.91 – 1.87 Ga (Snowbird phase of Hudsonian orogeny; Berman et al., 2007), and 1.86 – 1.75 Ga (main phase of Hudsonian orogeny; e.g. Hoffman, 1988; Lewry and Collerson, 1990; Ansdell, 1995; St-Onge et al., 2002;

Berman et al., 2005). In many cases, these geochronological data have been linked to metamorphic assemblages and/or calculated pressure - temperature conditions, making it possible to infer overall metamorphic conditions at specific times. However in very few places has pre-2.5 Ga metamorphism been dated convincingly in spite of the widespread presence of ca. 2.6 Ga and older basement. In the few places where ca. 2.6 Ga and older metamorphism has been dated (e.g. Skulski et al., 2003; Davis et al., 2004; Berman et al., 2005, 2008, 2010a; Mahan et al., 2006; Harper et al., 2007; Martel et al., 2007), the metamorphic grade is largely obscured by younger events. For these reasons no attempt has been made to construct a metamorphic map of pre-2.5 Ga metamorphism.

After a brief description of the map legend, the five main orogenic events depicted on Sheets 1-3 are briefly described.

2. Acknowledgements:

The author is greatly appreciative of the excellent cartographic support he has received especially from M. Methot, but also from P. Arscott, C. Hemingway, T. Houlahan, and D. Paul during the early stages of the work that led to this publication. I am also very grateful for the contributions made by a number of geologists at various stages in the compilation of the metamorphic zones portrayed in these regions: Saskatchewan (K. Ashton, C. Card, R. Maxheiner), Manitoba (D. Schledewicz, H. Zwanzig, T. Corkery, P. Gilbert), southern and central Baffin Island (M. St-Onge), and Thelon tectonic zone (P. Thompson). Lastly, I thank those who provided unpublished geochronology results (see Table 1), D. Corrigan for his constructive review, and Sylvia Frohberg-Junginger for her very careful cartographic check.

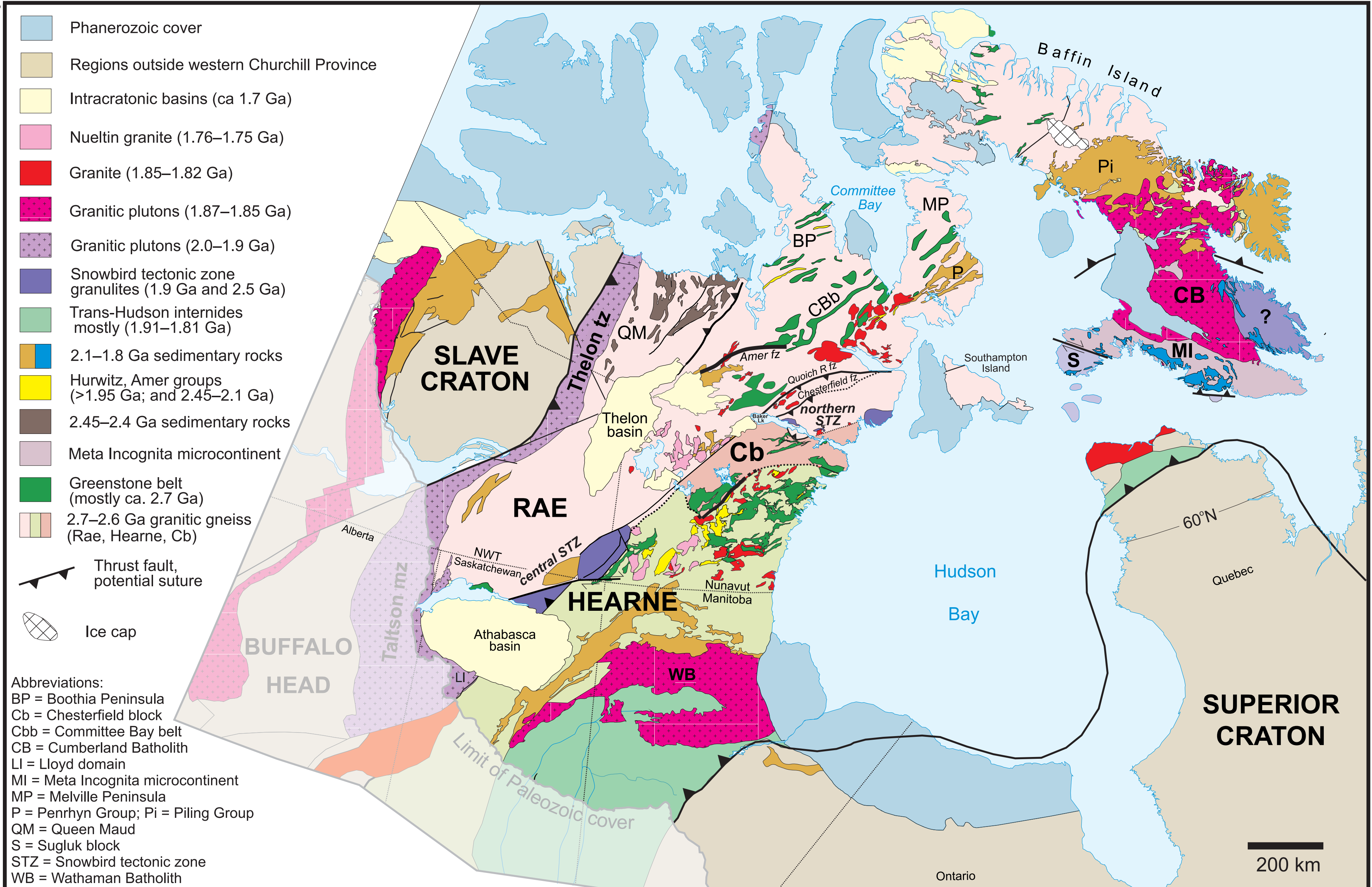

Figure 1. Regional geology of the western Churchill Province and environs

METAMORPHIC MAP LEGEND

The metamorphic map legend used on Sheets 1-3 follows the format described by Berman et al. (2000a) and applied elsewhere (Easton et al., 2005; Ashton et al., 2009). Increasing metamorphic grade is shown with generally warmer colours, except for granulite facies (purple). Where pressure information is available, the metamorphic pressure regime is shown by the orientation of superimposed coloured lines. Pressure is classified as high (vertical lines) or low-intermediate (horizontal lines), primarily on the basis of whether kyanite (Barrovian facies-series) or andalusite-sillimanite (Abukuma facies-series) is present in pelitic rocks. For granulite-facies rocks, a medium-pressure (diagonal lines) subdivision is also used, based on thermobarometric results.

Instead of showing complex overprinting relationships on a single map (e.g. Berman et al., 2000a; Easton et al., 2005), separate metamorphic time slices are presented on Sheets 1-3, as done on the metamorphic map of Saskatchewan (Ashton et al., 2009). On Sheets 1 and 2, two different ages of metamorphism are shown by means of a bright and a pastel colour series of metamorphic zones. In two small regions where separate metamorphic events overlap (Sheet 2), overprint “circles” are used to show the metamorphism of the younger event.

In order to convey the degree (or lack) of control on the age of assigned metamorphic zones, selected geochronological data (together with reference numbers given in parentheses) are shown on Sheets 1-3. Table 1 lists the reference that corresponds to each of these reference numbers. Because of spatial limitations on the maps, the displayed data do not represent a comprehensive compilation, but show instead the most constraining data available in each region. The majority of displayed data are U-Pb monazite and zircon ages that are interpreted as metamorphic, typically crystallizing near the metamorphic peak. Most ^{40}Ar-^{39}Ar ages are not shown because they generally record younger cooling ages through the closure temperature of hornblende (500 °C; Harrison, 1981) and biotite (300 - 350°C; Harrison et al., 1981). Selected ^{40}Ar-^{39}Ar and K-Ar ages are used to show regions where metamorphic grade contrasts can be identified by differences in hornblende or biotite ages. In most of these regions, the preservation of old biotite or hornblende ages is used to indicate that these regions have experienced metamorphic grade lower than greenschist or lower amphibolite, respectively. Where other data are lacking, some K-Ar biotite or hornblende ages are also used to indicate regions that experienced at least greenschist- or amphibolite-facies metamorphism, respectively.

For those data that have not been previously published, the ages displayed on the maps (Sheets 1-3) are rounded to the nearest 5 Myr and associated uncertainties are not shown. The references for these geochronological data are presented in Table 1. An important point to note is that the distribution of the geochronological data provides a visual representation of the regions across which metamorphic zones have been interpolated or extrapolated. Areas of each map that are uncoloured represent either regions that were unmetamorphosed at the time of metamorphism shown on each map, or regions in which there are no data to evaluate whether they were metamorphosed.

Thermobarometric data are not shown on the metamorphic maps for several reasons. First, their display obscures metamorphic relationships shown on the maps, particularly where the data are most dense in parts of Saskatchewan and Manitoba. Second, unlike geochronological data which have very accurately known decay constants, thermobarometric data are based on empirical observations, sometimes conflicting sets of experimental data, and thermodynamic data that are not all accurately known (e.g. Berman, 1990; Spear, 1993). Third, numerical results are highly sensitive to interpretation of mineral zoning patterns (which are not always documented), and to whether results correspond to prograde, near-peak, or retrograde conditions.

GEOLOGICAL BASE

The metamorphic maps shown on Sheets 1-3 were compiled on a composite geologic base, simplified primarily from the following sources:

- Saskatchewan: Macdonald, 1999 (1:1,000,000 scale)
- Manitoba: Manitoba Geological Survey, 1979 (1:1,000,000 scale)
- Hearne/Rae craton (60-64 ^{o}N, 88-100 ^{o}W): Tella et al., 2007 (1:550,000 scale)
- Hearne/Rae craton (60-64 ^{o}N; 100-112 ^{o}W): Pehrsson et al., 2011 (1:550,000 scale)
- Rae craton (64-68 ^{o}N; 81-102 ^{o}W): Patterson and Lecheminant, 1985; Paul et al., 2002 (1:1,000,000), Wheeler et al., 1996 (1:5,000,000)
- Rae craton (68-70 ^{o}N; 81-86 ^{o}W): Schau, 1993 (1:500,000 scale)
- Rae craton (north of 68 ^{o}N) and Baffin Island: Wheeler et al., 1996 (1:5,000,000 scale), Fraser et al., 1978 (1:3,500,000 scale), Jackson and Berman (2000)

Table 1. References for geochronological data plotted on Sheets 1-3.

#	Reference	#	Reference
1	Annesley et. al. (1992a)	72	Heaman et. al. (1999)
2	Annesley et. al. (1992b)	73	Heaman et. al. (1993)
3	Annesley et. al. (1993)	74	Heaman et. al. (1991)
4	Annesley et. al. (1999)	75	Heaman et. al. (1994)
5	Annesley et. al. (1997)	76	Henderson et. al. (1992)
6	Ansdell and Stern (1997)	77	Hunt and Roddick (1987)
7	Ashton et. al. (1999a)	78	Hunt et. al. (1988)
8	Ashton et. al. (1999b)	79	Hunt et. al. (1993)
9	Ashton et. al. (1992)	80	Jackson (1978)
10	Aspler et. al. (2002)	81	Jackson et. al. (1990)
11	Baadsgaard and Godfrey (1967)	81a	Joyce, N. and Ryan, J. (2007, unpub. data)
12	Baldwin et. al. (2006)	82	Koster and Baadsgaard (1970)
13	Baldwin et. al. (2003)	82a	Leech et al. 1963)
14	Baldwin et. al. (2004)	83	Loveridge et. al. (1988)
15	Banks (1980)	84	Lowdon et. al. (1963)
16	Beaumont-Smith and Bohm (2002)	85	Machado et. al. (1990)
17	Bell (1985)	86	MacLachlan, K. and Davis, W. J. (2005, unpub. data)
18	Berman, R.G. (2005, unpub. data)	87	MacLachlan et. al. (2005a)
19	Berman, R.G., Ashton, K., Davis, W.J. (2005, unpub. data)	88	MacLachlan et. al. (2005b)
20	Berman, R.G., Card, C., Ashton, K., Davis, W.J. (2005, unpub. data)	89	MacLachlan et. al. (2004)
21	Berman et. al. (2002b)	90	Mahan et. al. (2006a)
22	Berman, R.G., Davis, W.J., LeCheminant, T. (2005, unpub. data)	91	Mahan et. al. (2006b)
23	Berman, R.G., Davis, W.J., Pehrsson, S. (2005, unpub. data)	92	Martel et. al. (2007)
24	Berman et. al. (2007)	93	McDonough et. al. (2000)
25	Berman et. al. (2002a)	94	McNicoll et. al. (2000)
26	Berman, R.G., Davis, W.J., Sanborn-Barrie, M. (2005, unpub. data)	95	Mills et. al. (2007)
27	Berman, R.G., Davis, W.J., Sandeman, H.S. (2004, unpub. data)	96	Parent et. al. (1999)
28	Berman, R.G., Davis, W.J., Tella, S. (2004, unpub. data)	97	Parrish (1989)
29	Berman, R.G., Maxeiner, R., Davis, W.J. (2005, unpub. data)	98	Rayner et. al. (2005b)
30	Berman, R.G., Qui, H., Davis, W.J., Pehrsson, S. (2006, unpub. data)	99	Reilly (1993)
31	Berman, R.G., Rayner, N. and Sanborn-Barrie, M. (2006, unpub. data)	100	Roddick et. al. (1992)
32	Berman, R.G., Rayner, N., Sanborn-Barrie, M., Chakungal, J. (2009, unpub. data)	101	Roddick and van Breemen (1994)
33	Berman et. al. (2011)	102	Ryan et. al. (2000)
34	Berman, R.G., Ryan, J.J., Davis, W.J., Nadeau, L. (2008, unpub. data)	103	Sanborn-Barrie et. al. (2001)
35	Berman et. al. (2008)	104	Sandeman, H.A. and Berman, R.G. (2001, unpub. data)
36	Berman, R.G., Sanborn-Barrie, M., Rayner, N. (2003, unpub. data)	105	Sandeman (2001)
37	Berman et. al. (2010a)	106	Schultz et. al. (2007)
38	Berman et. al. (2005)	107	Scott, D. (2005, unpub. data)
39	Berman, R.G., Stern, R.A., Tella, S. (1997, unpub. data)	108	Scott and Gauthier (1996)
40	Bethune and Scammell (1997)	108a	Scott and St-Onge (1995)
41	Bethune and Scammell (2003)	109	Scott and Wodicka (1998)
42	Bickford et. al. (1994)	110	Scott (1997)
43	Bohm, C. (2006, unpub. data)	111	Scott (1999)
44	Bostock and van Breemen (1992)	112	Scott et. al. (2002)
45	Bostock et. al. (1991)	113	Shaerer and Deutsch (1990)
46	Bostock and Loveridge (1988)	114	Stern et. al. (2000)
47	Brouand et. al. (2002)	115	Stern et. al. (2003)
48	Carpenter et. al. (2005)	116	Stevens et. al. (1982a)
49	Carson et. al. (2004)	116a	Stevens et. al. (1982b)
50	Cavell et. al. (1992)	117	St-Onge et. al. (2002a)
51	Chiarenzelli et. al. (1998)	118	St-Onge et. al. (2002b)
52	Clark et. al. (1988)	119	St-Onge et. al. (2003a)
53	Clarke et. al. (2005)	120	St-Onge et. al. (2003b)
54	Corrigan et. al. (2001)	121	St-Onge et. al. (2003c)
55	Cranstone et. al. (1976)	122	St-Onge et. al. (2005a)
56	David et. al. (1996)	123	St-Onge et. al. (2005b)
57	Davis, W.J. (2005, unpub. data)	124	St-Onge et. al. (2005c)
58	Davis et. al. (2004)	125	St-Onge et. al. (2005d)
59	Davis et. al. (2006)	126	St-Onge et. al. (2007)
60	Fedorowich et. al. (1995)	127	ter Meer (2001)
61	Flowers et. al. (2006b)	128	Thomas (1994)
62	Flowers et. al. (2006a)	129	Tran (2001)
63	Flowers et. al. (2008)	130	van Breemen and Bostock (1994)
64	Frisch and Hunt (1988)	131	van Breemen et. al. (2007)
65	Frisch and Hunt (1993)	131a	Wanless et al. (1966)
66	Gordon et. al. (1990)	131b	Wanless et al. (1970)
67	Harper et. al. (2006)	131c	Wanless et al. (1972)
68	Hartlaub (2004)	131d	Wanless et al. (1974)
69	Hartlaub, R.P. (2006, unpub. data)	132	Wanless et. al. (1979)
70	Hartlaub et. al. (2005)	133	Williams and Jercinovic (2002)
71	Heaman, L.M. (2004, unpub. data).	134	Wodicka and Scott (1997)

refers to reference number given in parentheses with ages plotted on Sheets 1-3

DESCRIPTION OF METAMORPHIC EVENTS

The five main orogenic events depicted on Sheets 1-3 (ca. 2.56 – 2.50 Ga, ca. 2.50 – 2.30 Ga, ca. 2.0 – 1.92 Ga, ca. 1.91 – 1.87 Ga, ca. 1.86 – 1.75 Ga) are briefly described here with the aim of providing an overview of the main geochronological constraints, styles of metamorphism, and interpreted tectonic settings. For regions within Saskatchewan, most of the descriptions as well as the map linework and geochronological constraints are taken from Ashton et al. (2009). Also included below is brief mention of earlier metamorphic events that are not displayed on Sheets 1-3.

Pre-2.56 Ga Archean metamorphic events

Although there is no shortage of evidence of metamorphic events between 2.8 – 2.6 Ga, associated metamorphic assemblages are generally not known and the data are too dispersed geographically to be displayed meaningfully on a map. Regions with the most unambiguous evidence for specific, pre-2.56 Ga metamorphic events include: the high-grade, ca. 2.85 Ga metamorphism in the Chipman domain of the central Snowbird tectonic zone in northeastern Saskatchewan (Mahan et al., 2006) – southeastern Northwest Territories (Martel et al., 2007), ca. 2.66 Ga, largely greenschist-facies metamorphism of the Kaminak belt (Davis et al., 2004), a ca. 2.60 Ga granulite-facies belt on southern Boothia Peninsula (Hinchey et al., 2006; Berman et al., 2008), the cryptic, ca. 2.58 Ga, greenschist-facies regional contact metamorphism of the Committee Bay belt (Skulski et al., 2003; Berman et al., 2010a), and the ca. 2.7 Ga greenschist- to amphibolite-facies metamorphism of the Eqe Bay region on Baffin Island (Bethune and Scammell, 2003).

MacQuoid Orogeny (ca. 2.56 - 2.50 Ga; pastel colour series on Sheet 1)

Evidence for this event was first discovered in the MacQuoid supracrustal belt within the Chesterfield block (Berman et al., 2000b; Stern and Berman, 2000). A strong, northwest-dipping, northeasterly-strking tectonic fabric is associated with low- to intermediate-P, middle amphibolite-facies metamorphism dated between 2.56 and 2.50 Ga (*in situ* monazite and zircon) in the MacQuoid Lake supracrustal belt (Berman et al., 2000b; Stern and Berman, 2000). Similar ages have been obtained in detailed studies of low-P, upper amphibolite to granulite facies rocks in domain 2 at Angikuni Lake (*in situ* U-Pb monazite; Berman et al., 2002b) and in the Yathkyed Lake region (K-Ar hornblende, Hunt and Roddick, 1987; ^{40}Ar-^{39}Ar hornblende, Sandeman, 2001; U-Pb monazite, MacLachlan et al., 2005a; U-Pb titanite, MacLachlan et al., 2005a; W. J. Davis, unpub. data, 2005). This event is also documented in structurally deeper, granulite-facies rocks of the Snowbird tectonic zone, both to the southwest in in northwestern Saskatchewan (Baldwin et al., 2003; Mahan et al., 2006a, 2006b; Flowers et al., 2008) and to the northeast in the Big Lake shear zone (Ryan et al., 2000), Kramanituar complex (Sanborn-Barrie et al., 2001) and Uvauk complex (Mills et al., 2007). Scant geochronological data (see Sheet 1) suggest that this event extends into felsic gneisses further to the northeast of Chesterfield Inlet, but the grade of metamorphism there is unknown (assumed to be amphibolite facies), except for an upper amphibolite- to granulite-facies region mapped to the east of the Uvauk complex (Tella et al., 1993) and assumed here to be ca. 2.5 Ga in age.

High-grade rocks with ca. 2.55 – 2.53 Ga metamorphic ages have also been identified in northern Manitoba (C. Bohm, unpub. data, 2006) and southern Nunavut (Loveridge et al., 1988) as well as in the region of Peter Lake domain, Sasketchewan (Annesley et al., 1992a, b; R. G. Berman, R. Maxeiner, R., W. J. Davis, unpub. data, 2005). These metamorphic rocks are considered to be unrelated to those discussed above because the Hearne craton is thought to have had a separate and independent history from the Rae craton until their collision at ca. 1.9 Ga (Berman et al., 2007).

A possible tectonic setting for the MacQuoid orogeny is that it initiated as a continental arc (northwest-dipping subduction under the southeast margin of the Rae - Chesterfield block) that generated a large volume of 2.61 – 2.58 Ga plutons in the Rae - Chesterfield block (e.g. van Schmus et al., 1986; LeCheminant and Roddick, 1991; Skulski et al., 2003; Davis et al., 2006; Hinchey et al. 2006), and caused low- to intermediate-*P* metamorphism at ca. 2.56 Ga in the MacQuoid supracrustal belt (Stern and Berman, 2000). A subsequent collisional event may have produced the higher pressure, more calcic outer core compositions and microporphyroblasts observed in garnet from the MacQuoid supracrustal belt and dated at ca. 2.50 Ga (Berman et al., 2000; Stern and Berman, 2000). If this model is correct, it implies that the colliding block rifted away between 2.2 – 2.1 Ga, prior to the ca. 1.9 Ga Snowbird phase of the Hudsonian orogeny (see below, and Sheet 2). The general lack of recognition of a ca. 2.5 Ga tectonic fabric in parts of the Rae craton outside of the Snowbird tectonic zone may reflect strain being taken up in the Rae-Chesterfield boundary zone, as documented in the granulite-facies Big Lake shear zone (Ryan et al., 2000).

Arrowsmith Orogeny (ca. 2.5 - 2.3 Ga; bright colour series on Sheet 1)

The Arrowsmith Orogeny was defined on the basis of ca. 2.35 Ga monazite interpreted to date a distinct deformation and metamorphic event in the western and northwestern parts of the Committee Bay belt (Berman et al., 2005). The clockwise P-T paths and absence of pre- or syn-metamorphic magmatism (Sanborn-Barrie et al., 2011) indicate that low-P (~ 4 kbar), lower amphibolite-facies metamorphism in this region was a response to crustal thickening (Berman et al., 2005). Subsequent work to the west has revealed ca. 2.5 – 2.43 Ga felsic plutonism, sedimentation (Sherman gneisses), and ca. 2.38 Ga metamorphism (monazite growth) in low-*P* (~ 5 kbar, R. G. Berman, unpub. data), granulite-facies rocks of the eastern part of the Queen Maud block, south of Queen Maud Gulf (Schultz et al., 2007). On the southern part of Boothia Peninsula a cryptic, ca. 2.43 – 2.35 Ga high grade event has been documented in metasedimentary rocks that experienced granulite facies at ca. 2.6 Ga (Berman et al., 2008). North of this ca. 2.6 Ga granulite-facies belt, metasedimentary rocks record episodic, low-*P* (~ 5 kbar), at least upper amphibolite-facies monazite growth at ca. 2.5 Ga, ca. 2.43 Ga, and ca. 2.38 Ga, with the ca. 2.5 Ga monazite dating a regional tectonic fabric (R. G. Berman, W. J. Davis, J. J. Ryan, and L. Nadeau, unpub. data, 2008).

For large areas of the Rae craton, data are not available to evaluate the timing of metamorphism. Nevertheless much of the western and northern Rae craton is considered to have been affected by this orogenic event on the basis of widely distributed, but sparse geochronological data. In the northern Rae craton, this event is suggested by ca. 2.48, 2.4, and 2.32 Ga zircon ages of granulite-facies gneiss on northern Boothia peninsula (Frisch and Hunt, 1993), a ca. 2.52 Ga zircon age of granulite-facies gneiss on Devon Island (Frisch and Hunt, 1988), a ca. 2.53 Ga zircon age of enderbite on Bylot Island (northeast Baffin Island; D. Scott, unpub.data, 2005), and ca. 2.31 and 2.39 Ga Rb-Sr whole rock isochron ages and a ca. 2.5 Ga zircon age for upper amphibolite-facies gneiss on northwest Baffin Island (Jackson, 1978; Jackson et al., 1990). The above three age determinations on northern Baffin Island suggest that this region lies within the Arrowsmith orogen, and the 6 – 8 kbar pressures determined from these granulite-facies areas yield a consistent indication of low- to medium-*P* granulite facies metamorphism at this time. There is insufficient geochronological control to allow the southern extent of Arrowsmith orogen on northern Baffin Island to be delineated from Hudsonian orogenic reworking. With the exception of the high-*P* northeast Baffin thrust zone (see Sheet 3), Hudsonian metamorphism is thought to decrease in grade northward from its culmination in the high-*P* Dexterity granulite belt on the north side of the Isortoq fault zone (Jackson and Berman, 2000; Sheet 3), an interpretation supported by the preservation of pre-Hudsonian Rb-Sr isochrons (ca. 1.97 Ga and 2.39 Ga, Jackson, 1978; ca. 2.31 Ga, Jackson et al., 1990) on northwestern Baffin Island. Thus upper amphibolite-facies regions of northwestern Baffin Island (Fraser et al., 1978) more likely represent an earlier metamorphic event, which is assumed here to be the Arrowsmith orogeny. This interpretation gains some support from a 2.52 Ga upper intercept zircon age from migmatitic gneiss near the tip of Cambridge Fjord (Jackson et al., 1990), which is used here as the southern limit of the Arrowsmith orogen, but it may well extend further south to the Isortoq fault zone, coincident with the southern limit of Jackson and Berman's (2000) Committee orogen. It should be noted that data are not available to evaluate the extent that northern Melville Peninsula may have been affected by the Arrowsmith orogeny. Granulite-facies plutonism and/or metamorphism on northwestern Melville Peninsula (Schau et al., 1993) is assumed to be ca. 2.6 Ga in age, based on its proximity to the 2.6 Ga granulite belt on southern Boothia Peninsula (Berman et al., 2008).

In the southwestern Rae craton, the Arrowsmith orogen is marked by low-*P* (~ 4 – 7 kbar), granulite-facies gneisses southwest of the Thelon basin (Sheet 3), which record ca. 2.5 – 2.3 Ga monazite ages (Berman, R.G., Qui, H., Davis, W.J., Pehrsson, S., unpub. data, 2006), and by upper amphibolite- to granulite-facies gneiss with 2.38 – 2.32 Ga monazite and zircon ages in the Taltson basement (Bostock and Loveridge, 1988; McNicoll et al., 2000). Granitic plutons of 2.33-2.29 Ga age in the Uranium City region of northwestern Saskatchewan are interpreted as a product of syn- to post-orogenic magmatism associated with the Arrowsmith orogeny (Hartlaub et al., 2007). Rare ca. 2.35 Ga chemical monazite ages in variably overprinted pelitic gneisses probably record the Arrowsmith metamorphic event in this region (Hartlaub, 2004). Granitoid rocks north of Tazin Lake, Saskatchewan yield ca. 2.5 – 2.3 Ga K/Ar ages in hornblende (Koster and Baadsgaard, 1970; Banks, 1980), suggesting that peak temperatures there exceeded the 500 ± 40°C estimated closure temperature of hornblende (Harrison, 1981), thus reaching at least lower amphibolite facies. Beyond these inferences, the grade of metamorphism in northwestern Saskatchewan is largely unconstrained as is the extent of this event, which is currently based on known occurrences of the 2.33-2.29 Ga plutons (Hartlaub et al., 2007; Ashton et al., 2009).

"Arrowsmith" ages have also been found on the east side of the Rae craton in parts of the Snowbird tectonic zone, occurring in high-grade gneiss and greenschist-grade mylonite at Angikuni Lake (MacLachlan and Davis, unpub. data, 2005) and in upper amphibolite- to granulite-facies gabbro on the south side of the Kramanituar complex (Sanborn-Barrie et al., 2001). The ca. 2.45 Ga ages observed in the Sask craton exposed in structural windows to the south are considered to be unrelated to the Arrowsmith orogeny (Ashton et al., 2009).

The Arrowsmith orogeny is considered to represent the transition from a ca. 2.5 – 2.4 Ga continental arc (east-dipping subduction beneath the western flank of the Rae craton; e.g. Hoffman, 1988) to a ca. 2.4 – 2.3 Ga collisional orogen (Berman et al., 2005), and the evidence for ca. 2.5 Ga regional foliation formation in the Boothia Peninsula region (see above) support this model. However, Schultz et al. (2007) interpret their Queen Maud data to indicate an extensional environment, possibly a continental rift, with sedimentary basin formation (Sherman gneisses) penecontemporaneous with ca. 2.5 – 2.43 Ga plutonism. This region may thus represent a back-arc setting or relatively short-lived rift within an overall convergent tectonic setting.

Thelon-Taltson Orogeny (ca. 2.0 – 1.92 Ga; pastel colour series on Sheet 2)

While metamorphic zones and the Barrovian style of metamorphism in Thelon orogen north of the MacDonald fault have been quite well defined (Henderson et al., 1999), the timing of metamorphism in this region is only very loosely constrained between ca. 2.0 and 1.9 Ga (Henderson and van Breemen, 1992; Roddick and van Breemen, 1994).

Limited geochronological data indicate that this event extends to the northeast where it has been dated in high-grade gneiss at 1.91 Ga on northern Boothia Peninsula (Frisch and Hunt, 1993), 1.93 Ga within the Haughton impact structure on Devon Island (Shaerer and Deutsch, 1990), and 1.94 Ga on Ellesmere Island (Frisch and Hunt, 1988). The eastward extent of reworking during the Thelon orogeny is very poorly constrained. In the eastern Queen Maud block (south of Queen Maud Gulf), only one grain of ca. 1.9 Ga zircon was found among abundant ca. 2.6 – 2.43 Ga detrital zircon in a sample of the Sherman gneiss (Sheet 2; Schultz et al., 2007). It is not clear what grade of metamorphism produced this ca. 1.9 Ga monazite, nor how much of the area experienced, but may not record this event in the nonreactive granulite-facies rocks that were likely dehydrated during the Arrowsmith orogeny (see Sheet 1). On northern Boothia Peninsula, this event is assumed to extend over the region displayed on Sheet 2 based on similarities in structural trends (T. Frisch, personal communication, 2009). Preliminary data (R.G. Berman, W. J. Davis, J. J. Ryan, and L. Nadeau, unpub. data, 2008) suggest that structural and thermal reworking of the Rae craton extended south on Boothia Peninsula to where two amphibolite-facies samples record ca. 1.92 Ga monazite growth (Sheet 2).

South of the MacDonald fault, the evolution of the low-*P*, ultrahigh-T Taltson orogen is well documented. Following continental arc magmatism (east-dipping subduction beneath the western Rae craton) at 1.98 Ga (Thériault, 1992), synorogenic granitic plutons were intruded between 1.96 – 1.92 Ga (Bostock et al., 1991; McNicoll et al., 2000**)**. In order to account for metamorphic temperatures approaching 900 °C, Berman and Bostock (1997) suggested that an oceanic ridge may have been subducted. It should be noted, however, that De et al. (2000) advocate an intracontinental setting, rather than a continental margin setting, on the basis of geochemical characteristics.

In Saskatchewan, a regional northwest-striking deformational fabric is thought to record a tectonic accretionary event in the Taltson orogen (Ashton et al., 2009). North of the Lake Athabasca, metamorphism thought to be coeval with this northwest-striking fabric (Ashton et al., 2005b), is represented by: a) widespread ca. 1.93 Ga leucogranites derived during crustal melting (Hartlaub et al., 2005), b) the growth of 1.94-1.92 Ga metamorphic monazite in pelitic gneisses west of the Black Bay fault (Sheet 2; R.G. Berman, K. Ashton, W.J. Davis, unpub. data, 2005), and c) 1.93-1.92 Ga lower intercept ages from Archean granitoid rocks west of the Black Bay fault (Ashton et al., 2007). The crustal melt leucogranites occur as far east as the central Beaverlodge domain. The dominantly upper-amphibolite facies conditions assigned to this area are based on the partially melted nature and mineral assemblages of rocks defining the northwest-striking fabric (Ashton et al., 2009).

In distinguishing the effects of the Taltson from the Snowbird orogeny, it is important to note that the metamorphic ages within the core of the Taltson orogen are generally no younger than ca. 1.92 Ga (Ashton et al., 2009). In contrast, the Snowbird orogen is generally characterized by metamorphic ages younger than 1.91 Ga (see below). This distinction in ages strongly suggests that the early deformation (characterized by a shallow, northwest-striking foliation) and medium-*P*, granulite-facies metamorphism dated at 1.92 Ga immediately west of the Snowbird tectonic zone and east of Wholdaia Lake (Sheet 2; Martel et al., 2007) are causally related to the Taltson orogeny. It is on this basis that the high-grade, Taltson aged metamorphism has been interpolated across northern Saskatchewan (Ashton et al., 2009) and the southern Northwest Territories.

The occurrence of ca. 2.3 Ga K-Ar biotite ages north of Tazin Lake, northwest Saskatchewan (Koster and Baadsgaard, 1970; Banks, 1980) suggests that metamorphic grade there during the Taltson event did not exceed subgreenschist facies. Towards the north, K-Ar biotite ages progressively young down to ca. 1890 Ma (Banks, 1980), suggesting that the metamorphic grade of the Taltson event increases up to at least greenschist facies, but probably not higher since ca. 2.5 – 2.4 Ga K-Ar hornblende ages are preserved (see Sheet 2). To the northeast of the K-Ar data in this region of the southwestern Northwest Territories, there is only one monazite age (R.G. Berman, H. Qui, W.J. Davis, S. Pehrsson, unpub. data, 2006) which is interpreted to indicate Taltson age amphibolite-facies metamorphism of a ca. 2.3 Ga (Arrowsmith orogeny) granulite-facies gneiss. Based on this sample, a large region of the southwestern Northwest Territories, extending up to the MacDonald fault in the southern Thelon tectonic zone, is considered to have experienced Taltson – Thelon age amphibolite-facies metamorphism (Sheet 2).

South of the Athabasca Basin, most of the exposed basement, with the exception of rocks near the Virgin River shear zone, was subjected to low- to moderate-*P*, granulite-facies conditions (Ashton et al., 2009). This thermal event was accompanied by partial melting in most rock types and the formation of a northwest-striking gneissosity (e.g. Scott, 1985). Well-developed northwest-striking magnetic striping in the sub-Phanerozoic portion of this region (Lloyd domain) is thought to be a regional manifestation of that gneissosity (Ashton et al., 2009). Metamorphic zircon of ca. 1.94-1.93 Ga age in ca. 1.98 Ga arc-type plutonic rocks (Stern et al., 2003) is considered coeval with development of the northwest-striking fabric and the granulite-facies assemblages.

The northwest-striking fabrics and presumed ca. 1.94-1.93 Ga assemblages continue northward beneath the unmetamorphosed Athabasca Group and are exposed in the basement core of the Carswell structure. Anatectic granitoid rocks sampled from the basement to the Athabasca Group just south of the Carswell structure also returned 1.93-1.91 Ga U-Pb zircon ion microprobe ages (Brouand et al., 2002). In addition, monazite in pelitic rocks from the basement core of the Carswell structure are ca 1.93 Ga in age (R.G. Berman, K. Ashton, W. J. Davis, unpub. data, 2005). These ages suggest that the metamorphic culmination during the Taltson orogeny extended southeastward from the Northwest Territories and northeastern Alberta to the Virgin River shear zone (Ashton et al., 2009).

Snowbird phase of Hudsonian Orogeny (ca. 1.91 – 1.87 Ga; bright colour series on Sheet 2)

The occurrence of a high-pressure, ca. 1.9 Ga metamorphic belt along much of the length of the Snowbird tectonic zone has been proposed to reflect the ca. 1.9 Ga collision between the Hearne craton and the composite Rae-Chesterfield block (Berman et al., 2007; see also Hoffman, 1988). This orogenic event is considered to represent the earliest phase of the Hudsonian orogeny, involving pre-1.865 Ga (Cumberland – Wathaman Batholiths) microcontinent collisions with the southeast flank Rae craton (Berman et al., 2007) following terminal collisions in the Taltson-Thelon orogen on the west side of the Rae craton. The geochronological and metamorphic data defining this belt form three domains. First, Barrovian-style (kyanite-staurolite-garnet), middle amphibolite-facies metapelitic rocks in the Josephine River belt crystallized during a tectonometamorphic event that is regionally well dated at 1.89 Ga (Sheet 2; Berman et al., 2002; 2007). The Paleoproterozoic age of these sediments, which contain detrital zircon as young as 2.3 Ga (Berman et al., 2007), suggests that they may have formed on the margin of the composite Rae-Chesterfield block before its interpreted collision with the Hearne craton at ca. 1.91 Ga (Berman et al., 2007). This timing would provide ~20 Ma to thicken and incubate these sediments to the temperature (lower amphibolite-facies) required for ca. 1.89 Ga monazite growth. This domain extends northeast to the western part of exposed bedrock on Southampton Island**,** where ca. 1.88 Ga monazite inclusions form part of an early fabric within garnet porphyroblasts in samples from the western side of the exposed bedrock (Berman et al., 2011). The inferred eastern limit of the Snowbird orogen on Southampton Island (Sheet 2) lines up with the geophysically defined eastern edge of the Hearne craton, providing a plausible explanation for the apparent absence of ca. 1.88 Ga monazite to the east of this boundary on Southampton Island (Berman et al., 2010c). Further east on southwestern Baffin Island, a thermal event marked by ca. 1.87 Ga monazite growth may be related to the ca. 1.9 Ga accretion of an Archean (Sugluk, Corrigan et al., 2009) crustal block to Meta Incognita microcontinent (Wodicka et al., 2010) prior to the ca. 1.88-1.865 Ga collision of this composite block with the Rae craton (see below and Sheet 3).

The second high-*P* domain is interpolated through a large portion of the Chesterfield block on the basis of several samples near MacQuoid and Angikuni Lakes (Sheet 2). In this region garnet porphyroblasts show post-tectonic, ~10 kbar overgrowths at middle amphibolite to upper amphibolite facies (Berman et al., 2000, 2002b; 2007). The best constraints on timing suggest that this post-tectonic, high-*P* event occurred at 1.88 Ga (Berman et al., 2007). In the vicinity of Yathkyed Lake, a subgreenschist-facies region must have been at higher crustal levels to preserve K-Ar biotite (and hornblende) cooling ages > 2300 Ma (Wanless et al., 1979; Sandeman, 2001). The extent of this low-grade region is only very approximately constrained by available isotopic data (Sheet 2). Therefore the metamorphic patterns shown for ca. 1.9 Ga are assumed to be the same at ca. 1.8 Ga (see Sheet 3), when there is more control from the distribution of greenschist-facies Paleoproterozoic Hurwitz Group metasedimentary rocks.

The third high-*P* domain, which defines the northern segment of the Snowbird tectonic zone, consists of 12-13 kbar, mafic gabbro-anorthosite that intruded at granulite facies into granitic lower crust at ca. 1.901 Ma (Kramanituar complex; Sanborn-Barrie et al., 2000) at 1.907 Ga (Uvauk complex, Mills et al., 2007), and at 1.905 Ga (Daly Bay complex; Berman et al., 2007). In this segment of the STZ, metamorphism began prior to gabbro-anorthosite emplacement, based on ca. 1.92 – 1.91 Ga monazite in granulite-facies metasediments at Kramanituar complex (Sanborn-Barrie et al., 2001) and 1.917 Ga granulite-facies tonalite at Daly Bay complex (Berman et al., 2007). The region of > 10 kbar lower crust, which contains smaller lenses of high-*P* amphibolite, extends north to the Chesterfield fault zone, a northwest vergent thrust interpreted as responsible for at least partial exhumation of this high-*P* domain (Berman et al., 2007). The timing of this exhumation is constrained by ^{40}Ar-^{39}Ar hornblende ages between ca. 1.92 – 1.88 Ga (H. A. Sandeman, R. G. Berman, unpub. data, 2001). In the eastern part of this region (with the possible exception of western Southampton Island discussed above), there are no mineralogical or quantitative data to allow assessment of the pressure or timing of crystallization, which is assumed to be ca. 1.9 Ga on Sheet 2.

The third high-*P* domain described above in the northern STZ is similar to ca. 1.91 – 1.90 Ga, 10-12 kbar granulite-facies rocks that occur in the east-vergent, hanging wall of the Chipman shear zone (Martel et al., 2007; Legs Lake shear zone of Mahan et al., 2003), and comprise the central STZ in northwest Saskatchewan (East Athabasca mylonite zone; Williams et al., 1999; Baldwin et al., 2003; Williams and Hanmer, 2005; Flowers et al., 2006; Mahan et al., 2008) and southern Northwest Territories (Snowbird Lake map sheet; Martel et al., 2007). Recent detailed studies suggest that most of the region has undergone two periods of high-*P* metamorphism and deformation at ca. 2.56 – 2.50 Ga (Baldwin et al., 2003, 2006; Mahan et al., 2006a; Flowers et al., 2008) and ca. 1.9 Ga (Baldwin et al., 2004; Flowers et al., 2006; Mahan et al., 2006a, 2006b) prior to ca. 1.85 – 1.80 Ga partial exhumation along the Chipman shear zone (Mahan et al., 2003; Martel et al., 2007) and ca. 1.8 Ga dextral offset along the Grease River shear zone (Mahan et al., 2005; Dumond et al., 2008). Whereas pressures of 10-12 kbar characterize much of this area in Saskatchewan, 1.904 Ga eclogite (Sheet 2) records conditions up to 1000 °C and 20 kbar (Baldwin et al., 2004). In the eastern part of this region, mafic dykes (Chipman swarm) dated at 1.896 ± 0.001 Ga (Flowers et al., 2006) were emplaced synkinematically with respect to a steep, northeast-striking fabric that regionally records sinistral strike slip displacement and overprints an earlier, shallow-dipping gneissosity (Williams et al., 1995; Mahan et al., 2008). Petrological studies suggest contrasting ca. 1.9 Ga P-T paths, with the eclogite recording prograde burial followed by isothermal decompression (Baldwin et al., 2003, 2004), whereas mafic granulites to the east are interpreted to record isobaric heating prior to cooling and decompression (Mahan et al., 2008). Also enigmatic is the interpretation that rocks in the central STZ of Saskatchewan experienced long-term residence in the deep crust between ca. 2.5 Ga and 1.9 Ga (e.g. Williams and Hanmer, 2005; Flowers et al., 2006; Mahan et al., 2008), whereas the Rae craton in the adjacent region of the Northwest Territories (the area east of Wholdaia Lake; Snowbird Lake map sheet) is characterized by ca. 2.56 Ga basement gneiss which had been exhumed prior to deposition of < 2.07 Ga (youngest detrital zircon) sediments that were buried prior to experiencing lower pressure (7.6 kbar) granulite-facies metamorphism and deformation (northwest-striking, shallow foliation and gneissosity) at 1.92 Ga (Martel et al., 2007).

Based on metamorphic zircon and monazite ages in northern Saskatchewan, the region affected by the 1.91-1.90 Ga "Snowbird" event extends west of the central Snowbird tectonic zone, broadly coinciding with the structural overprint of regional northeast-trending folds on the earlier (ca. 1.92 Ga) northwest-striking fabric (Ashton et al., 2009). Because of the overlap in this area between the Taltson and Snowbird events, "overprint circles" display the granulite-facies metamorphism in this region (Sheet 2) that is characterized by somewhat lower pressures than those within the central Snowbird tectonic zone, suggesting that a tilted crustal section exposes medium-*P* granulites west of the Grease River shear zone (Ashton et al., 2009). East of the Black Bay fault (see inset on Sheet 2), the overprinting of these two events has produced an as yet unresolved range of 1.93-1.90 Ga chemical monazite ages from pelitic gneisses, even though metamorphic zircon and titanite yield well-constrained 1.91-1.90 Ga ages (Hartlaub, 2004). Because of this lack of geochronological resolution, it is unclear at what time (Taltson or Snowbird) the metamorphic zonation from lower amphibolite-facies to upper amphibolite-facies formed (Ashton et al., 2009). The Black Bay Fault may have been initiated at this time as a northwest-verging thrust fault (Ashton et al., 2009). West of the Black Bay Fault, metamorphic grade is inferred to decrease towards the greenschist facies that is recorded further west in the Waugh Lake group (see inset on Sheet 2).

South of the Athabasca Basin, northwest-striking fabrics (in the Lloyd domain) are overprinted by northeast-trending folds that tighten towards the Virgin River shear zone (Ashton et al., 2009). The earliest northeast-striking mylonitic fabric in the shear zone is thought to be the same age as that folding. Mafic dykes that cut the older gneissosity were deformed by the northeast-trending folds. The presence of locally derived leucosomes axial planar to the northeast-trending folds, hornblende replacing primary pyroxenes in the dykes, and hornblende in the oldest mylonitic rocks suggest that this deformation developed under upper amphibolite-facies conditions (Ashton et al., 2009). The main northeast-striking fabric is older than a suite of ca. 1.83-1.82 Ga biotite granites (e.g. Stern et al., 2003) that stitch the Virgin River shear zone. Those granites contain only a weak northeast-striking foliation that is thought to result from reactivation of the shear zone (Ashton et al., 2009). Zircon overgrowths have been dated at ca. 1.90 Ga in ca. 1.98 Ga arc-type plutonic rocks (Stern et al., 2003) immediately west of the Virgin River shear zone. Given that the northeast-striking fabric has been established to have formed at 1.90 Ga in the STZ northeast of the Athabasca basin (Flowers et al., 2006), these ca. 1.90 Ga zircon overgrowths probably record coeval development of the northeast-striking structures in this region.

The Virgin River shear zone was subjected to repeated reactivation during its uplift history. The Virgin Schist Group is a lower to middle amphibolite-facies, commonly phyllonitic, supracrustal succession exposed in the core of the Virgin River shear zone. The provenance of these rocks remains unclear as does the age of superimposed metamorphism, although 1.90 Ga is considered a minimum age (Ashton et al., 2009).

Main phase of Hudsonian Orogeny (ca. 1.86 – 1.75 Ga; Sheet 3)

The Trans-Hudson Orogen extends across the western Churchill Province from northern Saskatchewan to Baffin Island. Following decades of focussed research in the Manitoba-Saskatchewan and North Quebec – Baffin Island segments, it is the best characterized of all orogenic events affecting this part of the Canadian shield.

Saskatchewan – Manitoba segment of the Trans-Hudson orogen

In the Saskatchewan – Manitoba segment of the orogen, metamorphism and deformation resulted from arc accretion followed by collision of the Hearne, Sask, and Superior crustal blocks. Metamorphism within the zone of accreted juvenile rocks mainly resulted from their interaction with the Sask craton in the south and Hearne craton in the northwest.

Along the Hearne craton margin, northwest-directed subduction under the continent may have commenced as early as 1.92 Ga (Tran et al., 2003; R. Maxeiner et al., in prep) and been accompanied by arc accretion. Alternatively, subduction with the opposite polarity may have been followed by ca. 1.88 – 1.865 Ga accretion of the La Ronge – Lynn Lake arcs to the southeastern Hearne craton margin (Bickford *et al.* 1990; 1994; Corrigan *et al.* 2005). By 1.86 Ga, continued westward transport of the oceanic plate resulted in development of a major continental arc (Wathaman Batholith) on the eastern margin of the Hearne craton (e.g. Meyer et al., 1992). Evidence for early thermotectonic processes (not shown on Sheets 2 or 3) comes from 1917 Ma volcanogenic rocks of the Porter Bay Complex in the Peter Lake domain (Maxeiner et al., in prep.), which were subjected to a low-grade thermotectonic overprint prior to emplacement of 1913 Ma Porter Bay Complex intrusions (Rayner et al., 2005a). As shown on Sheet 3, the lower sequence of the Campbell River Group and its basement in the Peter Lake domain are interpreted (Ashton et al., 2009) to have been metamorphosed to upper amphibolite facies during northward subduction and emplacement of the 1.92-1.91 Porter Bay Complex, a proposed continental arc.

Earliest deformation in the Rottenstone domain, Saskatchewan is reportedly older than 1.87 Ga (Bickford et al., 1987; Maxeiner et al., 2004). Bickford et al. (1990) and Meyer et al. (1992) relate this early deformation to arc accretion (i.e. La Ronge arc) along the eastern margin of the Hearne craton via southeast-directed subduction at about 1.87 Ga. Alternatively, early deformation, at least in the La Ronge domain, Saskatchewan, may have been due to intra-oceanic amalgamation processes (Maxeiner et al., 2005) and interaction with the Flin Flon-Glennie Complex (Lucas et al., 1997). The latter was amalgamated with attendant deformation and possible metamorphism from juvenile ocean floor and arc assemblages at 1.88-1.87 Ga (Lucas et al., 1996). Evidence of these earlier metamorphic conditions is scarce due to subsequent overprinting by the main <1.83 Ga metamorphic event (Ashton et al., 2009).

The Flin Flon – Glennie Complex together with sedimentary rocks of the Kisseynew domain, were thrust over the Archean Sask craton (exposed in several tectonic windows) at about 1.83 Ga (Ashton et al., 2005a), resulting in widespread partial melting. Isograds are modified after Plint et al. (1996). Rare relict kyanite observed west of the Saskatchewan – Manitoba border (Ashton and Wheatley, 1986) suggests that these early metamorphic conditions locally involved higher pressures than those inferred from the final stable assemblages (Ashton et al., 2009). The Flin Flon block must have remained at relatively high structural levels during most of the collisional phase of the orogeny as it preserves greenschist- to subgreenschist- facies assemblages.

The resulting episode of renewed tectonic thickening produced widespread upper amphibolite-facies metamorphism in the Rottenstone domain (Ashton et al., 2009), recorded by the emplacement of abundant anatectic granitic sheets that range from ca. 1.84 to 1.81 Ga in age

(MacLachlan et al. 2004; Clark et al., 2005). Metamorphic monazite in pelitic migmatite records an age of ca. 1.82 Ga. In slightly lower-grade rocks of the former Crew Lake Belt of the eastern Rottenstone domain, which exhibit only incipient partial melting (Ashton et al., 2009), metamorphic monazite in psammite ranges from 1.83 to 1.80 Ga (N. Rayner, unpub. data, 2006). Metamorphic titanite from various rock types in the central Rottenstone domain ranges from ca. 1.81 to 1.79 Ga (MacLachlan et al. 2004) and likely records the timing of cooling through the ~600° C closure temperature (Heaman and Parrish, 1991). In the Mudjatik and Wollaston domains of the Hearne craton, this early pulse of high-grade Trans-Hudson metamorphism is recorded by upper amphibolite to granulite-facies mineral assemblages in Wollaston Supergroup rocks (Schledewitz, 1978; Harper et al., 2005; Card et al., 2006).

Ashton et al. (2009) interpet that renewed tectonic thickening attendant to terminal collision caused a thermal pulse responsible for widespread recrystallization of mylonite zones and variable overprinting of the early metamorphic assemblages via terrane accretion (e.g. Tran, 1997). Although the orientation of isograds has been modified by late folding in some areas (e.g. the contact between the northeastern Glennie domain and the Kisseynew domain), elsewhere (e.g. the area northwest of Flin Flon) they crosscut the main regional fabric at high angles (Ashton et al., 2009). Portions of the Flin Flon – Glennie and La Ronge arc complexes that remained at shallow crustal levels during this deformation have been preserved as granite-greenstone belts marked by metamorphic ‘lows’ (Ashton et al., 2009). Most of the Hearne craton, except for the Peter Lake domain, was also reworked at this time under upper amphibolite to granulite-facies conditions (e.g. Orrell et al., 1999; Tran, 2001; Chakungal et al., 2004). This main metamorphic event took place at moderate pressures (i.e. 5-7 GPa) and culminated at about

1.82-1.80 Ga (e.g. Annesley et al., 1992, 1997; Syme et al., 1998; Ashton et al., 1999; Harper et al., 2006**;** van Breemen et al., 2007).

Titanite ages throughout the orogen (e.g. Ashton et al., 2005a; van Breemen et al., 2007), as well as ^{40}Ar-^{39}Ar hornblende and mica ages east of the Tabbernor fault (Schneider et al., 2007), document uplift and cooling to about 1.77 Ga, at which time felsic magmas (Nueltin and Jan Lake granite suites) were intruded over much of the region (e.g. Heaman et al., 2003; Bickford et al., 2005). Sporadic 1.76-1.75 Ga titanite and rutile metamorphic ages (Sheet 3), as well as ca. 1.74 – 1.72 Ga ^{40}Ar-^{39}Ar mica ages west of the Tabbernor fault, may result from high heat flow associated with this pulse of granitic magmatism (Ashton et al., 2009).

North Quebec – Baffin Island segment of the Trans-Hudson orogen

A large part of the metamorphic map for Baffin Island is taken from that of Fraser et al. (1978), updated on the basis of more recent studies in the Eqe Bay region (Bethune and Scammell, 2003), central Baffin Island (St-Onge et al., 2005; St-Onge, personal communication), southern Baffin Island (St-Onge et al., 1999; 2006), and southwest Baffin Island (Sanborn-Barrie et al., 2008; St-Onge, personal communication). The following description of the North Quebec – Baffin Island segment of the Trans-Hudson orogen follows that summarized by St-Onge et al. (2002; 2006; see also references within). This segment of the orogen preserves a record or several accretionary and magmatic events prior to collisional reworking of three main tectonic assemblages (from the southern, lowermost assemblage to the northern, uppermost assemblage): (1) lower plate, Archean rocks of the Superior craton with overlying Paleoproterozoic sedimentary and volcanic cover rocks, (2) upper plate, accreted Paleoproterozoic ophiolite (c. 2.0 Ga), fore-arc clastic rocks, and the ca. 1.86 – 1.82 Ga Narsajuac magmatic arc, and (3) upper plate, accreted rocks representing (a) Meta Incognita microcontinent basement with a <1.93 Ga north-facing clastic – carbonate shelf succession (Lake Harbour Group), (b) Archean Rae craton basement with 2.2 – 1.92 Ga south-facing clastic and carbonate cover rocks (Piling Group), and (c) 1.865 – 1.848 Ga charnockitic rocks of the Cumberland Batholith.

The earliest event (D_0) involved ca. 1.88 – 1.865 Ga accretion of Meta Incognita microcontinent to the southern Rae craton, which produced a north-vergent, thin-skinned thrust and fold belt (north of the Cumberland Batholith, see Piling Group, Fig. 1) with ca. 1.88 – 1.85 Ga (M_0; see Sheet 2 for older ages) cordierite-andalusite assemblages recording P-T conditions of 550 – 600 °C and 3-4 kbar (Allan and Pattison, 2003).

Accretion of Narsajuac arc (or Sugluk terrane, Corrigan et al., 2009) to the southern margin of Meta Incognita microcontinent produced D_1 deformation, and together with intrusion of the Cumberland Batholith, M_{1a} regional lower granulite-facies (garnet-cordierite-K feldspar) metamorphism. This ca. 1849 – 1835 Ma metamorphic event records conditions of 790 – 845 °C and 7-8 kbar on southern Baffin Island (St-Onge et al., 2007), where subsequent thermal events are recorded at 1833 – 1829 (M_{1b}) and at 1836 – 1825 Ma (M_{1c} at lower structural levels). On southwest Baffin Island, monazite records the main tectonometamorphic event between ca. 1850 and 1835 Ma at metamorphic conditions that vary from upper amphibolite facies (~700 °C – 5 kbar) to lower amphibolite facies (Rayner et al., 2008). In a lower to middle-amphibolite facies supracrustal belt on southwestern Baffin Island, peak P-T conditions of 620 °C - 4 kbar appear to have been attained on an anti-clockwise P-T-t path (Smye et al., 2009) which contrasts with the clockwise P-T-t paths determined for samples further east (St-Onge et al., 2006; Smye et al., 2009). These differences are considered to reflect the pre-deformation emplacement of Narsajuac arc plutons in the west and syn- to post-deformation intrusion of the Cumberland Batholith in the east (Smye et al., 2009).

Further north in the Eqe Bay region, a low-*P* greenschist to amphibolite-facies region dated at ca. 1.85 – 1.82 Ga is juxtaposed against granulite-facies rocks to the north (Jackson and Berman, 2000; Bethune and Scammell, 2003). The south-dipping Isortoq fault zone that separates these areas (Sheet 3) is thought to represent a ca. 1.85 Ga northwest-vergent thrust that was reactivated as a normal fault (Bethune and Scammell, 2003). Further north and northeast, amphibolite- to granulite-facies gneisses (some with thermobarometric data suggesting high pressures, Sheet 3) are thought to have been exhumed during a later stage of southwest-vergent thrusting (Jackson and Berman, 2000).

Collision between the Superior craton and the composite upper plate produced ca. 1820 – 1805 Ma D_2 south-vergent, recumbent folds and thrusts in the upper plate, with M_2 retrograde upper amphibolite-facies metamorphism dated between ca. 1820 – 1808 Ma (St-Onge et al., 2006). In the lower plate, ca. 1830 – 1815 Ma piggy back sequence thrusting produced greenschist- to amphibolite-facies, Barrovian style metamorphism (up to 575 °C and 9.1 kbar) in the Cape Smith belt (St-Onge et al., 2000). Between ca. 1815 to 1785 Ma (Scott and St-Onge, 1995) the Superior craton underwent high-*P* (8 – 10 kbar; 585 – 725 °C; St-Onge and Ijewliw, 1996) hydration reactions at the basement-cover contact. This amphibolite-facies hydration zone has been assumed to extend across the width of the basal décollement at the base of the Cape Smith belt (Sheet 3). The occurrence of young (ca. 1660 – 1590 Ma) K-Ar hornblende cooling ages (Wanless et al., 1974) within two samples of the Superior craton north of the basal décollement (Sheet 3) has been taken to indicate that this portion of the Superior craton experienced at least greenschist-facies thermal conditions. Post-collisional, M_3 metamorphism at ca. 1795 – 1785 Ma involved reequilibration at amphibolite facies and emplacement of leucogranites in both the upper and lower plates.

Central Rae segment of the Trans-Hudson orogen

Metamorphic and geochronological studies in the central Rae craton west of Hudson Bay has demonstrated extensive reworking of this region between ca. 1.86 and 1.77 Ga (Carson et al., 2004; Berman et al., 2005, Berman et al., 2010a). The northwestern part of the Committee Bay belt provides a window into early tectonometamorphism, with metapelitic rocks recording monazite and zircon ages between 1.86 – 1.84 Ga (Carson et al., 2004; Berman et al., 2005; Berman et al., 2010a). Low-*P* (5 kbar maximum) metamorphism as a response to tectonic thickening (dominantly northwest-vergent D_2 folds; Sanborn-Barrie et al., 2011) is indicated by the the absence of plutonism prior to or during metamorphism (Sanborn-Barrie et al., 2011), clockwise P-T paths retrieved from lower to upper amphibolite-facies metapelitic rocks, and garnet and staurolite growth that is synchronous or late with respect to deformation. Given the ~10 to 30 Ma time lag for thermal incubation after onset of thickening (depending on the heat productivity of the rocks and peak temperatures relative to the temperature of monazite growth; Berman et al., 2010b), this metamorphism is considered to reflect ca. 1.87 Ga collision of a Meta Incognita microcontinent with the Rae craton (Berman et al., 2005, 2010a, 2010b), consistent with the 1.88 – 1.865 Ga constraints for the timing of this collision (St-Onge et al., 2002). The eastern end of the Committee Bay belt records D_2 deformation and metamorphism at ca. 1.82 – 1.81 Ga (Berman et al., 2010a; Sanborn-Barrie et al., 2011), which is thought to reflect its relative geographic proximity to the eastern promontory of the Superior craton during its ca. 1.82 Ga collision with the composite Rae - Meta Incognita crustal block (Berman et al., 2010a, 2010b). Further east on southern Melville Peninsula, data are available to evaluate the timing of near-peak metamorphism and deformaton, except for a ca. 1.80 Ga whole rock Rb-Sr age (Henderson, 1983).

On southern Boothia Peninsula north of the Committee Bay belt, the main Paleoproterozoic deformation and metamorphism is recorded by later monazite growth (ca. 1.83 Ga; Berman et al., 2008) than in the western Committee Bay belt, a difference that can be attributed to the deeper structural levels exposed in the Boothia region (~ 7 kbar; Berman et al., 2008) than in the Committee Bay belt (~5 kbar; Berman et al., 2005, 2010a). To the south in the region of Woodburn Lake, the main Paleoproterozoic deformation and lower to middle amphibolite-facies metamorphism is dated at 1.84 – 1.83 Ga (R. G. Berman, W.J. Davis, S. Pehrsson, unpub. data, 2005). On Southampton Island, metamorphic grade ranges from upper amphibolite facies to granulite-facies, with most *in situ* monazite dating (Berman et al., 2011) indicating ca. 1.86 – 1.82 Ga ages. However, some of the granulite-facies metamorphism is very likely older than ca. 1.9 Ga (not shown on Sheet 2), as dated charnockites have yielded disparate ages of ca. 3.0 Ga, 2.67 Ga, and 1.93 Ga (N. Rayner,, unpub. data, 2009). Similarly, upper amphibolite to granulite-facies metamorphism of metaplutonic rocks on southern Melville Peninsula (sheet 3) is considered in part Archean and in part Paleoproterozoic in age. The diachroneity in the timing of deformation and metamorphism across the central Rae craton illustrates the domainal pattern of reworking which appears to be controlled by heterogeneity in transmission of tectonic stresses, in exposed crustal levels (e.g. deeper levels in the Boothia mainland region (Berman et al., 2008), in the heat productivity of the thickened rocks which controls the rate at, and grade to which rocks are heated and thermally weakened (e.g. 1.86 Ga monazite growth in the western CBb; Berman et al., 2010b), and in the relative timing of heat input from plutonic rocks (e.g. 1.86 Ga monazite growth on Southampton Island; Berman et al., 2011).

In contrast to the 1.86 – 1.81 Ga tectonometamorphism described above in the Committee Bay belt, the central part of this belt records monazite growth at 1.79 Ga that is associated with dextral shearing at lower amphibolite facies in the south (Berman et al., 2010a) and with upper amphibolite facies in the north (Carson et al., 2004). Further northwest on southern Boothia Peninsula, ca. 1.79 Ga monazite ages are anomalously young relative to the regions further east (Berman et al., 2008). This younger episode of metamorphic crystallization is thought to reflect slightly deeper crustal levels (and temperatures above that required for monazite growth) within a north to northwest-trending synform produced during cross-folding in the final stages of amalgamation of Laurentia (Berman et al., 2008; 2010a, 2010b; Sanborn-Barrie et al., 2011).

The areal extent of reworking of Rae crust during the Hudsonian orogeny is still quite poorly known, with few constraints outside of the areas discussed above. Large regions with uncertain grade of 1.86 – 1.75 Ga metamorphism occur between Chesterfield Inlet and Wager Bay, and on Boothia and Melville Peninsulas as well as the adjacent area on northwestern Baffin Island. These regions are shown as either undivided amphibolite-facies or greenschist- to amphibolite-facies metamorphic grade, primarily based on the sparse occurrence of K-Ar ages younger than ca. 1750 Ma for hornblende or biotite, respectively (Wanless et al., 1966, 1979; Stevens et al., 1982a, b). The region of greenschist- to amphibolite facies on northwestern Melville Peninsula outlines an area that previously reached granulite facies (Schau et al., 1993), most likely at ca. 2.6 Ga. A K-Ar hornblende age of 2657 Ma (Stevens et al., 1982a) suggests that one area on the northern flank of the Penrhyn Group (just south of Committee Bay) did not exceed greenschist facies.

Northwest Baffin Island is characterized by upper amphibolite-facies regions, but scant geochronological data suggest these are part of the ca. 2.5 – 2.3 Ga Arrowsmith orogen (see Sheet 1), not part of the Trans-Hudson orogen as originally thought (e.g. Fraser et al., 1978). Preservation in this region of ca. 2.31 – 2.39 Ga Rb-Sr isochron ages suggest that Hudsonian metamorphic grade did not exceed lower amphibolite facies. These interpretations suggest that metamorphic grade decreases northward from the ca. 1.82 Ga Dexterity granulite belt (Jackson and Berman, 2000) which has been juxtaposed against greenschist- to lower amphibolite-facies rocks along the Isortoq fault zone (Bethune and Scammell, 2003). On the northeast side of Baffin Island, upper amphibolite- to granulite-facies areas are considered part of the Northeast Baffin thrust belt (Jackson and Berman, 2000). Within this belt, reconnaissance level thermobarometric results (without geochronological control) indicate high-*P* granulite facies conditions (Jackson and Berman, 2000). Because upper amphibolite-facies surrounding regions do not have any pressure indicators, it is not known whether high-*P* conditions characterize the whole belt.

The north and northwestern limits of reworking related to the Hudsonian orogeny are also poorly constrained. Two geochronological studies (Schultz et al., 2007; Berman et al., 2005) have found no evidence for Hudsonian age zircon or monazite growth in the Queen Maud region, and K-Ar data are lacking there. On southern Boothia peninsula, Hudsonian reworking extends north to a ca. 2.6 Ga granulite belt (Berman et al., 2008; shown as an upper amphibolite-facies belt on Sheet 1;), which is interpreted to have been thrust southwards (Berman et al., 2008), leading to earlier cooling through hornblende closure (ca. 1.8 Ga, N. Joyce and J. Ryan, unpublished data) than the region to the south of this belt (ca. 1.76 Ga, N. Joyce and J. Ryan, unpublished data). Further north, there is an indication from monazite ages in one sample (Sheet 3) of cryptic Hudsonian reworking. On northern Boothia Peninsula, two K-Ar hornblende cooling ages are 1918 ± 12 Ma (Roddick et al., 1992) and 1831 ± 20 Ma (Frisch and Hunt, 1993), whereas K-Ar biotite ages are younger than ca. 1740 Ma (Wanless et al., 1979; Frisch and Hunt, 1993). These data suggest that Hudsonian thermal reworking was at least greenschist facies, and locally exceeded lower amphibolite facies (Sheet 3).

In northwestern Saskatchewan, the Hudsonian event has not been recognized west of the Virgin River shear zone (Orrell et al., 1999). North of Lake Athabasca, the effects of the Hudsonian orogeny diminish west of the Black Bay fault (Ashton et al., 2009). Along strike in the Northwest Territories, the few available data suggest that much of the Rae was at least thermally reworked at amphibolite facies at least as far west as the Howard Lake shear zone. Further west, the Nonacho sediments were metamorphosed to lower subgreenschist facies (Aspler, 1985).

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