Canadian Journal of Earth Sciences
Revue canadienne des sciences de la Terre

Transtensional origin of multi-order cross-folds in a high-grade gneiss complex, southwestern Grenville Province: formation during post-peak gravitational collapse

Journal:	<i>Canadian Journal of Earth Sciences</i>
Manuscript ID	cjes-2015-0212.R2
Manuscript Type:	Article
Date Submitted by the Author:	08-Jul-2016
Complete List of Authors:	Schwerdtner, Walfried; University of Toronto, Geology Rivers, Toby; Memorial University of Newfoundland, Earth Sciences Tsolas, Jason; GoldCorp Waddington, Dennis; Waddington Consulting Page, Sydney; University of Toronto Yang, Jack; Formerly University of Toronto
Keyword:	gneiss complex, cross-fold, constrictional strain, gravitational collapse, metamorphic core complex

SCHOLARONE™
Manuscripts

Transtensional origin of multi-order cross-folds in a high-grade gneiss complex, southwestern Grenville Province: formation during post-peak gravitational collapse

W.M. (Fried) Schwerdtner, Toby Rivers, Jason Tsolas, Dennis H. Waddington, Sydney Page and Jack Yang

W.M. Schwerdtner. Department of Earth Sciences, University of Toronto, 22 Russell Street, Toronto, ON, M5S 3B1, Canada

T. Rivers. Department of Earth Sciences, Memorial University of Newfoundland, 300 Prince Philip Drive, St. John's, NL, A1B 3X5, Canada

J. Tsolas. Red Lake Gold Mines, 17 Mine Rd, Balmertown, ON, P0V 1C0, Canada.

D.H. Waddington. Waddington Consulting, 3230 Yonge St, Unit #1918, Toronto, ON, M4N 3P6, Canada

S. Page. Lassonde Institute of Mining, University of Toronto, 170 College Street, Toronto, ON, M5S 3E3, Canada

J. Yang. United Research Systems, Suite 240, 6025 Eleventh Street SE, Calgary, AB, T2H 2Z2, Canada

Corresponding author: T. Rivers (trivers@mun.ca)

Abstract: The well mapped western part of the Ottawa River Gneiss Complex (ORGC; new name), a large metamorphic core complex, hosts a system of gently-plunging cross-folds of outcrop to regional scale situated in the ductile detachment zone between the lower grade cover and high-grade core of the complex. The cross-folds are buckle structures that deform the attenuated gneissic layering, plunge parallel to the regional elongation lineation, range from upright to recumbent, and exhibit distinctive hinge-parallel elongation lineations, all features of pure-shear-dominated ductile transtension. Our L-S fabric data are consistent with kinematic modelling that predicts progressive constrictional strain in the hinge zones of transtensional folds, and with dynamic modelling that predicts rotation of flattened fold limbs into moderately dipping attitudes. On the basis of petrologic data, we show that cross-folding post-dated the peak Ottawa metamorphism and took place during retrogression and exhumation of the thrust-sheet stack.

The cross-folds form an inclined system that is principally developed in retrograde melt-weakened amphibolite-facies rocks, with the transtensional origin implying that exhumation and retrogression of the high-grade core of the complex took place in an oblique extensional setting. A transtensional origin for the cross-folds removes the need to appeal to orogen-parallel regional shortening, an implausible requirement of previous interpretations, and is compatible with data indicating that much of the visible fabric and structure of amphibolite-facies domains of the ORGC developed during post-peak exhumation, retrogression and gravitational collapse of the thrust-sheet stack. [235 words]

Key words: gneiss complex, cross-fold, constrictional strain, gravitational collapse, metamorphic core complex

Introduction

Cross-folds, also called transverse folds, occur in collisional orogens and have axial traces that make high angles with the traces of major thrust faults (Heim, 1921; Kvale 1953; Fyson 1964; Steltenpohl and Bartley 1988; Sati and Rautela 1993; Aller and Gallastegui 1995; Fisher and Olsen 2004). Where they occur in high-grade gneiss complexes (Culshaw et al. 2004; Van Reenen et al. 2004), they have been interpreted as late-stage structures superimposed on orogen-parallel fold nappes and ductile thrusts that formed during crustal thickening. In both settings they are multi-order structures that occur over a wide range of scales, but most attention has been paid to map-scale examples with hinge lengths of 1-50 km and amplitudes and wavelengths of 1-10 km. Various tectonic, kinematic or geodynamic scenarios for cross-folding have been advanced (e.g., Alsop et al. 1996, p. 783), many calling upon regionally heterogeneous, ductile high-strain deformation, but since their inferred timing ranges from during to after peak metamorphism and crustal thickening it appears there may be no single cause. In the present paper, we integrate outcrop and regional observations relating to a major system of late tectonic cross-folds with distinctive L-dominant fabrics in their hinge zones and provide a new interpretation for their origin. The cross-folds are developed in the Ottawa River Gneiss Complex (former Central Gneiss Belt; Fig. 1; see Appendix 1 for discussion), a large granulite- to upper amphibolite-facies metamorphic core complex with dimensions of >250 x 400 km in the south-western Grenville Province of Ontario and western Quebec.

Previous interpretations of cross-folds in the Ottawa River Gneiss Complex

Despite progress in understanding the tectonic evolution of the Ottawa River Gneiss Complex (e.g., Culshaw et al. 1997; Wodicka et al. 2000; Rivers 2008, 2012; Rivers et al. 2012;

Rivers and Schwerdtner 2015), the significance of a prominent system of late multi-order cross-folds remains controversial. Figure 2 shows the regional distribution of the cross-fold system, which is characterized by upright folds with moderately plunging hinge lines from 10-100 km in length (data from Hewitt 1967; Waddington 1973; Davidson and Morgan 1981; Davidson et al. 1982; Culshaw et al. 1983, 2004; Easton 1992; Ontario Geological Survey 1992; Lumbers 1996). Related cm- to m-scale cross-folds also form prominent structures in outcrop, but orientations of their axial surfaces are more varied and include recumbent structures (Fig. 3). Termed set-2 folds by Schwerdtner (1987), they refold the gneissic fabric, are heterogeneously distributed, trend parallel to the regional elongation lineation, and are characterized by hinge-parallel elongation lineations defined by statistically aligned mineral grains and polycrystalline aggregates.

As alluded to above, the origin of the cross-folds in the ORGC is a long-standing issue. Although most workers have favoured buckling as the folding mechanism (Schwerdtner and van Berkel 1991; Gower 1992; Culshaw 2005; Schwerdtner and Klemens 2008), possibly preceded by large amounts of layer-parallel shortening, their tectonic context has not been unequivocally established and three conflicting scenarios have been proposed. Schwerdtner (1987) attributed cross-folding to NW-directed convergent bulk flow towards the 'Big Bend' in the trace of the Grenville Front (see Fig.1), with or without a component of dextral shear. This was criticized by Gower (1992) who proposed cross-folding took place during crustal thickening in an early phase of NE-directed sinistral transpression. Finally, in a third scenario Culshaw et al. (1994, 1997) ascribed the cross-folds to ductile reactivation of the thrust-sheet stack by pervasive extensional shearing parallel to the SE-plunging dip line of the regional foliation.

A problem common to all three interpretations, however, is that none provides a plausible geodynamic explanation for the implied component of orogen-parallel shortening required for

regional buckling at high angle to the orogenic front. In this context, Schwerdtner et al. (2014) and Rivers and Schwerdtner (2015) briefly mentioned another possibility we investigate in more detail here – that the cross-folds formed in an episode of late-orogenic, pure-shear-dominated, ductile transtension accompanying exhumation of the early-Ottawan thrust-sheet stack. To do this, we briefly discuss the tectonic context of the ORGC as a large metamorphic core complex in a collapsing large hot orogen, review modes of formation of ductile pure L, L>S, and L>>S (prolate) fabrics, and summarise results of recent modelling of transtension, including the relative magnitudes of its simple-shear and pure-shear parts and their influence on the orientation and geometry of transtensional folds. We then present new structural data and metamorphic observations from the southwest ORGC supporting the hypothesis that the cross-folds formed in a transtensional setting during retrogression and exhumation of the hot mid crust in a large metamorphic core complex beneath a collapsing orogenic plateau.

Large hot orogens

Numerical modelling and empirical data both support the deduction that the Grenville Province is a remnant of a large hot orogen of Himalaya-Tibet scale with an orogenic plateau underlain by double-thickness crust in its hinterland (Beaumont et al. 2006; Rivers et al. 2012). In the Grenville Province, this hinterland consists of an exhumed stack of high-grade thrust sheets of mid- and lower-crustal origin (Davidson et al. 1982; Culshaw et al. 1983, 1997; Davidson 1984a, 1984b; Rivers et al. 1989, 2002; Easton 1992; White et al. 1994, 2000; Carr et al. 2000; Ketchum and Davidson 2000; Martignole et al. 2000; Tollo et al. 2004; Slagstad et al. 2004; Hynes and Rivers 2010) that is structurally juxtaposed against parts of the upper orogenic crust (Rivers 2008, 2012). The present distribution of crustal levels suggests the juxtaposition

was a result of gravitationally-driven extensional collapse (Culshaw et al. 1994; Martignole and Calvert 1996; Klemens and Schwerdtner 1997; Ketchum et al. 1998; Schwerdtner et al. 2005, 2014; Rivers 2008; McLelland et al. 2010; McLelland and Selleck 2011; Jamieson and Beaumont 2011), such that the orogenic mid-crust is exposed in a series of large metamorphic core complexes (Rivers 2012; Rivers et al. 2012; Rivers and Schwerdtner 2015; Soucy La Roche et al. 2015). In this context, the long time-span of the Grenvillian Orogeny, from ca. 1090 to 980 Ma (Carr et al. 2000; Rivers et al. 2012), the widespread high grade of metamorphism in the exhumed Grenvillian middle and lower crust, and the collapse of the orogenic plateau are all compatible with a large hot orogen.

L>S shape fabrics and constrictional finite strain

Since its introduction by Flinn (1965), the L-S concept has been widely used to characterize shape fabrics in ductile rocks, despite some evidence that (a) visual assessment may result in over-estimation of the L-component, (b) the shape fabric depends on whether the rock was initially mechanically isotropic (e.g., megacrystic granite) or anisotropic (e.g., layered gneisses), and (c) primary shape fabrics and viscosity contrasts between inclusions and matrix render direct correlation between L-S shape fabric and the geometry of a finite-strain ellipsoid problematic (Schwerdtner et al. 1977). Nonetheless, clusters of L-dominant prolate shape fabrics are widely regarded to be indicative of an important constrictional component of strain, especially where the interpretation is consistent with other information (e.g., Schwerdtner et al. 1983, 2010a; Zulauf 1997; Krabbendam and Dewey 1998; Schwerdtner and Cote 2001; Van Kranendonk et al. 2009; Leslie et al. 2010; Sullivan 2013).

Although, in theory, the finite strain and related L-S shape of a ductile tectonite may have been acquired along any one of an infinite number of possible paths, geologically realistic paths are probably relatively simple and three plausible regimes for the formation of L-dominant tectonites from initially spherical objects in a single phase of deformation were discussed by Sullivan (2013) and Chen (2014). These are: (i) constriction (Kobberger and Zulauf 1995; Zulauf and Zulauf 2005), (ii) collinear simple shear along two or more glide planes, a deformation mechanism characteristic of lateral ramps in ductile thrust systems (Leslie et al. 2010) and known as ‘cross slip’ in crystal plasticity (Cottrell 1953; Friedel 1957), and (iii) non-coaxial finite strain during ductile transtension (Sanderson and Marchini 1984; Fossen et al. 1994; Dewey et al. 1998). In addition, it has been suggested that prolate fabrics may form by the superposition of two non-constrictional finite strains (e.g., Ramsay 1967, p. 165; Schwerdtner and Gapais 1983; Schwerdtner 1987; Schwerdtner et al. 2010b). As noted above, we deduce a transtensional origin for the $L > S$ prolate fabrics in the hinge zones of cross-folds in the Ottawa River Gneiss Belt and provide support for this interpretation in the remainder of this paper.

The concept of ductile transtension

The terms transpression and transtension were introduced by Harland (1971) to denote the progressive bulk deformation in marginal zones of obliquely converging or diverging tectonic plates. Sanderson and Marchini (1984) extended their use to ductile zones bounded by smaller structural discontinuities within tectonic plates and specified the cumulative bulk deformation by means of asymmetric Cartesian tensors. Unless further generalized to include components of triaxial strain (e.g., Tikoff and Teyssier 1994; Jones et al. 1997, Jones and Holdsworth 1998; Fossen and Tikoff 1998), these tensors have a simple-shear part that distorts

and rotates material lines in the horizontal plane, and a pure-shear part that distorts material lines in the vertical plane perpendicular to the simple-shear direction. In order to depict the overall geometric effects of ductile transtension, most workers have used simple ‘cartoon’ models such as that shown in Figure 4. Typically, these consist of two rigid vertically-sided blocks bordering an intermediate region of horizontally-layered isotropic material that undergoes homogeneous ductile deformation, illustrated by folding of passive planar markers. More complex models with inclined non-parallel or non-planar boundaries of the ductile region have been considered (e.g., Dewey et al. 1998), but even in these cases the ductile deformation accords with tensors that are referred to horizontal and vertical Cartesian axes.

Cartoon models assuming homogeneous ductile deformation and different arrays of tensor components have been used to illustrate the systematic variations in the shape and orientation of the strain ellipsoid with divergence angle (Tikoff and Teyssier 1994; Fossen et al. 1994, 2013; Dewey et al. 1998; Fossen and Tikoff 1998). However, the underlying assumption that ductile transtensional strain is homogeneous is known to be problematic on both mechanical and dynamic grounds (Schwerdtner 1989; Robin and Cruden 1994; Dutton 1997; Dewey et al. 1998, their Fig. 1). Thus, although cartoon models with homogeneously-strained ductile regions are useful to illustrate general features of transtension, it is inappropriate to rigorously apply results derived from them to natural transtension zones in which strain was likely heterogeneous. Similarly, distinction between natural examples of simple-shear-dominated transtension, in which Z (short axis of strain ellipsoid) is horizontal, and pure-shear-dominated transtension in which Z is vertical, that are based solely on inferences from models using an assumption of homogeneous strain may also be unjustified.

Natural examples and models of transtensional fold systems

Well described Phanerozoic multi-order cross-fold systems deduced to have developed in a transtensional setting include those from orogens that have undergone extensional collapse, with the Western Gneiss Region of Norway being a classic example (Fossen 1992; Krabbendam and Dewey 1998; Osmundsen and Andersen 2001; Fossen et al. 2013). The Western Gneiss Region is a large Caledonian metamorphic core complex, in which high-pressure and ultra-high pressure (HP and UHP) gneisses are exhumed along several ductile and brittle detachments from beneath a cover of low-pressure (LP) metamorphic rocks. In the HP and UHP rocks below the main detachment, the hinge lines of cross-folds are parallel to elongation lineations defined by $L \gg S$ and $L > S$ shape fabrics, attesting to pervasive constriction-dominant strain (Krabbendam and Dewey 1998; Fossen et al. 2013). Above the main detachment, LP rocks are also deformed by the cross-folds, which extend structurally up-section into the fill of a transtensional basin that opened as the HP and UHP rocks were exhumed from mantle depths (Osmundsen and Andersen 2001). From this arrangement, Krabbendam and Dewey (1998) concluded that the Western Gneiss Region provided compelling evidence for the important role of transtension in promoting the exhumation of HP and UHP metamorphic rocks.

Turning to the results of modelling, Fossen et al. (2013) employed numerical kinematic (strain) analysis to study both the amplification and reorientation of asperities in originally horizontal passive layers during homogeneous progressive transtensional strain, and the three-dimensional geometry of the resulting folds. They reached four important conclusions: (1) transtensional folds initiate as upright structures, regardless of the proportions of simple and pure shear; (2) final amplitudes of the transtensional folds are smallest where the divergence angle is close to 90° (i.e., the simple-shear part of the deformation is small); (3) the fold-hinge lines

rotate progressively towards the divergence vector and the long axis of the bulk strain ellipsoid (X); and (4) transtensional folds are characterized by $L > S$, $L \gg S$ and locally pure L fabrics and hinge-parallel elongation lineations.

Fossen et al. (2013) also considered the distortion of inclined or vertical passive multilayers with a strike approximately parallel to the divergence vector. They showed that asperities in such multilayers were transformed into trains of inclined to recumbent folds replete with hinge-parallel elongation lineations. Moreover, they also demonstrated how, at an advanced stage of high-amplitude transtensional folding of initially subhorizontal layers, fold limbs that attain steep attitudes deform in a similar manner to originally subvertical layers (their Figs. 11, 12).

With respect to dynamic modelling, the analogue experiments of Venkat-Ramani and Tikoff (2002) were the first to reproduce pure-shear-dominated transtensional folds. Using horizontal mechanically-weak multilayers, they illustrated both the rotation of the fold hinge lines into parallelism with X of the bulk strain ellipsoid and ductile extension parallel to them. Analogue simulation of transtensional buckling of mechanically strong multilayers is confined to the limiting case of purely constrictional bulk strain. For this case, the experiments by Kobberger and Zulauf (1995) produced upright buckle folds accompanied by extensional pull-apart zones perpendicular to the fold hinge lines in competent layers, and Zulauf and Zulauf (2005) showed that successive episodes of coaxial constrictional strain resulted in complex structures due to the superposition of boudins and buckle folds.

Role of original structures on the style of transtensional folds

Inasmuch as the Ottawa River Gneiss Complex is developed from a ductile thrust sheet stack, it is appropriate to consider the role of the original thrust structures on its later transtensional evolution. For instance, Hatcher and Hooper (1992) concluded that ductile mid-crustal thickening gave rise to km-scale fold nappes with sheath geometry and strongly-attenuated overturned limbs, the latter evolving into narrow shear zones and discontinuity surfaces they termed crystalline thrusts (see also Hatcher 1995). Well-preserved examples of such structures have been described from the Inner Piedmont in the southern Appalachians (Mersch et al. 2005; Hatcher 2010), and from the Caledonian thrust-sheet stack in northern Scotland (Powell and Phillips 1985; Harris and Johnson 1991), both of which escaped later reworking by gravitational collapse. Comparable structures were probably also present in the Ottawa River Gneiss Complex, but the original architecture of ductile thrust-sheet stacks in large hot orogens is difficult to reconstruct because of the superimposition of later ductile extensional structures, including cross-folds, during the orogenic collapse stage.

The structural architecture of regions affected by ductile cross-folding has been analyzed in several collisional orogens (e.g., Cantabrian zone of the Variscan Orogen, Aller and Gallastegui 1995; Grenville inliers in the Appalachians, Fisher and Olsen 2004), and compared with models of superimposed buckling and passive folding (e.g., Ramsay 1967; Ghosh and Ramberg 1968; Ghosh et al. 1993). Especially on the km-scale, the map patterns of ductile cross-folding are characterized by bifurcating axial traces (conjugate non-cylindrical structures) and oval domes and basins (doubly-plunging folds), compatible with model results.

In this paper, we apply results of the three-dimensional strain modelling of Fossen et al. (2013) and two-dimensional dynamic buckling experiments of Schmalholz and Schmid (2012) to interpret the cross-folds of the ORGC (three-dimensional models of buckling have yet to be

produced). However as intimated in the previous paragraphs, a severe limitation of all such idealised models with respect to their application to nature is that the original geometry of thrust-sheet stacks is significantly more complex than the passive homogeneous planar horizontal multilayers employed by modellers.

Cross-folds in the Ottawa River Gneiss Complex

Distribution, geometric style, groups, and age

Distribution: The outcrop pattern of amphibolite-facies rocks in the well mapped western ORGC is dominated by a regional system of cross-folds with principally NW-striking axial surfaces, and a minority with more N- or W-striking orientations (Fig. 2b; e.g., Hewitt 1967; Waddington 1973; Lumbers 1975; Schwerdtner et al. 1977; Schwerdtner and Waddington 1978; Davidson and Morgan 1981; Davidson et al. 1982; Culshaw et al. 1983, 2004; van Berkel and Schwerdtner 1986; Lumbers and Vertolli 2000a, b; Lumbers et al. 2000a, b). Apparently similar structures are also present on geological maps of less studied portions of the ORGC, both in the northeast part of Figure 2a (Lumbers 1971, 1976a, 1976b; Ontario Geological Survey 1992; Davidson 1996; Van De Kerckhove 2015) and farther east in Québec (Rive 1973a, 1973b). However, cross-folds are not a typical feature of the Algonquin and Parry Sound domains of the ORGC, both largely composed of granulite-facies rocks (e.g., Culshaw et al. 1997; 2010), and they are also absent from the overlying upper-crustal (greenschist- to amphibolite-facies) Composite Arc Belt (Ontario Geological Survey 1991, 1992), except at the base of the Composite Arc Belt boundary zone (e.g., near Barry's Bay; Easton 2001; see Fig. 1 for location). This distribution indicates cross-folding was principally restricted to ductile amphibolite-facies rocks in the ORGC, and cross-folds are especially well-developed in structural levels 1 and 3

(Fig. 2c; stacking nomenclature of Culshaw et al. 1997; Ketchum and Davidson 2000; Jamieson and Beaumont 2011). Figure 2d shows the ORGC recast as a giant metamorphic core complex resulting from orogenic collapse, the tectonic setting relevant to their formation (after Rivers and Schwerdtner 2015). In this context it can be seen that cross-folds are mostly developed in the detachment zone between the top of the granulite-facies core (Algonquin domain and structurally equivalent subdomains) and the base of the cover (Composite Arc Belt and Parry Sound domain).

Size and geometric style: The largest cross-folds in the western ORGC occur in the Muskoka domain, where the hinge lines of the Moon River and Seguin synforms are >40 km long and define folds with wavelengths of ≥ 10 km and amplitudes of ~ 7 km (MRS and SS, Fig. 2b). As noted, they are situated in the detachment zone of the core complex (Fig. 2d), and the two synforms are separated by the doubly-plunging Bracebridge antiform in the underlying Rosseau subdomain at the top of the core (BA, Fig. 2b; Schwerdtner et al. 1977). A cross-section of the Moon River synform (Fig. 5), the best studied first-order cross-fold in the ORGC (Waddington 1973; van Berkel and Schwerdtner 1986; Schwerdtner and van Berkel 1991; Gower 1992; Lumbers and Vertolli 2000a; Culshaw et al. 2004, 2010), displays the steep axial surface and large inter-limb angle characteristic of these structures. It is bounded on its southwest flank by the Twelve Mile Bay shear zone (Fig. 2; Davidson 1984a, 1984b; Culshaw et al. 2010; Marsh et al. 2013), but marginal shear zones are not typical of most cross-folds.

Where the Moon River and Seguin synforms abut against the Parry Sound domain, their hinge lines decompose into smaller S- and E-trending folds that converge down plunge into the main SE-plunging synforms. These smaller structures are fold abutment structures (Schwerdtner and van Berkel 1991), in which the folded ductile amphibolite-facies, predominantly felsic

straight gneiss is in contact with more competent granulite-facies mafic gneiss and the contact has undergone incipient cusped–lobate folding.

Cross-folds are also abundant in the Britt domain farther northwest in the ORGC, where they form a moderately SE-plunging system in which individual fold hinge lines are continuous for >20 km (>80 km in the case of the Britt-Nobel antiform; BNA Fig. 2b). However, their wavelengths and amplitudes are smaller than those of the Moon River and Seguin synforms, being generally <5 km. The map patterns of cross-folds in the two areas also differ. The down-plunge continuity of straight gneiss in the wide Moon River and Seguin synforms in the Muskoka domain implies fabric transposition was important and indicates the plunge of these structures is close to the dip of the erosion surface. On the other hand, cross-folds in the Britt domain generate V-shaped outcrop patterns, suggesting transposition was less significant and that they plunge more steeply southeast than the dip of the erosion surface. We suspect the hinge lines of cross-folds in the Britt domain were steepened after formation, compatible with other evidence for Rigolet (~1000 Ma) deformation in the vicinity of the Grenville Front Tectonic Zone.

Cross-fold groups: It can be seen in Figure 2 that cross-folding affects all levels of the ORGC, from the Britt domain at its base to the Muskoka domain at its upper contact with the Composite Arc Belt. However, despite individual hinge-lines of almost 100 km length, no single cross-fold spans the entire ORGC; rather three spatially distinct groups can be identified (Fig. 2b) that are limited to different parts of the gneiss complex and terminate at domain boundaries. Specifically, some cross-folds (termed Group A) in the southeast are limited to the Muskoka domain and the uppermost part of the underlying Algonquin domain (i.e., the large NW-trending Moon River and Seguin synforms in the Muskoka domain, and the intervening NW-trending

Bracebridge antiform), with several smaller, mostly NNW-trending unnamed cross-folds in the Rosseau and Go Home subdomains comprising a distinct subgroup. Group B cross-folds in the northwest ORGC are NW-trending and occur within the amphibolite-facies Britt domain and either terminate within that domain or cross the Allochthon Boundary and terminate at the Parry Sound domain (several examples, including the Britt-Nobel antiform). Group C cross-folds include two antiformal structures that deform the granulite-facies Parry Sound domain and its amphibolite-facies footwall (i.e., the ~50 km long, WNW-trending Wawashkesh–Ahmic Lakes antiform and the shorter Golden Valley antiform; WALA and GVA, Fig. 2b). They are distinctive in terms of their WNW-trend, lack of obvious intervening synforms, and wide spacing, features possibly related to the competent rheology of the Parry Sound domain, which is largely composed of granulite-facies mafic rocks.

Spatial separation of cross-fold groups A and B occurs because of fold abutment structures and the termination of cross-fold hinge lines at the boundaries of the Parry Sound domain, which appears to have determined the km-scale geometry of strain during cross-folding. In addition to their mutually exclusive spatial distributions, cross-folds in Groups A and B appear to have developed under slightly different post-peak metamorphic conditions. The axial surfaces of some Group A cross-folds are decorated with syn-folding leucosome (e.g., Fig. 3a), compatible with other evidence for at least local suprasolidus conditions (Rivers and Schwerdtner 2015), whereas those in Group B formed after leucosome had crystallized (Culshaw et al. 1991, 1994), implying formation under subsolidus conditions. Both groups are locally cut by thin undeformed pegmatite dykes (e.g., Fig. 3c) suggesting the presence of a suprasolidus zone at greater depth.

Timing and conditions of cross-folding: In terms of relative timing, the cross-folds clearly deform the gneissic fabric of the ORGC, locally forming Type 2 fold interference patterns (Fig. 5; Schwerdtner 1987; Schwerdtner and van Berkel 1991; Gower 1992). They are deduced to have developed after the metamorphic peak during dynamic retrogression in the detachment zone of the metamorphic core complex (Rivers and Schwerdtner 2015; see also below). The lack of mutual interference of the three cross-fold groups, despite their varied style and orientation, is compatible with experimental (e.g., Ghosh et al. 1995) and empirical (e.g., Aller and Gallastegui 1995; Fisher and Olsen 2004) studies that folds of varied orientation and style can develop in a single phase of deformation when superimposed on previously strained multilayers with complex structure and varied rheology.

Absolute constraints on the time of cross-folding remain imprecise. With respect to a maximum age bound, Rivers and Schwerdtner (2015) inferred that peak metamorphism in the Muskoka domain was early Ottawa (~1085 Ma), with suprasolidus retrogression and cross-folding occurring in mid- to late-Ottawa time, perhaps initiating as early as ~1060 Ma. On the other hand, cross-folding in the Britt and Shawanaga domains post-dated the Ottawa metamorphic peak (dated at 1070-1030 Ma there); in addition, since some cross-folds deform the Allochthon Boundary, they must have formed at the time of, or after, its reworking as an extensional structure at ~1020 Ma (Culshaw et al. 1997; Ketchum et al. 1998). A minimum age bound for cross-folding in the Britt domain is given by the ~1.0 Ga Rigolet-related thrusting in the Grenville Front Tectonic Zone (Haggart et al. 1993; Dudás et al. 1994; Bethune 1997), and by the ~990 Ma age of late undeformed pegmatites (Corrigan et al. 1994; Bussy et al. 1995).

Evaluation of cross-folds as transtensional structures

In this section, we evaluate whether cross-folds in the ORGC conform to model transtensional folds developed in an oblique extensional setting. The geometry, km-scale variations, and associated mesoscopic structures of cross-folds have been studied systematically in the Moon River synform, Wawashkesh-Ahmic Lakes antiform, Britt-Nobel antiform and a few other examples (Fig. 2b; Waddington 1973; Janes 1974; Bennett 1975; Schwerdtner et al. 1977; Klemens 1996; Schwerdtner and Klemens 2008; Schwerdtner et al. 2010a), and in the following text we examine features of representative cross-folds using five diagnostic criteria for a transtensional origin.

Buckle origin of multi-order cross-folds

Systematic variation in M-S-Z style of minor folds in multi-order fold systems is a classic shear-sense criterion with which to evaluate the vergence of regional fold systems (e.g., Ramberg 1963a) and by extension the role of buckling in mechanically anisotropic multilayer sequences such as the gneisses comprising the ORGC. Adopting a measurement protocol to eliminate the effects of plunge reversals along strike, we have used an extension of the vergence principle known as tangential unit shear (Schwerdtner et al. 2005, 2010b) to assess the shear sense on the enveloping surfaces of map-scale cross-folds at three widely separated localities in Figure 2. These are: (i) the *Britt-Nobel antiform* (Fig. 6a) that transects the Shawanaga orthogneiss unit in the immediate hanging wall of the Allochthon Boundary (Culshaw et al. 1994; Dickin et al. 2008); (ii) the conjugate *Grundy Lake synform and Pakesley antiform* (Figs. 6b, 7b), which deform strongly foliated metaplutonic rocks of the Key Harbour gneiss association in the Britt domain (Culshaw et al. 1997, 2004, 2006); and (iii) the central portion of the *Moon River synform* in the Muskoka domain (Fig. 8), where the high-strain orthogneiss fabric is deformed by the first-order Moon River synform, the second-order Inner Moon River and Torrence synforms,

and third-order outcrop-scale folds. In all three locations, on scales ranging from a few metres to several kilometres, the variations in vergence and shear sense on the enveloping surfaces conform closely to the ideal pattern for buckling, thereby supporting a buckle origin for cross-folding (a conclusion also supported by data for the Wawashkesh-Ahmic Lakes antiform; Klemens 1996; Schwerdtner and Klemens 2008).

Determination of the amount of shortening accommodated by buckling, i.e., 'unfolding' the cross-folds and estimating of the amount of layer-parallel shortening, is not presently possible. The difficulty is highlighted by the fact that although the apical angles of the km-scale folds are large (Fig. 5), those of the m-scale folds are highly variable (Fig. 3), indicating the scale variation of mechanical heterogeneity and deformation.

Quasi-cylindrical and non-cylindrical cross-folds

Analogue and numerical modelling of oblique ductile extension in idealised horizontal planar multilayer sequences results in the formation of transtensional folds with horizontal, cylindrical hinge lines (e.g., Venkat-Ramani and Tikoff 2002; Fossen et al. 2013). However, in the ORGC, the hinge lines of most cross-folds are gently plunging ($\leq 30^\circ$) and only quasi-cylindrical, with some m-scale examples being markedly non-cylindrical (i.e., both cross-sectional geometry and attitude of hinge-line segments change with position). In an extreme case, a 10-20 m wide zone in the Moon River synform is characterised by upright to steeply inclined mesoscopic folds in which the amount of plunge of the hinge lines changes by $>60^\circ$ (Fig. 9c-e), leading to the interpretation it is a sub-horizontal sheath fold (Schwerdtner and van Berkel 1991). Moreover, detailed structural analyses of km-scale segments of the Moon River synform and Wawashkesh-Ahmic Lakes antiform have demonstrated that the regional cross-

folds are also non-cylindrical (Fig. 8b; Waddington 1973, Schwerdtner et al. 1977; Schwerdtner and van Berkel 1991; Klemens 1996; Schwerdtner and Klemens 2012). The non-cylindricity has been attributed to the original curvature of early-Ottawan thrust sheets (Davidson 1984a; Gower 1992; Wodicka et al. 2000), rheological heterogeneity of the folded multilayers, or later superposition of km-scale boudins and pinch-and-swell structures (Schwerdtner et al. 2014). All three may account for the disharmonic style of the Wawashkesh-Ahmic Lakes antiform. In the case of the Moon River and Seguin synforms, non-cylindrical cusate-lobate folding of the amphibolite-facies contact with the Parry Sound domain, discussed previously, is also compatible with observations that the Moon River synform folds the retrogressed (amphibolite-facies) margin of the Parry Sound allochthon (Culshaw et al. 1997; Wodicka et al. 2000; Jamieson et al. 2007; Marsh et al. 2013). In summary, we consider the variable plunge and widespread non-cylindricity of cross-folds to be a result of their superimposition on a dipping, mechanically heterogeneous multilayer sequence with pre-existing structures rather than an intrinsic aspect of the cross-folding process.

Cross-fold hinge lines parallel to the regional elongation direction

Despite widespread annealing, the regional elongation lineation is a readily observed fabric element in the ORGC, typically being defined by streaky mineral aggregates, leucosomes, quartz ribbons, and grain-size reduction of feldspar augen. In the southeast (Muskoka domain), the presence of peak anhydrous minerals such as *Opx* and *Cpx* partially rimmed and replaced by post-peak hydrous minerals such as *Hbl* and *Bt* (mineral abbreviations after Whitney and Evans 2010) implies the lineation is post-peak and formed during hydrous retrogression of peak granulite-facies rocks (Rivers and Schwerdtner 2015), whereas in the northwest (Britt domain)

the fabric in metapelitic rocks is deduced to have formed in the stability field of the upper amphibolite-facies sub-assemblage $Ky-Sil-Kfs$ after leucosome crystallization (Culshaw et al. 1994). Additional evidence the L-fabric developed after the metamorphic peak, rather than during thrusting as previously proposed, is given in Appendix 1 and Rivers and Schwerdtner (2015).

The elongation lineation plunges gently towards the southeast in much of the ORGC, hence is effectively parallel to traces of the axial surfaces of the cross-folds (Culshaw et al. 1994; their Figs. 7, 10), and we have documented a similar pattern on the scale of the smaller parasitic cross-folds. For instance, Figure 7a shows our data for the attitudes of fold-hinge lines and elongation lineations in the vicinity of the SE-plunging Grundy Lake synform and Pakesley antiform in the Britt domain. Similarly, in the central Moon River synform the elongation lineations are subhorizontal or plunge gently southeast, tracking the orientations of nearby cross-folds even where there are plunge reversals (Figs. 8, 9a, f; Waddington 1973; van Berkel and Schwerdtner 1986; Schwerdtner 1987).

Cross-fold development in retrograde attenuated layering during decompression

In the Muskoka domain, the layering deformed by the cross-folds is typically a retrogressed leucosome-rich straight-gneiss fabric defined by alternations of pink or grey granitoid gneiss with thin amphibolite layers (heterogeneous gneiss) and more homogeneous orthogneiss (Fig. 10a-c; Rivers and Schwerdtner 2015). Unequivocal evidence for high strain, including narrow mylonite zones, symmetrical and asymmetrical porphyroclasts and grain aggregates etc. can be found locally, but in general the fabric exhibits an equigranular microstructure with only weak crystallographic expression, a result of thorough annealing.

Nonetheless, the widespread macroscopic evidence for pinch-and-swell and boudinage structures in retrogressed rocks attests to important post-peak attenuation (Fig. 10d-f). In terms of metamorphic grade, the common presence of coarse-grained granulite-facies inclusions and boudins surrounded and partially replaced by the finer-grained high-strain amphibolite-facies fabric, with abundant leucosome in boudin necks (Figs. 10c, f), provides convincing evidence that the straight-gneiss fabric is retrograde, a product of amphibolite-facies retrogression and associated partial melting of granulite-facies precursors (Rivers and Schwerdtner 2015). The post-peak origin is further supported by the presence of *An*-rich plagioclase aggregates replacing former *Grs*-rich garnet in mafic rocks (Fig. 3f), which is commonly also associated with replacement of *Cpx* by *Hbl*, thereby linking fabric formation to post-peak decompression, exhumation and retrogression of the peak (HP-) granulite-facies rocks.

Essentially comparable relationships have been reported from the Britt domain, where the cross-folds deform straight gneiss in which late ductile extensional strain, comprising both discontinuous shear zones and distributed shear in the intervening rocks, occurred under post-peak upper amphibolite-facies conditions (Culshaw et al. 1994). However, *Pl* aggregates replacing *Grt* in mafic rocks were not reported, suggesting the peak metamorphic pressure was lower there.

L>S fabric parallel to hinge lines of cross-folds

To address the issue of L-S fabric style in the hinge zones of km-scale cross-folds, we present strain and anisotropy of magnetic susceptibility (AMS) data from the hinge zone of the Inner Moon River synform. At several localities, including the MacTier interchange between Highways 69 and 400 and two places in central Bala (Fig. 8), the amphibolite-facies layering is

deformed into upright mesoscopic cross-folds with subhorizontal enveloping surfaces (Fig. 9b-c). Most of these folds are gently plunging and quasi-cylindrical, but some appear to have sheath forms as described previously. In central Bala, the hinge zone of the Inner Moon River synform locally deforms a layer of mafic gneiss with strongly aligned, almost perfectly prolate, plagioclase-rich pseudomorphs after large garnet porphyroblasts (Fig. 9f). Despite their strong alignment, the X/Z axial ratios of the distorted pseudomorphs range from 1.5 to 9.0, suggesting a span of finite strain from relatively low to moderately high, and axial ratios of neighbouring pseudomorphs differ by up to 5.5 (based on data in Schwerdtner et al. 1974, their Table 1). This range is incompatible with coeval homogeneous deformation of randomly oriented dodecahedral passive markers. Combined with similar observations elsewhere in the Muskoka domain, it implies that garnet breakdown was spatially and temporally heterogeneous, staggering the onset of ductile deformation and creating a record of the progressive strain. Accordingly, we deduce that the variation in principal axial ratios among neighbouring pseudomorphs preserves a record of *progressive constrictional deformation*, a characteristic of transtensional folding (Dewey et al. 1998; Fossen et al. 2013). Moreover as noted, petrological considerations indicate the pseudomorphs provide a record of garnet breakdown during post-peak decompression at high temperature, thereby linking the constrictional deformation to exhumation and orogenic collapse (Rivers and Schwerdtner 2015).

This result can be compared with hornblende orientation data and AMS analyses of cores from a hornblende-bearing gneiss unit (Schwerdtner and Waddington 1978, p. 208-209). The data for a fold in central Bala with a sub-horizontal hinge line parallel to the regional Inner Moon River synform are shown in Figure 11a, in which it can be seen that the hornblende lineation accounts for both the L>S character of the mineral-grain and magnetic fabrics (maximum

susceptibility direction and the girdle containing the intermediate and minimum directions), suggesting all were the result of a single act of constrictional deformation. Data for two moderately to steeply NW-plunging, upright mesoscopic cross-folds in Bala and at the MacTier interchange are shown in Figure 11b-c. For the Bala sample, the maximum susceptibility direction and hornblende-grain lineations are effectively parallel and subhorizontal, but nearly perpendicular to the plunging fold-hinge line (Fig. 11b), whereas in the MacTier sample the hornblende-grain lineation varies from parallel (X) to oblique (X') to the fold-hinge line, and the magnetic lineation is almost horizontal and parallel to X', making an angle of $\sim 40^\circ$ with the plunging fold-hinge line (Fig. 11c). Two interpretations of these relationships are feasible: (i) the plunging hinge lines are parts of metre-scale sheath forms that developed at the same time as the subhorizontal hinge line and L-fabric of the Inner Moon River synform, or (ii) the magnetic and hornblende-grain lineations were superimposed on a pre-existing mesoscopic fold. In earlier papers, the data were interpreted in terms of superposition (Schwerdtner and Waddington 1978; Schwerdtner 1987, 1988), but we now conclude on both structural and petrological grounds that all the fabric elements developed together in a single phase of ductile deformation during retrograde *Hbl* crystallization. Thus, on a regional scale both the subhorizontal elongation directions defined by the mineral-grain and magnetic lineations, and the prolate shapes of plagioclase-rich pseudomorphs after garnet, are effectively parallel to the regional hinge line of the Inner Moon River synform (Waddington 1973, p. 46-49) and related to the strain accumulated during regional cross-folding. The hinge lines of the local sheath forms, on the other hand, are more steeply plunging than their mineral fabrics and the regional structure. However, since the fabric mineralogy and orientation is identical we now deduce that the folds formed at the same time as the regional structure and that their steep plunges and sheath-like

forms are a result of fold nucleation on steeply dipping surfaces, such as the hinge zones of early-Ottawan recumbent folds. This accommodates the occurrence of dip variations within the thrust stack prior to cross-folding, a feature not considered in the original interpretation.

L-S fabric variation around hinge lines of cross-folds

There is a systematic change in shape fabric between the hinge zones and limbs of cross-folds that is especially evident where relatively homogeneous isotropic lithologies such as megacrystic orthogneiss, metagabbro and mafic gneiss are present (e.g., Schwerdtner et al. 2010b, their Figs. 5a-b). Where affected by cross-folding, the hinge zones exhibit distinctive L- or L>S shape fabrics defined by deformed igneous megacrysts, retrograde *Pl* aggregates after *Grt*, or amphibolite-facies polycrystalline mafic clots, whereas the limbs exhibit S>L fabrics and much weaker hinge-parallel lineations (Figs. 3e, 9a; Schwerdtner and Mawer 1982, their Fig. 26.4; Davidson, Jr. 1983; Van Kranendonk 1991). Even in equigranular rocks, the dichotomy in L-S shape fabric between the hinge zone and limbs is present (e.g., Schwerdtner et al. 1981), suggesting it is an important property of the cross-folds.

In previous studies by the first author, the L>S fabrics in cross-fold hinge zones were interpreted to result from passive superposition of shape fabrics during the refolding of orogen-parallel thrust nappes by upright cross-folds (Schwerdtner et al. 1977, 1981, 2010a; Schwerdtner 1987, 1988; see also Gower 1992). However, while not questioning that cross-folding resulted in fold interference (see Fig. 5), we no longer consider the prominent L>S shape fabrics in cross-fold hinge zones were a product of this process. In light of recent kinematic modelling of transtensional folding and empirical observations of hinge-parallel L- to L>S fabrics in large cross-fold systems elsewhere (e.g., the Western Gneiss Region), it is apparent that L>S fabrics

can develop in a single act of constrictional deformation during ductile transtension. We consider this interpretation to be more compatible with our data from the ORGC in which the L- to L>S fabrics are deduced to have formed in a single act of constrictional strain and are defined by retrograde mineral assemblages indicating development after the metamorphic peak during retrogression and exhumation.

Two possible explanations for the development of shape-fabric variation around fold hinge zones have been proposed recently, one kinematic and the other dynamic. With regard to the former, Sullivan (2013; his Fig. 9) illustrated the distribution of ductile L-S fabrics in a km-scale synform that is similar to the pattern around some large cross-folds in the ORGC, with an L>S fabric 'domain' in non-layered, mechanically isotropic megacrystic orthogneiss in the hinge zone and S>L fabric 'domains' in layered, mechanically anisotropic gneisses in the limbs. Although not specifying a mechanism, Sullivan attributed the domainal L-S strain fabrics around the fold to the rheological heterogeneity of the folded rocks.

The second (dynamic) explanation follows from results of two-dimensional numerical modelling of buckle folding (e.g., Ramberg 1963a, 1963b, 1964; Dieterich 1970; Lam and Hudleston 1995; Schmalholz and Schmid 2012). For instance, in their experiments under conditions of low to moderate contrast in effective viscosity, Schmalholz and Schmid (2012) observed that progressive deformation of the multilayers commences with quasi-homogeneous pure shear thickening, but that during subsequent buckling the deformation becomes markedly heterogeneous. Thus in describing their results, they stated (p. 1819) that *'the [fold] limbs become regions of essentially passive [rigid-body] rotation and high effective viscosity, while the hinges locally exhibit high rates of [ductile] deformation and consequently low effective viscosity. ... In the final stages of [plane bulk] deformation, the limbs rotate into the field of*

[hinge-perpendicular] *extension and* [undergo] *stretching and thinning*.' In order to extend the results of their two-dimensional experiments to three-dimensional transtensional buckling, the initial quasi-homogeneous pure shear thickening would have to be transformed into a constrictional bulk flow giving rise to a non-coaxial triaxial strain. As in the two-dimensional case, however, the onset of transtensional buckling would be governed by the contrasts in effective viscosity and progressive deformation between the hinge zones and limb regions.

Of these two explanations, we consider that dynamic rotation provides a more coherent model for the development of the observed shape fabrics around fold hinge zones. Although concurring in principle with Sullivan (2013) that strain variations could be interpreted in terms of different mechanical responses of rheologically isotropic and anisotropic units in an overall homogeneous strain field, the lack of an identified mechanism to produce the strain heterogeneity leads us to prefer a three-dimensional version of the limb-rotation effect of Schmalholz and Schmid (2012).

Axial surface orientations of cross-folds

The axial surfaces of most cross-folds in the ORGC are steep to upright, with those of the km-scale structures typically being almost vertical (Fig. 5). However, the axial surfaces of some mesoscopic cross-folds, although striking approximately parallel to those of their km-scale hosts, exhibit a much wider range of dip from upright through inclined to recumbent (Fig. 3). As discussed, the strain modelling of Fossen et al. (2013) implies that such variation is to be expected in a regime of ductile oblique extension where inclined to recumbent minor folds form on the steeply-dipping limbs of pre-existing or contemporaneous structures (their Fig. 10).

Moreover, this variation in axial surface orientation within a single fold generation is not readily explained by other mechanisms.

Evaluation of criteria for transtensional folding

In the previous section we have evaluated cross-folds in the ORGC in terms of an origin by multi-order buckling, the parallelism of their hinge lines to the regional elongation direction, and the presence of L>S fabrics in their hinge zones, all features characteristic of transtensional folds and fulfilled by the examples examined. In addition, we have shown that the cross-folds display a tendency to non-cylindricity, a feature not observed in modelling experiments of 'idealised' multilayer sequences, but which we have argued should be anticipated in natural thrust-sheet stacks comprising units with different mechanical properties and pre-existing structures such as km-scale nappes and m-scale isoclinal and sheath folds. Moreover, we have shown that m-scale cross-folds exhibit a range of axial surface orientation from upright to recumbent, a diagnostic feature predicted by strain modelling of progressive transtension and not seen in folds formed in other settings. Finally, we have noted that the high-grade ductile layering deformed by the cross-folds exhibits evidence for high strain, attenuation and retrogression to amphibolite-facies assemblages, implying folding was accompanied by crustal thinning and took place after the metamorphic peak in rocks undergoing retrogression and exhumation. All these criteria are compatible with a transtensional origin for the cross-folds. We do not have a dataset with which to evaluate the rotation of their hinge lines into approximate parallelism with the regional principal strain direction (X), but we have demonstrated that the two are effectively parallel.

Simple-shear and pure-shear parts of the transtension tensor

Although the criteria discussed in the previous section are necessary to deduce a transtension origin for the cross-folds, they are not sufficient. The orientations and relative magnitudes of the simple-shear and pure-shear parts of the transtension tensor are also required. The strong signal of crustal attenuation during cross-folding supports a model of pure-shear-dominated transtension, which implies the divergence angle α was large. This first-order deduction can be refined by recalling that during progressive ductile pure-shear-dominated transtension, the hinge lines of growing transtensional folds rotate towards the divergence direction. As a result, the axial traces of the cross-folds with the most highly strained hinge zones and most prominent linear shape fabrics are predicted to be most nearly parallel to the divergence vector (Fossen et al. 2013, p. 101). Applying this rule-of-thumb to the Moon River synform, the cross-fold with the best evidence for hinge-parallel elongation, we conclude the direction of the divergence vector was close to the NW-trend of its central segment.

As described previously, the lower and upper boundaries of the crust affected by transtension are the Grenville Front Tectonic Zone and the Composite Arc Belt boundary zone respectively, and hence the simple-shear part of the transtension tensor would have been approximately parallel to these gently-dipping surfaces. Both are NE- to ENE-striking in the study area (Fig. 1). However, although containing older structures the main architecture of the Grenville Front Tectonic Zone did not develop until ~1000 Ma (Rigolet orogenic phase; i.e. post-dating cross-folding), so its orientation at the time the cross-folds developed is uncertain. The Composite Arc Belt boundary, on the other hand, is not known to carry a signal of Rigolet strain and may therefore retain its original relationship with respect to the cross-folds in its footwall. This ENE-striking boundary thus provides the best estimate for the orientation of the simple-

shear part of the transtension tensor. Resolution of a NW-trending divergence vector on its ENE-trace yields a tangential shear component with sinistral sense (Fig. 12).

The kinematic signal in tectonites along the length of the Composite Arc Belt boundary is principally indicative of NW-directed thrusting followed by SE-directed extension (Fig. 2c-d). However, the occurrence of SE-plunging mesoscopic cross-folds with hinge-parallel elongation lineations and a strong predominance of *S*-asymmetry (Hanmer and Ciesielski 1984; Hanmer 1988; Hanmer and McEachern 1992; 9:1 predominance measured by Schwerdtner et al. 2010b) supports a component of tangential, foliation-parallel sinistral shear strain bounding the zone affected by oblique extension. After discounting a possible connection with thrusting in a putative transpressional regime, Hanmer and McEachern (1992) suggested that the *S*-folds '*may represent a regionally uniform sense of [NE-directed] extensional shearing at higher levels of the now eroded Grenvillian mountain edifice*' (p. 1786) that was approximately coeval with, but subordinate to the NW-directed thrusting. On the other hand, Schwerdtner et al. (2010b) considered that the evidence favoured '*a regional component of orogen-parallel sinistral shear that probably post-dates the ductile thrusting*' (p. 773). Although unequivocal geochronological data to discriminate between these scenarios are not available, we consider the balance of evidence supports the interpretation of Schwerdtner et al. (2010b), thereby permitting linkage with the regional cross-fold system in the underlying ORGC (Fig. 12).

Aeromagnetic expression of km-scale cross-folds

Waddington (1973) noted the northwest closure of the Moon River synform was the site of a negative magnetic anomaly, and the whole ~60 km long structure is clearly imaged on more recent maps of both the total magnetic field and the vertical magnetic-gradient (Fig. 13a; Gupta

1991a, 1991b). We have used the latter map to identify two previously unrecognized km-scale cross-folds in the eastern ORGC and followed up with reconnaissance field work to determine whether they exhibit structural features compatible with a transtensional origin.

The first example is an oval NW-trending negative anomaly near Whitney (Fig. 13b), which our reconnaissance has indicated is underlain by amphibolite-facies orthogneiss with a quasi-cylindrical system of deformed leucosome veins that defines an $L>S$ fabric (Fig. 14a-d). The steeply-dipping foliation in the cross-fold limbs is deformed into recumbent buckle folds with $L>S$ mineral-shape fabrics and granitoid leucosome veins parallel to their axial surfaces (Figs. 14e-f), all features compatible with a transtensional setting. Integrating this structural and magnetic information with regional foliation trends defines a NW-trending synform of comparable scale to the Moon River and Seguin lobes in the Muskoka domain that we have named the Wallace synform after the eponymous hamlet near its centre (Fig. 12). Each of these km-scale lobes is thus characterised by: (i) NW-trending synformal foliation patterns in high-strain retrogressed rocks, (ii) upright axial surfaces and large apical angles, (iii) $L>S$ fabrics in the hinge zones, and (iv) paired negative and positive anomalies in the vertical magnetic gradient of variable intensity.

The second example is from ~20 km farther northwest in the underlying Algonquin domain where several smaller NW-trending anomalies occur (Fig. 13b), and we focus here on an area where the road through Algonquin Park (Hwy. 60) passes between two of them crossing the northeast limb of a gently SE-plunging, upright km-scale structure we term the West Smith antiform (Fig. 15). The limbs of the West Smith antiform are composed of strongly foliated ($S>>L$), variably retrogressed, veined hornblende-bearing gneisses (Figs. 16a-b), whereas most of its hinge zone consists of highly lineated ($L>S$), vein-free megacrystic orthogneiss (Figs. 16c-

e). Part of the hinge zone is occupied by a moderately stretched metagabbro body that exhibits prolate plagioclase-rich pseudomorphs after former garnet porphyroblasts with a range of X/Z ratios (Fig. 16f). As in the Moon River synform, the $L>S$ fabric defined by the stretched pseudomorphs is parallel to the cross-fold hinge line, from which we similarly infer that decompression leading to formation of the pseudomorphs and their subsequent constrictional strain took place during exhumation, retrogression and transtensional cross-folding.

The apparent lack of prior ductile deformation in the megacrystic orthogneiss with $L>S$ shape fabric in the hinge zone of the West Smith antiform is open to two interpretations: (i) the orthogneiss intruded as concordant sheets into dilatant zones that opened *during* km-scale cross-folding, perhaps resembling the synkinematic emplacement of phacoliths (Dietrich 1954; Billings 1972) or saddle reefs during upright buckling (Ramsay and Huber 1987; Vines and Law 2000); or (ii) the orthogneiss was undeformed or only weakly deformed prior to cross-folding. Given the widespread occurrence of variably deformed bodies of 1.27-1.24 Ga megacrystic orthogneiss in the ORGC (e.g., Culshaw et al. 2013), the lack of geochronological evidence for late Ottawa magmatism (e.g., Table 2 in Rivers et al. 2012), and our observations that $L>S$ fabrics in 1.26 Ga megacrystic orthogneiss are common elsewhere in the Ontario segment of the ORGC, we tentatively suggest (ii) may be more likely.

In summary, in both the regional Wallace synform in the Muskoka domain and the smaller West Smith antiform in the Algonquin domain there is a correlation among: (i) the NW-trending cross-fold pattern defined by the retrogressed post-peak foliation, (ii) the record of progressive constrictional strain in the hinge zone, and (iii) anomalous patterns in the vertical magnetic gradient. This strengthens our confidence that NW-trending, km-scale mostly negative aeromagnetic anomalies are reliable proxies for transtensional cross-folds throughout the ORGC.

Moreover, although both examples are from the hanging wall of the Allochthon Boundary, cross-folds in its footwall are equally well-imaged (e.g., Grundy Lake synform and Pakesley antiform, Fig. 7). Thus we feel justified in provisionally extrapolating our conclusions to other parts of the ORGC, including western Québec, wherever regional magnetic and geological maps show NW-trending magnetic anomalies and corresponding cross-fold patterns.

Discussion

The aim of this paper is to evaluate the case for a transtensional origin for the cross-folds in the ORGC. This is a model-driven approach, as explicitly noted by one journal reviewer, and is predicated on both the unsatisfactory and conflicting existing scenarios, and because in our estimation an oblique extensional setting is not only compatible with the structural and metamorphic data, but also sheds new light on the mechanism of exhumation of a large high-grade gneiss terrane recently interpreted as a metamorphic core complex (Rivers and Schwerdtner 2015). Thus the preceding sections constitute a series of tests of the transtension hypothesis. We contend that we have not ignored other possibilities, but rather they have been considered and found wanting, whereas a transtension origin appears to fit the data. However, given that the models on which transtension theory is based are simplifications of natural settings, it is appropriate to consider their limitations and possible refinements before coming to a conclusion.

Limitations of cartoon models of ductile transtension

Present understanding of the complex patterns of ductile transtensional strain is grounded in analogue and numerical modelling experiments. These experiments, in turn, are based on

simplified and idealised physical systems ('cartoon' models; Figs. 4, 17a) in which initially rectangular bodies of homogeneous, perfectly passive, horizontal multilayers bounded by vertically-sided rigid blocks are subjected to ductile oblique extension with a fixed horizontal divergence vector, leading to a bulk monoclinic symmetry of strain. Despite the mechanically questionable boundary conditions and assumption of homogeneous strain, it is appropriate to consider whether deformation in the central regions of such models may nonetheless approximate homogeneous strain related to the divergence vector. To investigate this further, we highlight five points in which the simplified model set-up departs significantly from the setting in the ORGC.

- In contrast to its horizontal orientation in the models, the layered ductile crust in the Muskoka and Britt domains that was affected by oblique extension dips gently to moderately southeast, leading to a triclinic symmetry of transtensional deformation and a plunging divergence vector.
- In contrast to their vertical orientation in the models, the margins of the bounding blocks, comprising the granulite-facies Algonquin domain and related subdomains and the greenschist- to amphibolite-facies Composite Arc Belt, dip gently to moderately southeast. Collectively these two points define a regime of *inclined transtension* (by analogy with terminology for transpression; e.g., Jones et al. 2004), with the Algonquin and equivalent domains forming the footwall and the Composite Arc Belt the hanging wall to the zone of transtensional strain (Fig. 17b). Moreover, at the time cross-folding was initiated, the dips of the straight gneiss foliation in the Muskoka domain and its bounding footwall would probably have been steeper than at present, since pure-shear-

dominated transtensional folding occurs in a regime of ductile flattening / vertical shortening.

- Although field evidence indicates that the footwall and hanging wall domains were significantly more competent than the Muskoka domain during oblique extension, they did not act as perfectly rigid bodies. Dynamic amphibolite-facies retrogression of footwall granulite-facies rocks in the Rosseau subdomain (part of the Algonquin domain) near its boundary with the Muskoka domain, described by Rivers and Schwerdtner (2014, 2015), implies the top of the footwall was progressively incorporated into the ductile zone affected by transtensional strain (Fig. 17c). This interpretation is also supported by map patterns indicating the presence of cross-folds in the uppermost parts of the Algonquin domain and related subdomains near their contact with the Muskoka domain (Figs. 2b, 13b). On the other hand, although cross-folds occur locally in the Composite Arc Belt boundary zone, they do not propagate into the Composite Arc Belt proper, suggesting the hanging wall more closely fulfilled the role of rigid body in the models.
- The modelling assumes that transtensional strain was homogeneous and constant volume. As noted, homogeneous strain is not only improbable on theoretical grounds (e.g., Schwerdtner 1989; Robin and Cruden 1994; Dewey et al. 1998), but given the presence within the Muskoka domain of km-scale mega-boudins with relict low-strain granulite-facies mineralogy (Fig. 12; Schwerdtner et al. 2014) and the buttressing role of the Parry Sound domain with respect to the distribution of cross-folds, it is clearly implausible in this case. Indeed, we have tentatively ascribed the distribution of cross-fold groups in Figure 2, including their sparse occurrence and different trend within the granulite-facies Parry Sound domain, to strain heterogeneity resulting from mechanical anisotropy and

local boundary conditions. Another indication of heterogeneous strain is the contrasting amplitudes and wavelengths of the regional cross-fold systems in the Muskoka and Britt domains. The constant volume assumption may also be invalid since cross-fold formation in the Muskoka domain took place under suprasolidus conditions (Rivers and Schwerdtner 2015) raising the likelihood of volume increase due to partial melting during transtensional strain. Moreover as mentioned, there is convincing evidence for dynamic increase in the thickness of the amphibolite-facies Muskoka domain at the expense of its adjacent granulite-facies footwall (Rivers and Schwerdtner 2014; Fig. 17c).

- Following from the latter point, the case can be made that the zone affected by transtension may have been wedge-shaped rather than parallel-sided, as schematically illustrated in Figure 17c.

Other refinements to the transtension model

Other possible refinements to the transtension model also involve questions about the initial assumptions. For example, is it appropriate to follow Fossen et al. (2013) in assuming the direction of the divergence-increment vector remained invariant throughout a possibly prolonged period of ductile transtension? This question can be qualitatively addressed by using the elongation lineation as a monitor of the principal extension direction (X) on a regional scale, and hence also of the approximate orientation of the divergence vector. Available data show that the elongation lineation plunges gently towards the southeast throughout much of the western ORGC (e.g., Fig. 8), suggesting that an assumption of a constant divergence vector may be reasonable. If correct, this would support the interpretation, discussed previously, that variations in lineation plunge on the 0-10m scale are a result of mechanical anisotropy or local boundary conditions in

the thrust-sheet stack (Figs. 9, 11). However, in general there is a lack of robust data to address these issues and, as noted by Dewey et al. (1998, p. 2), distinction between a regionally heterogeneous strain field and local strain heterogeneities due to mechanical anisotropies is difficult in practice.

For instance, concerning the WNW orientation of the Group C cross-fold hinge-lines in the Parry Sound domain compared to the NNW orientation of groups A and B (Fig. 2), three interpretations appear feasible: (i) Group C cross-folds formed at a different time, reflecting a temporal change in the orientation of the divergence vector; (ii) they were coeval, their different orientation being a result of stress refraction and heterogeneous strain in the mechanically competent granulite-facies mafic rocks comprising the domain; or (iii) the orientation of the cross-fold hinge lines was modified after formation. We cite the gradual change in orientation of the axial trace of the Wawashkesh-Ahmic Lakes antiform from WNW in the Parry Sound domain to NW in the Britt domain (Fig. 2), and the regionally consistent SE plunge of the elongation lineation elsewhere in much of the western ORGC, in support of our tentative preference for (ii). Thus we assume the regional orientation of the divergence vector shown in Figure 12 is also applicable to the formation of Group C cross-folds in the Parry Sound domain. The third possibility, later modification of the original orientation, may be applicable elsewhere in the ORGC, for instance to the Bracebridge antiform which was locally affected by the superimposition of late-Ottawan mega-boudins (Schwerdtner et al. 1977, 2014; Rivers and Schwerdtner 2014, 2015).

Fabric evidence supporting pure-shear-dominated transtension

The L>S fabrics in the ORGC are best defined in three lithologies, megacrystic orthogneiss, metagabbro and mafic gneiss with *Pl* pseudomorphs after *Grt*. In the first two, the

lineation potentially carries an aggregate signal of the symmetry and magnitude of the total strain experienced by the rock since its igneous crystallization, whereas in the mafic gneiss it defines the local symmetry and magnitude of the strain associated with decompression/retrogression after peak metamorphism. Inasmuch as we have deduced that transtension took place during decompression and retrogression, we are particularly interested in the latter case and the cryptic information that the *Pl* pseudomorphs after *Grt* porphyroblasts carry about the nature of associated ductile strain. However, we also point out that wherever megacrystic orthogneiss or metagabbro occur with mafic gneiss with *Pl* pseudomorphs, their strain signals are compatible, from which we deduce: (i) the pre-Grenvillian intrusions were competent bodies that were not penetratively deformed during early-Ottawan thrusting, and (ii) the post-peak SE-plunging retrograde lineation was a penetrative fabric in the ORGC.

Concerning the degree of preferred orientation of the mineral aggregates defining the elongation lineation, during progressive ductile strain in an end-member constrictional (prolate) setting the principal directions of the strain ellipsoid are pegged to material lines in a passively deforming sphere. Regardless of strain magnitude, this leads to perfectly aligned ellipsoids at any locality within the deformed zone. An essentially comparable situation occurs in large-magnitude, pure-shear-dominated ductile transtension, in which the principal diameters of neighbouring ellipsoids will accordingly be effectively parallel. Application of this principle to the *Pl* pseudomorphs that progressively replaced *Grt* porphyroblasts during decompression would result in statistical parallelism of their long axes within measurement uncertainties, despite individual pseudomorphs recording different amounts of strain.

In contrast, during simple-shear-dominated ductile transtension the progressive constrictional strain is markedly non-coaxial. Accordingly, in the case of *Pl* pseudomorphs after

Grt, this would lead to pseudomorphs formed at successive stages of deformation having nonparallel stretching directions and exhibiting a lower degree of statistical alignment. As a result, unless the finite-strain magnitude is extremely high, simple-shear-dominated transtension leads to weaker linear fabrics with a less well developed preferred orientation.

In all cases in the ORGC for which we have data, including the hinge zones of the Moon River synform and the West Smith antiform illustrated in Figs. 8 and 16 respectively, proximal *Pl* pseudomorphs exhibit very high degrees of alignment despite preserving a record of contrasting strain intensity, thereby supporting the deduction that they formed in a pure-shear-dominated transtensional regime.

Large-scale vertical thinning and compaction melt bands

The interpretation of pure-shear-dominated ductile transtension implies that the triaxially strained rocks have undergone substantial vertical thinning. For instance, Dewey et al. (1993) estimated that crust in the Western Gneiss Region had been thinned by as much as 80%, an interpretation derived from the shapes of strongly lineated feldspar augen reported by Krabbendam and Dewey (1998, p. 171). In the detachment zone of the ORGC, such a deduction is qualitatively in accord with: (i) the attenuated gneissic foliation in the Muskoka domain (Fig. 10a-b, d), (ii) retrogression and decompression of granulite- and locally eclogite-facies assemblages in the hanging wall of the Allochthon Boundary during amphibolite-facies extensional exhumation of the mid crust (Fig. 10c, e-f), and (iii) the large inter-limb angles of the major cross-folds (e.g., Fig. 5), which may be a product of partial unfolding. However, although petrological data suggest the amount of regional crustal thinning associated with transtension

must have been substantial, the lack of robust post-peak P - T - t constraints linked to strain markers precludes quantitative estimates of regional thinning associated with transtension.

Another issue relating to crustal thinning/shortening is the presence of syn-folding leucosome parallel to the axial surfaces of both upright and recumbent cross-folds (Figs. 3a, 14b, f). This arrangement poses an enigma since axial surfaces are planes of shortening strain and as such not typically sites of coeval magma accumulation. A possible explanation was advanced by Weinberg et al. (2015) on the basis of 2-D numerical strain modelling of porous media with non-linear rheology and containing an interstitial fluid (subsolidus gneiss/restite and suprasolidus leucosome respectively). Under these conditions, compaction instabilities develop parallel to planes of shortening strain such as the axial surfaces of coeval folds, inducing formation of *compaction melt bands*. In a 3-D context of strongly constrictional strain, the 2-D axial planar melt bands in their models would become L>S structures, supporting the field evidence that the axial planar and linear leucosomes formed *during* buckle folding (e.g., Fig. 14a, c-d).

Roles of the Allochthon Boundary and Grenville Front Tectonic Zone during cross-folding

Several studies have concluded that the Allochthon Boundary is a first-order structure in the Grenville Province, the site of early Ottawa (~1085-1060 Ma) thrust emplacement of granulite-facies and locally eclogite-facies rocks from the orogenic hinterland onto lower grade rocks of the contemporaneous orogenic margin (e.g., Rivers et al. 1989; Culshaw et al. 1997; Carr et al. 2000; Dickin and Guo 2001; Indares and Dunning 2004; van Gool et al. 2008; Rivers 2009). In the ORGC, the Allochthon Boundary was reworked as a normal-sense shear zone under amphibolite-facies conditions in late Ottawa time (~1020 Ma; Culshaw et al. 1997; Ketchum et al. 1998; Carr et al. 2000), and we have shown that at the time of cross-folding the

rocks in the vicinity of the Allochthon Boundary were extensively retrogressed and the boundary itself was passively folded with its adjacent footwall and hanging wall (Fig. 6a), indicating it was no longer a zone of active displacement.

In contrast, the Grenville Front Tectonic Zone is a young structure that post-dated cross-folding and developed during the Rigolet phase of the Grenvillian Orogeny at ~1000 Ma (Haggart et al. 1993; Dudás et al. 1994; Krogh 1994; Bethune 1997). As a result, Rigolet crustal shortening in the southern Grenville Front Tectonic Zone overprinted the cross-folds at their apparent northwestern limit, leading to the dextral sense of ductile deflection of their axial traces (Fig. 12a; also illustrated by the deflection of the northern lobe of the Mulock batholith).

Regional variations in late-Ottawan transtensional strain with position

The model of inclined regional ductile transtension (Fig. 17b) implies the SE-dipping surfaces of the 'rigid' bounding domains were sites of sinistral transcurrent displacement. In order to investigate this deformation further we focus on the signal at its upper contact with the Composite Arc Belt boundary zone, where mesoscopic buckle folds with *S*-asymmetry occur as isolated pairs rather than long fold trains, suggesting they formed at random sites of strain perturbation during regional simple shear. Irrespective of buckling kinematics, this attests to wall-rock strain in which the tangential sinistral shear was *not* accompanied by significant finite shortening parallel to the Composite Arc Belt boundary (Schwerdtner et al. 2010b, p. 782). The long limbs of the *S*-fold pairs are concordant to the foliation implying the *S*-asymmetry pertains to the regional sense of foliation-parallel shear. This suggests that deformation in the Composite Arc Belt boundary was close to tangential simple shear, implying the same shear-sense across all planes co-zonal with the simple-shear direction (see Malavieille 1987, his Fig. 15; Klemens

1996). However, in contrast to the ORGC the transtensional deformation was distributed into orthogonal extension and shear components, giving rise to the down-dip elongation lineation and reclined *S*-shaped folds with collinear axes respectively.

An important implication is that during late-Ottawan oblique extension, deformation varied systematically with vertical position in the crust: from pure-shear-dominated ductile transtension in the detachment zone at the top of the ORGC, to distributed down-dip extension and simple-shear in the overlying Composite Arc Belt boundary (Fig. 18a-b). This implies that the pure-shear part of the transtension tensor (i.e., ductile vertical shortening) decreased across this boundary. Considering also the Bancroft shear zone, a ductile extensional structure with down-dip displacement in the northern Composite Arc Belt that formed during or following cross-folding (Fig. 18a; van der Pluijm and Carlson 1989), it is apparent that ductile transtension was restricted to the detachment zone at the top of the ORGC and that deformation in the overlying more competent rocks of the Composite Arc Belt was limited to narrower shear zones with distributed strike-slip and extensional displacement.

The role of the Big Bend

Since the Big Bend was previously implicated as a possible factor in the origin of cross-folds (Schwerdtner 1987), it is revisited briefly here. The Big Bend is defined by the curved traces of several prominent structural boundaries in the western Grenville Province on regional geological maps, including from northwest to southeast, the Grenville Front, the Grenville Front Tectonic Zone, the Composite Arc Belt boundary, the Allochthon Boundary and the basal thrust of the Parry Sound domain (Fig. 1; Schwerdtner et al. 2010b). It is thus a feature of orogenic scale. In three-dimensions these curved boundaries correspond to SE-plunging open folds with

large inter-limb angles; for instance, the Composite Arc Belt boundary near Minden defines a gently SE-plunging flexure with an inter-limb angle of $>130^\circ$ (Fig. 1; Schwerdtner et al. 2010b, p. 874), as illustrated schematically in Figure 18c. Two origins for the Big Bend appear possible: (i) it originated as a pre-Grenvillian re-entrant in the margin of SE Laurentia that governed subsequent ductile deformation of the orogenic crust (see Marshak 2004); or (ii) the early-Ottawan thrust-sheet stack was flexed at this location in the late-Grenvillian. In either case, the *S*-asymmetry of the mesoscopic buckle folds and the bend of the Composite Arc Belt boundary near Minden were not created by large-scale flexing of the wall rocks (Schwerdtner et al. 2010b), implying the Big Bend did not play a significant role in the transtensional deformation discussed in this paper.

Tectonic significance of transtension

The interpretation of a transtensional origin for Grenvillian cross-folding expands the range of tectonic settings for which extension is held to be the driver for exhumation of high-grade metamorphic rocks. Malavieille (1987, 1993) was the first to propose that extensional simple shear played an important role in the exhumation of upper- to mid-crustal Cordilleran core complexes in the Mesozoic. Subsequently, Kabbendam and Dewey (1998) showed that ductile transtension led to the exhumation of the HP and UHP Western Gneiss Region of the Scandinavian Caledonides during the Paleozoic. Our study appears to be the first to call upon transtension as the driving mechanism for exhumation of hot mid-crust in a Precambrian large hot collisional orogen. A common feature of all three settings is that the exhumed high-grade rocks occur in metamorphic core complexes situated beneath, and separated from, adjacent lower-grade rocks by detachment zones in which deformation is heterogeneously distributed and

ranges from extensional, through transtensional, to strike-slip. From the variety of tectonic settings and ages embraced by the three studies (Mesozoic continental-margin arc, Paleozoic (U)HP units in a collisional orogen, and latest Mesoproterozoic high-grade core of a collisional large hot orogen), it can be concluded that extension and transtension have been effective mechanisms for the exhumation of overthickened ductile crust since the onset of plate tectonics.

Conclusions

As predicted by kinematic modelling and confirmed by field studies, upright to recumbent, horizontal to moderately plunging buckle cross-folds of the attenuated layering, with hinge-parallel elongation lineations and trends ranging from oblique to perpendicular to the orogenic front, are distinctive signals of ductile pure-shear-dominated transtensional regimes. Together with progressive constrictional deformation in the fold-hinge zones, these features constitute the hallmarks of ductile oblique extension. We have described an inclined system of gently-plunging multi-order transtensional cross-folds ranging from metre to kilometre scale in the western Ottawa River Gneiss Complex that exhibits all these features. The folds are post-peak-metamorphic structures that occur in ductile, melt-weakened amphibolite-facies rocks situated in the detachment zone of a giant metamorphic core complex. The orientations of most m-scale cross-folds are similar to the km-scale structures on which they form, but some have much steeper plunges and more varied axial surface orientations, interpreted to be a result of superposition on the dipping surfaces of the early-Ottawan thrust-sheet stack. Irrespective of orientation, however, they exhibit constrictional $L > S$ fabrics that are parallel to those of their host km-scale structures.

The regional tectonic setting of the cross-folds in the ductile detachment zone separating the core and carapace of the metamorphic core complex is compatible with formation during post-peak gravitational collapse of a hot crustal-scale thrust stack. Evidence that cross-fold formation took place in a single, possibly protracted, phase of deformation after the metamorphic peak during retrogression and decompression is documented in retrogressed mafic granulites that occur in the hinge zones of some cross-folds. In these rocks, *An*-rich plagioclase aggregates partially to completely pseudomorph *Grs*-rich garnet porphyroblasts, the variably stretched prolate pseudomorphs defining a record of progressive constrictional strain during exhumation.

Absolute timing of the pure-shear-dominated ductile transtension in the Ottawa River Gneiss Complex remains poorly constrained. It followed some time after the peak-Ottawan metamorphism at ~1085 Ma, was initiated under suprasolidus conditions when some leucosome still remained, and continued at least until the late Ottawan, ~1020 Ma, under sub-solidus conditions.

Acknowledgements

This paper is a contribution to our ongoing study of the post-peak-metamorphic evolution of the Ottawa River Gneiss Complex. We are pleased to acknowledge that the interpretation of a transtensional setting for the cross-folds was strongly influenced by the seminal paper on the kinematic modelling of transtensional strain by Håkon Fossen, Christian Teyssier and Donna Whitney published recently in *The Journal of Structural Geology*. For WMS, this paper provided the stimulus to return to the cross-fold enigma, and for DHW the opportunity to re-interpret data from his MSc thesis in a new context. We thank Bob Hatcher and Bill Thomas for discussions

concerning proper use of the term gneiss complex, and are grateful to journal reviewers Peter Hudleston, Maarten Krabbendam, and two who chose to remain anonymous for their perceptive insights on transtension and comments that forced us to rigorously justify our conclusions. Careful editorial work by Associate Editor Kathryn Bethune is appreciated. We thank our field assistants, especially Daniel Landa, Sheng J. Lu and Charles Wang for their help and enthusiasm. Early drafts of some figures were drawn by Karin Gorra (University of Toronto). TR acknowledges support from the NSERC Discovery Grant program.

References

- Aller, J., and Gallastegui, J. 1995. Analysis of kilometric-scale superposed folding in the Central Coal Basin (Cantabrian Zone, NW Spain). *Journal of Structural Geology*, **17**: 961-969.
- Alsop, G.I., Holdsworth, R.E., and Strachan, R.A. 1996. Transport-parallel cross-folds within a mid-crustal Caledonian thrust stack, northern Scotland. *Journal of Structural Geology*, **18**: 783-790.
- Beaumont, C., Nguyen, M.H., Jamieson, R.A., and Ellis, S. 2006. Crustal flow modes in large hot orogens. *In* Channel Flow, Ductile Extrusion and Exhumation in Continental Collision Zones. *Edited by* R.D. Law, M.P. Searle, and L. Godin. Geological Society of London, Special Publication **268**: 91-145.
- Bethune, K.M. 1997. The Sudbury dyke swarm and its bearing on the tectonic development of the Grenville Front, Ontario, Canada. *Precambrian Research*, **85**: 117-146.
- Billings, M.P. 1972. *Structural Geology*. 3rd ed. Prentice-Hall Inc., Englewood Cliffs, N.J.
- Bussy, F., Krogh, T.E., Klemens, W.P., and Schwerdtner, W.M. 1995. Tectonic and metamorphic events in the westernmost Grenville Province, central Ontario: new results

- from high-precision U-Pb zircon geochronology. *Canadian Journal of Earth Sciences*, **32**: 660-671.
- Carr, S.D., Easton, R.M., Jamieson, R.A., and Culshaw, N.G. 2000. Geologic transect across the Grenville orogen of Ontario and New York. *Canadian Journal of Earth Sciences*, **37**: 193-216.
- Chen, W. 2014 The development of L-tectonites in high-strain zone settings: A multiscale modeling investigation. Ph.D. thesis, University of Western Ontario, London, ON. 115 p.
- Corrigan, D., Culshaw, N.G., and Mortensen, J.K. 1994. Pre-Grenvillian evolution and Grenvillian overprinting of the Parautochthonous Belt in the Key Harbour area, Ontario: U-Pb constraints. *Canadian Journal of Earth Sciences*, **31**: 583-596.
- Cottrell, A.H. 1953. *Dislocations and Plastic Flow in Crystals*. Oxford University Press, London, England.
- Culshaw, N.G. 2005. Buckle folding and deep-crustal shearing of high-grade gneisses at the junction of two major high-strain zones, Central Gneiss Belt, Grenville Province, Ontario. *Canadian Journal of Earth Sciences*, **42**: 1907-1925.
- Culshaw, N.G., Davidson, A., and Nadeau, L. 1983. Structural subdivisions in the Parry Sound – Algonquin region, Ontario. *In* Current Research, Part B. Geological Survey of Canada, Paper 83 – 1B, 243-252.
- Culshaw, N.G., Reynolds, P.H., and Check, G. 1991. A $^{40}\text{Ar}/^{39}\text{Ar}$ study of post-tectonic cooling in the Britt domain of the Grenville Province, Ontario. *Earth and Planetary Science Letters*, **105**: 405-415.

- Culshaw, N.G., Ketchum, J.W.F., Wodicka, N., and Wallace, P. 1994. Deep-crustal ductile extension following thrusting in the southwestern Grenville Province, Ontario. *Canadian Journal of Earth Sciences*, **31**: 160-175.
- Culshaw, N.G., Jamieson, R.A., Ketchum, J.W.F., Wodicka, N., Corrigan, D., and Reynolds, P.H. 1997. Transect across the southwestern Grenville orogen, Georgian Bay, Ontario: polystage convergence and extension in the lower orogenic crust. *Tectonics*, **16**: 966-982.
- Culshaw, N.G., Corrigan, D., Ketchum, J.W.F., Wallace, P., Wodicka, N., and Easton, R.M. 2004. Georgian Bay Geological Synthesis, Grenville Province: Explanatory Notes for Preliminary Maps P.3548 to P.3552. Ontario Geological Survey, Open File Report 6143.
- Culshaw, N.G., Beaumont, C., and Jamieson, R.A. 2006. The orogenic superstructure-infrastructure concept: Revisited, quantified, and revived. *Geology*, **34**: 733-736.
- Culshaw, N.G., Gerbi, C., and Marsh, J.H. 2010. Softening of the lower crust: Modes of syn-transport transposition around and adjacent to a deep-crustal granulite nappe, Parry Sound domain, Grenville Province, Ontario, Canada. *Tectonics*, **29**: TC5013.
- Culshaw, N.G., Slagstad, T., Raistrick, M., and Dostal, J. 2013. Geochemical, geochronological and isotopic constraints on the origin of the allochthonous Shawanaga and basal Parry Sound domains, Central Gneiss Belt, Ontario. *Precambrian Research*, **228**: 131-150.
- Culshaw, N., Gerbi, C., and Ratcliffe, L. 2015. Macro- and microstructural analysis of the North Tea Lake Mylonite Zone: an extensional shear zone in the Central Gneiss Belt, Grenville Province, Ontario. *Canadian Journal of Earth Sciences*, **52**: 1027-1044.
- Davidson, A., 1984a. Identification of ductile shear zones in the southwestern Grenville Province of the Canadian Shield. *In* *Precambrian Tectonics Illustrated*. Edited by A. Kröner and R. Greiling. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, Germany, 263-279.

- Davidson, A., 1984b. Tectonic boundaries in the Grenville Province of the Canadian Shield. *Journal of Geodynamics*, **1**: 433-444.
- Davidson, A. 1996. Geology of the Grenville Province, 1: 2,000,000 scale. Geological Survey of Canada, Open File 3346.
- Davidson, A., and Morgan, W.C. 1981. Preliminary notes on the geology east of Georgian Bay, Grenville Structural Province, Ontario. *In* Current Research, Part A. Geological Survey of Canada, Paper 81-1A, 291-298.
- Davidson, A., Culshaw, N.G., Nadeau, L. 1982. A tectono-metamorphic framework for part of the Grenville Province, Ontario. *In* Current Research, Part A. Geological Survey of Canada, Paper 82-1A, 275-290.
- Davidson, D. M., Jr. 1983. Strain analysis of deformed granitoid rocks (Helikian), Muskoka district, Ontario. *Journal of Structural Geology*, **5**: 181-195.
- Dewey, J.F., Ryan, P.D., and Andersen, T.B. 1993. Orogenic uplift and collapse, crustal thickness, fabrics and metamorphic phase changes; the role of eclogites. *In* Magmatic Processes and Plate Tectonics, *edited by* H.M. Alabaster, N.B.W. Harris, and C.R. Neary, Geological Society of London, Special Publication **76**: 325-343.
- Dewey, J.F., Holdsworth, R.E., and Strachan, R.A. 1998. Transpression and transtension zones. *In* Continental transpressional and transtensional tectonics, *edited by* R.E. Holdsworth, R.A. Strachan, and J.F. Dewey, Geological Society of London, Special Publication **135**: 1-14. doi:10.1144/GSL.SP.1998.135.01.01.
- Dickin, A.P., and Guo, A. 2001. The location of the Allochthon Boundary Thrust and the Archean-Proterozoic suture in the Mattawa area of the Grenville Province: Nd isotopic evidence. *Precambrian Research*, **107**: 31-43.

- Dickin A. P., Moretta, K. and North, R. 2008. Isotope mapping of the Allochthon Boundary Thrust in the Grenville Province of Ontario, Canada. *Precambrian Research*, **167**: 260-266.
- Dieterich, J.H. 1970. Computer experiments on mechanics of finite-amplitude folds. *Canadian Journal of Earth Sciences*, **7**: 467-476.
- Dietrich, R.V. 1954. Fish Creek phacolith, northwestern New York. *American Journal of Science* **252**: 513-531.
- Dudás, F.Ö., Davidson, A., Bethune, K.M. 1994. Age of the Sudbury dykes and their metamorphism in the Grenville Province. *In Radiometric Age and Isotopic Studies, Report 8, Geological Survey of Canada, Paper 1994-F: 97-106*
- Dutton, B.J. 1997. Finite strains in transpression zones with no boundary slip. *Journal of Structural Geology*, **19**: 1189-1200.
- Easton, R.M. 1992. The Grenville Province and the Proterozoic history of central and southern Ontario, *in Geology of Ontario, edited by P.C. Thurston, H.R. Williams, R.H. Sutcliffe, and G.M. Stott: Sudbury, Canada, Ontario Geological Survey. Special Volume 4, 715-790.*
- Easton, R.M., 2001. Precambrian geology, Barry's Bay area. Ontario Geological Survey, Preliminary map P.3444, scale 1:50,000.
- Fisher, G.M. and Olsen, S.N. 2004. The Baltimore gneiss domes of the Maryland Piedmont. *In Gneiss Domes in Orogeny, edited by D.L. Whitney, C. Teyssier, and C.S. Siddoway. Geological Society of America, Special Paper 380, 307-320.*
- Flinn, D. 1965. On folding during three-dimensional progressive deformation. *Geological Society of London Quarterly Journal*, **118**: 385-433.
- Fossen, H. 1992. The role of extensional tectonics in the Caledonides of south Norway. *Journal of Structural Geology*, **14**: 1033-1046.

- Fossen, H., Tikoff, B., and Teyssier, C. 1994. Strain modeling of transpressional and transtensional deformations. *Norsk Geologisk Tidsskrift*, **74**: 134-145.
- Fossen, H. and Tikoff, B. 1998. Extended models of transpression and transtension, and application to tectonic settings. *In Continental Transpressional and Transtensional Tectonics, edited by R.E. Holdsworth, R.A. Strachan, and D.F. Dewey*, Geological Society, London, Special Publications, **135**: 15-33.
- Fossen, H., Teyssier, C., and Whitney, D.L. 2013. Transtensional folding. *Journal of Structural Geology*, **56**: 89-102.
- Friedel, J. 1957. Dislocations and Mechanical Properties of Crystals. Wiley & Sons, New York.
- Fyson, W.K., 1964. Anatomy and evolution of a pull-apart basin, Stellarton, Nova Scotia. *Geological Society of America Bulletin*, **116**: 119-127.
- Ghosh, S.K. and Ramberg, H. 1968. Buckling experiments on intersecting fold patterns. *Tectonophysics*, **5**: 89-105.
- Ghosh, S.K., Mandal, N., Sengupta, S., Deb, S.K., Khan, D. 1993. Superimposed buckling in multilayers. *Journal of Structural Geology*, **15**: 95-111.
- Ghosh, S.K., Khan, D., and Sengupta, S. 1995. Interfering folds in constrictional deformation. *Journal of Structural Geology*, **17**: 1361-1373.
- Gower, R.J.W. 1992. Nappe emplacement direction in the Central Gneiss Belt, Grenville Province, Ontario, Canada; evidence for oblique collision. *Precambrian Research*, **59**: 73-94.
- Gupta, V.K. 1991a. Shaded image of total magnetic field of Ontario, southern sheet. Ontario Geological Survey, Map 2587, scale 1:1,000,000.

- Gupta, V.K. 1991b. Vertical magnetic gradient of Ontario, southern sheet. Ontario Geological Survey, Map 2591, Scale 1:1,000,000.
- Haggart, M.J., Jamieson, R.A., Reynolds, P.H., Krogh, T.E., Beaumont, C., and Culshaw, N.G. 1993. Last gasp of the Grenville orogeny: thermochronology of the Grenville Front Tectonic Zone near Killarney, Ontario. *Journal of Geology*, **101**: 575-589.
- Harland, B.W. 1971. Tectonic transpression in Caledonian Spitzbergen. *Geological Magazine* **146**: 419-421.
- Hanmer, S. 1988. Ductile thrusting at mid-crustal level. *Canadian Journal of Earth Sciences*, **25**: 1049-1059.
- Hanmer, S. and Ciesielski, A. 1984. A structural reconnaissance of the northwest boundary of the Central Metasedimentary Belt, Grenville Province, Ontario and Québec. *In* Current Research, Part B, Geological Survey of Canada, Paper 84-1B, 121-131.
- Hanmer, S. and McEachern, S. 1992. Kinematical and rheological evolution of a crustal-scale ductile thrust zone, Central Metasedimentary Belt, Grenville orogen, Ontario. *Canadian Journal of Earth Sciences*, **29**: 1779-1790.
- Harris, A.L., and Johnson, M.R.W. 1991. Moine. *In* *Geology of Scotland*, edited by C. Y. Craig. Geological Society, London, 87-118.
- Hatcher, R.D., Jr. 1995. *Structural Geology*. 2nd ed. Prentice-Hall Inc., Englewood-Cliffs, N.J.
- Hatcher, R.D., Jr. 2010. The Appalachian orogen: A brief summary. *In* *From Rodinia to Pangea: The Lithologic Record of the Appalachian Region*, edited by R.P. Tollo, M.J. Bartholomew, J.P. Hibbard, and P.M. Karabinos. Geological Society of America, Memoir 206, 1-20.

- Hatcher, R.D., Jr., and Hooper, R.J. 1992. Evolution of crystalline thrust sheets in the internal parts of mountain chains. *In* Thrust Tectonics, *edited by* K.R. McClay. Chapman & Hall, London, England, 217-233.
- Heim, A. 1921. Geologie der Schweiz. Tauschnitz Verlag, Leipzig, Germany, 259-476.
- Hewitt, D.F. 1967. Geology and mineral deposits of the Parry Sound-Huntsville area. Ontario Department of Mines, Report 52.
- Hynes, A., and Rivers, T. 2010. Protracted continental collision—evidence from the Grenville Orogen. *Canadian Journal of Earth Sciences*, **47**: 592-620.
- Indares, A., and Dunning, G. 2004. Crustal architecture above the high-pressure belt of the Grenville Province in the Manicouagan area: new structural, petrologic and U-Pb age constraints. *Precambrian Research*, **130**: 199-228.
- Jamieson, R.A., and Beaumont, C. 2011. Coeval thrusting and extension during lower-crustal ductile flow – implications for exhumation of high-grade metamorphic rocks. *Journal of Metamorphic Geology*, **29**: 33-51.
- Jamieson, R.A., Beaumont, C., Nguyen, M.H., and Culshaw, N.G. 2007. Synconvergent ductile flow in variable-strength ductile crust: Numerical models with application to the western Grenville orogen. *Tectonics* 26, TC5005, 1-23.
- Johnson, E.L., Goergen, E.T., and Fruchey, B.L. 2004. Right lateral oblique slip movements followed by post-Ottawan (1050–1020 Ma) orogenic collapse along the Carthage Colton shear zone: Data from the Dana Hill metagabbro body, Adirondack Mountains, New York. *In* Proterozoic Tectonic Evolution of the Grenville Orogen in North America, *edited by* R.P. Tollo, L. Corriveau, J. McLelland, and M.J. Bartholomew. Geological Society of America, Memoir **197**, 357-268.

- Jones, R.R., Holdsworth, R.E., and Bailey, W. 1997. Lateral extrusion in transpression zones: the importance of boundary conditions. *Journal of Structural Geology*, **19**: 1201-1218.
- Jones, R.R., and Holdsworth, R.E. 1998. Oblique simple shear in transpression zones. *Geological Society of London, Special Publications*, **135**: 35-40. doi:10.1144/GSL.SP.1998.135.01.01.
- Jones, R.R., Holdsworth, R.E., Clegg, P., McCaffrey, K., and Tavarnerelli, E. 2004. Inclined transpression. *Journal of Structural Geology*, **26**: 1531-1548.
- Ketchum, J.W.F., and Davidson, A. 2000. Crustal architecture and tectonic assembly of the Central Gneiss Belt, southwestern Grenville Province, Canada: a new interpretation. *Canadian Journal of Earth Sciences*, **37**: 217-234.
- Ketchum, J.W.F., Heaman, L.M., Krogh, T.E., Culshaw, N.G., and Jamieson, R.A. 1998. Timing and thermal influence of late orogenic extension in the lower crust: a U-Pb geochronological study from the southwest Grenville orogen, Canada. *Precambrian Research*, **89**: 25-45.
- King, R.F., and Ries, A.I. 1962. The measurement of the anisotropy of magnetic susceptibility of rocks by the torque method. *Journal of Geophysical Research*, **67**: 1565-1572.
- Klemens, W.P. 1996. Structural analysis of the Ahmic domain and its border zones, southwestern Grenville Orogen, Ontario, Canada. Ph.D. thesis, University of Toronto, Toronto, Ontario.
- Klemens, W.P. and Schwerdtner, W.M. 1997. Emplacement and deformation of granite-pegmatite dikes in a mid-crustal regime of late-orogenic extension, southwest Grenville Province, Ontario, Canada. *Geological Magazine*, **134** (3): 287-295. doi: 10.1017/S0016756897007048.
- Kobberger, G., and Zulauf, G. 1995. Experimental folding and boudinage under pure constrictional conditions. *Journal of Structural Geology*, **17**: 1055-1063.

- Krabbendam, M., and Dewey, J.F. 1998. Exhumation of UHP rocks by transtension in the Scandinavian Caledonides. *In* Continental transpressional and transtensional tectonics, *edited by* R.E. Holdsworth, R.A. Strachan, and J.F. Dewey, Geological Society of London, Special Publication **135**: 1–14. doi:10.1144/GSL.SP.1998.135.01.01.
- Krogh, T.E. 1994. Precise U-Pb ages for Grenvillian and pre-Grenvillian thrusting of Proterozoic and Archean metamorphic assemblages in the Grenville Front tectonic zone, Canada. *Tectonics*, **13**: 963-982.
- Kvale, A. 1953. Linear structures and their relation to movement in the Caledonides in Scandinavia and Scotland. *Geological Society of London Quarterly Journal*, **109**: 51-73.
- Lan, L. and Hudleston, P.J. 1995. The effects of rheology on single-layer buckle folds. *Journal of Structural Geology*, **17** (5): 727-738. doi: 10.1016/0191-8141(94)00095H.
- Leslie, A.G., Krabbendam, M., Kimbell, G.S., and Strachan, R.A. 2010. Regional-scale lateral variation and linkage in ductile thrust architecture: the Oykel Transverse Zone, and mullions, in the Moine nappe, NW Scotland. *Geological Society, London, Special Publications*, **335**: 359-381.
- Lumbers, S.B. 1971. Geology of the North Bay area, districts of Nipissing and Parry Sound. Ontario Department of Mines, Geological Report 94.
- Lumbers, S.B. 1975. Geology of the Burwash area, Districts of Nipissing and Parry Sound, Ontario Division of Mines, Report 116.
- Lumbers, S.B. 1976a. Mattawa-Deep River area (western half), District of Nipissing. Ontario Division of Mines, Preliminary Map P.1196, Geological Series.
- Lumbers, S.B. 1976b. Mattawa-Deep River area (eastern half), District of Nipissing, Ontario Division of Mines, Preliminary Map P.1197, Geological Series.

- Lumbers, S.B. 1996. Mapping Muskoka's rocks. *Rotunda Magazine*, Royal Ontario Museum, Toronto, **29**: 27-35.
- Lumbers, S.B., and Vertolli, V.M. 2000a. Precambrian geology, Lake Joseph area. Ontario Geological Survey Map P.3409, scale 1:50,000.
- Lumbers, S.B., and Vertolli, V.M. 2000b. Precambrian geology, Bracebridge area. Ontario Geological Survey Map P. 3411, scale 1:50,000.
- Lumbers, S.B., Vertolli, V.M., and Schwerdtner, W.M. 2000a. Precambrian Geology, Gravenhurst area. Ontario Geological Survey Map P.3409, scale 1:50,000.
- Lumbers, S. B., Vertolli, V. M, and Schwerdtner, W. M. 2000b. Precambrian Geology, Penetanguishene area. Ontario Geological Survey Map. P.3410, scale 1:50,000.
- Malavieille, J. 1987. Extensional shearing deformation and kilometre-scale 'a'-type folds in a Cordilleran metamorphic core complex (Raft River Mountains, NW Utah). *Tectonics*, **6**: 423-448.
- Malavieille, J. 1993. Late orogenic extension in mountain belts: insights from the Basin and Range and the late Paleozoic Variscan belts. *Tectonics*, **12** (5): 1115-1130. doi: 10.1029.93TC01129.
- Marsh, J.H., Culshaw, N.G., and Gerbi, C.C. 2013. Timing and conditions of polyphase metamorphism within the Twelve Mile Bay shear zone: implications for the evolution of mid-crustal décollement zones and western Grenville tectonics. *International Geology Review*, DOI:10.1080/00206814.2013.773768
- Marshak, S. 2004. Salients, recesses, arcs, oroclines, syntaxes - a review of ideas concerning the formation of map-view curves in fold-thrust belts. *In* *Thrust Tectonics and Hydrocarbon*

- Systems, *edited by* K.R. McClay. American Association of Petroleum Geologists Memoir, **82**: 131-156.
- Martignole, J., and Calvert, A.J. 1996. Crustal-scale shortening and extension across the Grenville Province of western Quebec. *Tectonics*, **15**: 376-386.
- Martignole, J., Calvert, A.J., Friedman, R., and Reynolds, P. 2000. Crustal evolution along a seismic section across the Grenville Province (western Quebec). *Canadian Journal of Earth Sciences*, **37** (2/3): 291-306.
- McLelland, J.M., and Selleck, B.W. 2011. Megacrystic Gore Mountain-type garnets in the Adirondack Highlands: Age, origin and tectonic implications. *Geosphere*, **7**: 1194-1208.
- McLelland, J.M., Selleck, B.W., and Bickford, M.E. 2010. Review of the Proterozoic evolution of the Grenville Province, its Adirondack outlier and the Mesoproterozoic inliers of the Appalachians. *In: From Rodinia to Pangea: The Litho-tectonic Record of the Appalachian Region*, *edited by* Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P.M. Geological Society of America, Boulder, Colorado. Memoir 206, 777-794. doi:10.1130/2010.1206 (30).
- Ontario Geological Survey 1991. Bedrock Geology of Ontario, southern sheet. Map 2544, scale 1: 1,000,000, Ontario Geological Survey, Sudbury, Ontario.
- Ontario Geological Survey 1992. Tectonic Assemblages of Ontario, southern sheet. Map 2578, scale 1: 1,000,000, Ontario Geological Survey, Sudbury, Ontario.
- Osmundsen, R.T. and Andersen, T.B. 2001. The middle Devonian basins of western Norway: sedimentary response to large-scale transtensional tectonics? *Tectonophysics*, **332**: 51-68.

- Powell, D. and Phillips, W.E.A. 1985. Time of deformation in the Caledonian orogen of Britain and Ireland. *In* The nature and Timing of Orogenic Activity in the Caledonian Rocks of the British Isles, *edited by* A.L. Harris. Geological Society of London, Memoir 9, pp. 17-40.
- Ramberg, H. 1963a. Evolution of drag folds. *Geological Magazine*, **100** (2): 97-106. doi: 10.1017/S0016756800055321.
- Ramberg, H. 1963b. Fluid dynamics of viscous buckling applicable to folding of layered rocks. *Bulletin of the American Association of Petroleum Geologists*, **47**: 484-505.
- Ramberg, H. 1964. Selective buckling of composite layers with contrasted rheological properties: a theory of simultaneous formation of several orders of folds. *Tectonophysics*, **1**: 307-341. doi: 10.1016/0040-1951(64)90020-4.
- Ramsay, J.G. 1967. Folding and fracturing of rocks. McGraw-Hill Book Company, New York, N.Y., USA.
- Ramsay, J.G. and Huber, M.I. 1987. The techniques of modern structural geology, Volume II, folds and fractures. Academic Press, London, England.
- Rive, M. 1973a. Géologie de la région des lacs Beauchene et Bleu. Ministère des Richesses Naturelles, Québec. Rapport no. 580.
- Rive, M. 1973b. Géologie de la région des lacs Ogascanane et Sairs. Ministère des Richesses Naturelles, Québec. Rapport no. 606.
- Rivers, T. 2008. Assembly and preservation of lower, mid and upper orogenic crust in the Grenville Province—implications for the evolution of large hot long-duration orogens. *Precambrian Research*, **167** (3-4): 237-259. doi:10.1016/j.precambres.2008.08.005
- Rivers, T. 2009. The Grenville Province as a large hot long-duration collisional orogen – insights from the spatial and thermal evolution of its orogenic fronts. *In* Ancient Orogens and

- Modern Analogues, *edited by* J.B. Murphy, J.D. Keppie, and A. Hynes, Geological Society of London, Special Publication **327**, 405-444. doi:10.1144/SP327.17
- Rivers, T. 2012. Upper-crustal orogenic lid and mid-crustal core complexes: signature of a collapsed orogenic plateau in the hinterland of the Grenville Province. *Canadian Journal of Earth Sciences* **49**: 1-42. doi: 10.1139/E11-114
- Rivers, T., and Schwerdtner, W.S. 2014. New ideas on the post-peak development of the Central Gneiss Belt in the Muskoka Region. Friends of the Grenville field trip guide, 45p. Accessible at <friendsofthegrenville.org>
- Rivers, T., and Schwerdtner, W.S. 2015. Post-peak evolution of the Muskoka domain, western Grenville Province: ductile detachment in a crustal-scale metamorphic core complex. *Geoscience Canada*, 42: 403-436. doi.org/10.12789/geocanj.2015.42.080
- Rivers, T., Martignole, J., Gower, C.F., Davidson, A. 1989. New tectonic divisions of the Grenville Province, southeast Canadian Shield. *Tectonics*, **8**: 63-84. doi: 10.1029/TC008i001p00063.
- Rivers, T., Ketchum, J., Indares, A., and Hynes, A. 2002. The High Pressure belt in the Grenville Province: architecture, timing, and exhumation. *Canadian Journal of Earth Sciences*, **39**: 867-893.
- Rivers, T., Culshaw, N.G., Hynes, A., Indares, A., Jamieson, R., and Martignole, J. 2012. The Grenville Orogen – a post-Lithoprobe perspective. *In* *Tectonic Styles in Canada: the Lithoprobe perspective*, *edited by* J.A. Percival, F.A. Cook, and R.M. Cloves. Geological Association of Canada, Special Paper 49, 97-238.

- Robin, P.-Y.F. and Cruden, A.R. 1994. Strain and vorticity patterns in ideally ductile transpression zones. *Journal of Structural Geology*, **16**: 447-466.
- Sanderson, D.J., and Marchini, W.R.D. 1984. Transpression. *Journal of Structural Geology*, **6**: 449-458.
- Sati D.C. and Rautela, P., 1993. The significance of cross folding in Himalayan tectonic history. *Episodes*, **16**, 389-394.
- Schmalholz, F.M., and Schmid, D.W. 2012. Folding in power-law viscous materials. *Philosophical Transactions of the Royal Society, series A*, **370**: 1798-1826. doi: 10.1098/rsta.2011.0421.
- Schwerdtner, W.M. 1987. Interplay of folding and ductile shearing in the Proterozoic crust of the Muskoka–Parry Sound region of central Ontario. *Canadian Journal of Earth Sciences*, **24**: 1507-1525.
- Schwerdtner, W.M. 1988. Recognition of pervasive prestrain in the total-strain pattern of large folds: *Journal of Structural Geology*, **10**: 33-40.
- Schwerdtner, W.M. 1989. The solid-body tilt of deformed paleohorizontal planes: application to an Archean transpression zone, southern Canadian Shield. *Canadian Journal of Earth Sciences*, **11**: 1012-1027.
- Schwerdtner, W.M., and Waddington, D.H. 1978. Structure and lithology of the Muskoka-southern Georgian Bay region, central Ontario. *In* Toronto '78, Field Trips Guidebook. *Edited by* A.L. Currie and W.O. Mackasey. Joint Annual Meeting of the Geological Society of America, the Geological Association of Canada and the Mineralogical Association of Canada, Toronto, Ontario, pp. 204-212.

- Schwerdtner, W.M., and Mawer, C. K. 1982. Geology of the Gravenhurst region, Grenville Structural Province, Ontario. *In* Current Research, part B. Geological Survey of Canada, Paper 82-1B, 197-207.
- Schwerdtner, W.M. and Gapais, D. 1983. Calculation of finite incremental deformations in ductile geological materials and structural models. *Tectonophysics*, **93**: T1-T7.
- Schwerdtner, W.M., and van Berkel, J.T. 1991. The origin of fold abutments in the map pattern of the westernmost Grenville Province, central Ontario: *Precambrian Research*, **49**: 39-59.
- Schwerdtner, W.M., and Cote, M.L. 2001 Patterns of pervasive shear strain near the boundaries of the La Ronge domain, inner Trans-Hudson Orogen, western Canadian Shield. *Precambrian Research*, **107**: 93–116.
- Schwerdtner, W.M., and Klemens, W.P. 2008. Structure of Ahmic domain and its vicinity, southwestern Central Gneiss Belt, Grenville Province of Ontario (Canada). *Precambrian Research*, **167**: 16-34.
- Schwerdtner, W.M., Waddington, D.H., and Stollery, G. 1974. Polycrystalline pseudomorphs as natural gauges of incremental paleostrain. *Neues Jahrbuch für Mineralogie, Monatshefte*, **314**: 174-182.
- Schwerdtner, W.M., Bennett, P.J., and Janes, T.W. 1977. Application of L–S fabric scheme to structural mapping and paleostrain analysis. *Canadian Journal of Earth Sciences*, **14** (5): 1012-1032. doi: 10.1139/e77-094.
- Schwerdtner, W.M., Mawer, C.K., and Hubbs, A.F. 1981. Geology of the Gravenhurst Region, Grenville Structural Province, Ontario: preliminary mapping results. *In* Current Research, part B. Geological Survey of Canada, Paper 81-1B, 167-169.

- Schwerdtner, W.M., Stott, G.M., and Sutcliffe, R.H. 1983. Strain patterns of crescentic granitoid plutons in the Archean greenstone terrain of Ontario. *Journal of Structural Geology*, 5: 419-430.
- Schwerdtner, W.M., Riller, U.P., and Borowik, A. 2005. Structural testing of tectonic hypotheses by field-based analysis of distributed tangential shear: Examples from major high-strain zones in the Grenville Province and other parts of the southern Canadian Shield. *Canadian Journal of Earth Sciences*, **42**: 1927-1947.
- Schwerdtner, W.M., Lu, S.J., and Yang, J.F. 2010a. Wall-rock structure at the present contact surfaces between repeatedly deformed thrust sheets, Grenville orogen of central Ontario, Canada. *Canadian Journal of Earth Sciences*, **47**: 875-899.
- Schwerdtner, W.M., Lu, S.J. and Landa, D., 2010b. S/Z buckle folds as shear-sense indicators in the ductile realm: Field examples from the Grenville Province of Ontario and the Appalachians of South Carolina. *In: From Rodinia to Pangea: The Lithotectonic Record of the Appalachian Region*, Geological Society of America. *Edited by* Tollo, R.P., Bartholomew, M.J., Hibbard, J.P., and Karabinos, P.M., Geological Society of America, Boulder, Colorado. *Memoir* **206**, 773-794. doi:10.1130/2010.1206 (30).
- Schwerdtner, W.M., Rivers, T., Zeeman, B., Wang, C.C., Tsolas, J., Yang, J., and Ahmed, M. 2014. Post-convergent structures in lower parts of the 1090–1050 Ma (early-Ottawan) thrust-sheet stack, Grenville Province of Ontario, southern Canadian Shield. *Canadian Journal of Earth Sciences*, **51**: 243-265.
- Slagstad, T., Hamilton, M.A., Jamieson, R.A., and Culshaw, N.G. 2004. Timing and duration of melting in the mid orogenic crust: constraints from Pb-U (SHRIMP) data, Muskoka and

- Shawanaga domains, Grenville Province, Ontario. *Canadian Journal of Earth Sciences* **41**: 1339-1365.
- Soucy La Roche, R., Gervais, F., Tremblay, A., Crowley, J.L., and Ruffet, G. 2015. Tectono-metamorphic history of the eastern Taureau shear zone, Mauricie area, Québec: Implications for the exhumation of the mid-crust in the Grenville Province. *Precambrian Research*, 257: 22-46.
- Steltenpohl, M.G., and Bartley, J.M. 1988. Cross folds and back folds in the Ofoten-Tysfjord area, north Norway, and their significance for Caledonian tectonics. *Geological Society of America Bulletin*, **100**: 140-151.
- Sullivan, W.A. 2013. L tectonites. *Journal of Structural Geology*, **50**: 161-175.
- Tikoff, B. and Teyssier, C. 1994. Strain modelling of displacement-field partitioning in transpressional orogens. *Journal of Structural Geology*, **16**: 1575-1588.
- Tollo, R.P., Corriveau, L., McLelland, J., and Bartholomew, M.J. 2004. Proterozoic evolution of the Grenville orogen in North America: An introduction. *In* Proterozoic Tectonic Evolution of the Grenville Orogen in North America, *edited by* R.P Tollo, L. Corriveau, J. McLelland, and M.J. Bartholomew. Geological Society of America, *Memoir* **197**, 1-18.
- van Berkel, J.T., and Schwerdtner, W.M. 1986. Structural geology of the Moon River area. Ontario Geological Survey Map P2954 (with marginal notes).
- Van De Kerckhove, S.R. 2015. Project Unit 13-029. Geologic studies in the Nepewassi domain, Central Gneiss Belt, Grenville Province. Summary of Field Work and Other Activities 2015, Ontario Geological Survey, Open File Report 6313, p. 20-1 to 20-8.
- van der Pluijm, B.A., and Carlson, K.A. 1989. Extension in the Central Metasedimentary Belt of the Ontario Grenville: timing and tectonic significance. *Geology*, **17**: 161-164.

- van Gool, J., Rivers, T., and Calon, T. 2008. The Grenville Front zone, Gagnon terrane, southwestern Labrador: configuration of a mid-crustal foreland fold-thrust belt. *Tectonics*, **27**: TC1004.doi:10.1029/2006TC002095
- Van Kranendonk, M.J. 1991. A magmatic sheet origin for thin meta-gabbroic anorthosite units in the Fishog Subdomain of the southern Central Gneiss Belt, Grenville Province, Ontario. *Canadian Journal of Earth Sciences* **28**: 431-446.
- Van Kranendonk, M.J, Kröner, A., Hegner, E., and Connelly, J. 2009. Age, lithology and structural evolution of the c. 3.53 Ga Theespruit Formation in the Tjakastad area, southwestern Barberton Greenstone Belt, South Africa, with implications for Archaean tectonics. *Chemical Geology*, **261**: 114-138.
- Van Reenen, D.D., Perchuk, L.L., Smit, C.A., Varlamov, D.A., Boshoff, R., Huizenga, J.M., Gerya, T.V. 2004. Structural and P-T evolution of a major cross fold in the Central Zone of the Limpopo high-grade terrain, South Africa. *Journal of Petrology*, **45**: 1413-1439.
- Venkat-Ramani, M., and Tikoff, B. 2002. Physical models of transtensional folding. *Geology*, **30**: 523-526.
- Vines, J.A., and Law, R.D. 2000. Emplacement of the Santa Rita Flat pluton as a pluton-scale saddle reef. *Geology*, **28**: 1115-1118.
- Waddington, D.H., 1973. Foliation and mineral lineation in the Moon River synform, Grenville Structural Province, Ontario. M.Sc. thesis, University of Toronto, Toronto, Ontario.
- Weinberg, R.F., Veveakis, E., and Regnaud-Lieb, K. 2015. Compaction-driven melt segregation in migmatites. *Geology*, **43**: 471-474. doi:10.1130/G36562.1

- White, D.J., Easton, R.M., Culshaw, N.G., Milkereit, B., Forsyth, D.A., Carr, S., Green, A.G., and Davidson, A. 1994. Seismic images of the Grenville Orogen in Ontario. *Canadian Journal of Earth Sciences*, **31**: 293-307.
- White, D.J., Forsyth, A.D., Asudeh, I., Carr, S.D., Wu, H., Easton, R.M., and Mereu, R.F. 2000. A seismic-based cross-section of the Grenville orogen in southern Ontario and western Québec. *Canadian Journal of Earth Sciences*, **37**: 183-192.
- Whitney, D.L., and Evans, B.W. 2010. Abbreviations for names of rock-forming minerals. *American Mineralogist*, **95**: 185-187.
- Wodicka, N., Parrish, Ketchum, J.W.F., Jamieson, R.A. 2000. Grenvillian metamorphism of monocyclic rocks, Georgian Bay, Ontario, Canada: implications for convergence history. *Canadian Mineralogist*, **38**: 471-510.
- Zulauf, G. 1997. Constriction due to subduction: evidence for slab pull in the Mariánské Lázně complex (central European Variscides). *Terra Nova*, **9**: 232-236.
- Zulauf, J., and Zulauf, G. 2005. Coeval folding and boudinage in four dimensions. *Journal of Structural Geology*, **27**: 1061-1068.

Figure Captions

Fig. 1: Sketch map showing the location and extent of the Ottawa River Gneiss Complex (ORGC); modified from Rivers et al. (2012, their Fig. 2). The double-headed arrow labelled Big Bend shows the curvature of the Grenville Front and other major tectonic boundaries in the western Grenville Province. Box indicates position of Figure 2. AB – Allochthon Boundary, CAB – Composite Arc Belt, CABb – Composite Arc Belt boundary, F-AB – Frontenac-Adirondack Belt. *AP* – Algonquin Park; *BB* – Barry's Bay, *Mi* – Minden, *NB* – North Bay; *NY* – New York, *ON* – Ontario, *QC* – Québec.

Fig. 2: (a) Structural domain map for the western ORGC showing structural levels 1-5 of the thrust-sheet stack; (b) Sketch map showing km-scale cross-fold system in the western ORGC. The three groups of cross-folds are separated by fold abutments (see text for explanation); (c) NW-SE crustal-scale cross-section showing thrust stacking and crustal levels; (d) Crustal section reconfigured as a metamorphic core complex showing extensional structures. (a) and (c) redrawn and slightly modified from Slagstad et al. (2004), after Culshaw et al. (1997), Ketchum and Davidson (2000), and Jamieson et al. (2010); (b) modified from Davidson et al. 1982, their Fig. 30.2; (d) modified from Rivers and Schwerdtner (2015). AB – Allochthon Boundary, BA – Bracebridge antiform, BNA – Britt-Nobel antiform, GFTZ – Grenville Front Tectonic Zone, GVA – Golden Valley antiform, MRS – Moon River synform, PSD – Parry Sound domain, SS – Seguin synform, TMBSZ – Twelve Mile Bay shear zone, WALA – Wawashkesh-Ahmic Lakes antiform.

Fig. 3: Examples of mesoscopic cross-folds (photographed approximately parallel to their hinge-lines) and shape fabrics in the ORGC; (a) Nearly symmetrical upright cross-folds, Moon River synform near MacTier, exit ramp from Hwy. 69 to Hwy 400; (b) *S*-asymmetric inclined cross-

folds at Georgian Bay shore, northwest of Parry Sound; (c) Inclined cross-fold, Hwy 69 north of Pointe au Baril Station; (d) Recumbent cross-folds at southeast termination of the Britt-Nobel antiform; (e) Sub-horizontal elongation lineation in grey gneiss defined by stretched, primary mineral constituents (total-strain indicators), hanging wall of the Allochthon Boundary near Shawanaga, northeast side of Hwy. 559; (f) Heterogeneously deformed, *An*-rich plagioclase pseudomorphs after *Grs*-rich garnet, Hwy. 118 east of Bracebridge. Low strain of pseudomorphs documents strain accumulated during exhumation and retrogression. Note xenoblastic garnet relics in the cores of some pseudomorphs. (a) and (c) from Rivers and Schwerdtner (2015).

Fig. 4: ‘Cartoon’ block diagram illustrating the geometry and bulk kinematic setup used in modelling ductile transtension, and orientation of resultant transtensional folds and triaxial strain ellipsoid (redrawn and modified from Fossen et al. 2013, their Fig. 1).

Fig. 5: Down-plunge projection of the northwest Moon River synform and vicinity (slightly modified from Gower 1992, his Fig. 11c, with the publisher’s permission). Outcrop-scale parasitic folds on the larger structure are not shown.

Fig. 6: (a) Buckle cross-folds of the Allochthon Boundary and its hanging wall and footwall southwest of the Britt-Nobel antiform, Pointe au Baril area (modified from Dickin et al. 2008, their Fig. 5, after Culshaw et al. 1994, their Figs. 1, 2); (b) Buckle cross-folds in the Key River area, western Britt domain (after Davidson et al. 1982, their Fig. 30.2).

Fig. 7: Sketch maps showing structures relating to the km-scale Grundy Lake synform and Pakesley antiform, Key River area, western Britt domain: (a) Distribution and attitudes of mesoscopic cross-folds and associated hinge-parallel elongation lineations; (b) Variation in M-, S- and Z-style in outcrop-scale parasitic cross-folds.

Fig. 8: Structural maps of the central Moon River synform (a) Foliation trend map (data from Lumbers and Vertolli, 2000a, b; Lumbers et al., 2000a, b, used with permission; domain boundary traces after Davidson et al. 1982); (b) Lineation trend-line map showing changes in plunge magnitude and reversals in plunge direction along strike (redrawn from Schwerdtner and van Berkel 1991, their Fig. 10, with the publisher's permission); (c) Sketch map showing the variation in M-S-Z style of mesoscopic buckle folds. Stereoplot shows poles to the folded fabric; vertical NE-SW-trending great circle is profile plane for this part of the regional Moon River synform; black square labelled CA is the cylindrical fold axis.

Fig. 9: Elongation lineations and mesoscopic cross-folds, central Moon River synform: (a) SE-trending, subhorizontal lineation defined by mafic-mineral streaks in a cross-fold hinge zone, central Bala; (b, c) Horizontal SE-trending quasi-cylindrical segments of upright cross-folds, same locality; (d) Steeply NW-plunging cross-fold (white arrow), northbound lanes of Hwy. 400, just south of exit to Hwy 69 near MacTier; (e) Oval cross-sections of sheath-like folds, same locality as (d); (f) Stretched plagioclase pseudomorphs after garnet define subhorizontal elongation lineation in SE-trending foliation, central Bala.

Fig. 10: Field photographs of the attenuated, leucosome-rich, retrogressed straight-gneiss foliation deformed by cross-folds, Muskoka domain: (a) High-strain monzonitic straight gneiss with amphibolite layers containing *Hbl*±*Bt*; (b-c) Granodioritic to monzonitic straight gneiss with amphibolite layers, with tadpole-shaped amphibolite inclusion in foliation in b, and folded clinopyroxenite layers partially retrogressed to *Hbl* in *Bt*-bearing straight-gneiss fabric in c; (d) Metre-scale boudinage of amphibolite in straight-gneiss layering; (e) Dismembered amphibolite layers exhibiting cm-scale boudinage with leucosome in boudin necks, indicating formation

under suprasolidus conditions; (f) *Opx* boudin surrounded by *Hbl* in straight-gneiss fabric. From Rivers and Schwerdtner (2015).

Fig. 11: Lower-hemisphere stereoplots (redrawn from Waddington 1973) showing angular relationships between mesoscopic structural elements and the principal directions of magnetic-susceptibility anisotropy, determined in small cylinders with a torque meter (King and Rees 1962). (a) Gently-plunging fold in hornblende-bearing gneiss near junction of MR 18 and 189, Bala; (b) NW-plunging, mesoscopic fold, same locality as (a); (c) NW-plunging mesoscopic fold near the junction between Hwys. 400 and 69, MacTier. The stereo-plot in (a) is referred to mesoscopic fabric directions, those in (b-c) are projected onto a geographic reference frame.

Fig. 12: (a) Sketch map of the western ORGC and adjacent Composite Arc Belt (modified from Davidson 1984a, his Fig. 1) showing regional cross-folds. Thick dashed blue arrows show inferred limits and orientations of the pure shear and simple shear parts of the transtension tensor. The Muskoka domain is highlighted in yellow, and the Moon River, Seguin and Wallace synforms (MRS, SS, WS) and Bracebridge antiform (BA) are shown. The Wallace synform is defined in this paper. Foliation mega-boudins (dark yellow; CL – Clear Lake, G – Germania, RL – Rockaway Lake) are weakly strained granulite-facies remnants within the Muskoka domain that escaped pervasive amphibolite-facies retrogression (Schwerdtner et al. 2014; Rivers and Schwerdtner 2015). AB – Allochthon Boundary, BSZ – Bancroft shear zone, CAB – Composite Arc Belt, CABb – Composite Arc Belt boundary zone, GF – Grenville Front, GFTZ – Grenville Front Tectonic Zone, MB – Mulock batholith (the northern part of which is deflected into a NE-trend in the GFTZ; see text for discussion). (b) Two-dimensional geometric sketch showing orientations and relative magnitudes of the pure-shear and simple-shear parts of the transtension

tensor (dashed blue arrows), the divergence angle α and the divergence vector parallel to the axial trace of the Moon River synform.

Fig. 13: Sketches of vertical magnetic-gradient patterns of first-order cross-folds; (a) Moon River synform; (b) Wallace synform and other large cross-folds near Whitney. The intensity of the vertical magnetic-gradient ranges from +3 to -3 nT.km⁻¹. Simplified from Gupta (1991b), Ontario Geological Survey Map 2591; reproduced with permission.

Fig. 14: Mesoscopic structures and fabrics in cross-folded orthogneiss, (a-f) in northwest Wallace synform along Hwy. 60, and (g) in Key River area. (a) NE-dipping L>S fabric in grey orthogneiss with stretched leucosome; (b) Profile view of recumbent cross-folded foliation in orthogneiss with coeval leucosome; (c) Straight traces of deformed leucosome in orthogneiss on a horizontal surface parallel to upright SE-striking foliation; (d) Straight traces of leucosome veins on a foliation surface in orthogneiss showing the mineral-streak lineation; (e) Recumbent cross-folds in grey orthogneiss cut by a diffuse pegmatite sheet; (f) Recumbent cross-folds in grey orthogneiss with axial-planar compaction melt bands and hinge-parallel elongation lineations in grey gneiss; (g) SE-trending L-fabric parallel to cross-fold hinges, Key River area, Britt domain (note lack of leucosome).

Fig. 15: Structural sketch map of the West Smith antiform showing hinge-line trace and gentle plunge of fold closure parallel to local elongation lineations (for location, see Fig. 13). Basemap redrawn from Google Maps.

Fig. 16: Mesoscopic structures and fabrics from hinge zone of West Smith antiform, Hwy. 60 northwest of Whitney. (a) Gently SE-plunging, highly strained *Pl* pseudomorphs after *Grt* porphyroblasts in mafic gneiss, 50 m west of Pinetree Lake portage, east side of Hwy 60; (b) SW-dipping, foliated and veined monzodiorite gneiss with *Hbl-Bt* (amphibolite-facies; pink unit

at top) in contact with lineated grey straight gneiss (grey unit at bottom), west side of Hwy. 60 near Pinetree Lake portage; (c-d) Gently SE-plunging lineation defined by stretched megacrysts in monzonitic orthogneiss, hinge zone of West Smith antiform; (e) Folded contact between purely lineated monzodioritic orthogneiss (light grey) and amphibolite (dark grey), cut by late-tectonic pegmatite with pinch-and-swell structures (below scale bar) indicating subhorizontal extension; same locality as b; (f) Gently SE-plunging elongate *Pl* pseudomorphs after *Grs*-rich garnet porphyroblasts in metagabbro, ~500 m south of Brewer Lake, hinge zone of West Smith antiform.

Fig. 17: Modifications to standard transtension model (a; after Sanderson and Marchini 1984; Jones et al. 2004; Fossen et al. 2013) in which transtensional folds are upright cylindrical structures with horizontal hinge lines; (b-c) applications to the ORGC. (b) Inclined zone of ductile transtension with ‘regional’ heterogeneous strain due to variable vertical thinning, strike-slip displacement and horizontal extension within ductile zone (greatest in centre and declining towards margins), and ‘local’ heterogeneous strain due to viscosity contrasts in deformed multilayer sequence, illustrated by pinch-and-swell structures; transtensional folds have plunging hinge lines and non-cylindrical shapes. (c) Wedge-shaped zone of inclined ductile transtension with volume increase due to incorporation of material from dynamically retrogressed footwall due to strain softening and possible syntectonic magma addition. An additional source of strain heterogeneity in (b) and (c) would be present if the base of the ductile zone did not remain horizontal. Note that transtensional strain in the middle of the ductile zone in (b) and (c) may be approximately represented by the standard model. AD - Algonquin domain, BD - Britt domain, CAB - Composite Arc Belt, MD - Muskoka domain.

Fig. 18: (a) Map of part of ORGC and adjacent Composite Arc Belt configured in terms of the core, detachment and cover of a metamorphic core complex, with at right oriented sketches showing inferred geometry of oblique extension in the detachment zone, Composite Arc Belt boundary, and Composite Arc Belt. Variations in divergence angle α , orientation of divergence vector, and attitudes and magnitudes of the pure-shear and simple-shear parts of the transtension tensor are shown. In the detachment zone of the ORGC, the resolved vector of pure-shear-dominated ductile transtension is towards the southeast (large red arrow), in the Composite Arc Belt boundary transtensional deformation is distributed in discrete extensional and simple shear parts (black arrow and shear couple respectively), and in the extensional Bancroft shear zone in the northern Composite Arc Belt the displacement vector is approximately dip-slip (black arrows on shear plane). (b) Crustal-scale cross-section showing the location, surface extent and projection of the detachment zone to depth. Axial traces of most cross-folds are approximately parallel to the line of section. (c) Block diagram showing trace of Composite Arc Belt boundary (CABb) around the Big Bend, orientations of isolated reclined S-shaped cross-folds, and their relation to the resolved vector of pure-shear-dominated transtension in the underlying detachment zone of the ORGC. Map and cross-section in (a) and (b) modified from Slagstad et al. (2004), (c) modified from Schwerdtner et al. (2010b).

Figure 1

Fig. 1: Sketch map showing the location and extent of the Ottawa River Gneiss Complex (ORG); modified from Rivers et al. (2012, their Fig. 2). The double-headed arrow labelled Big Bend shows the curvature of the Grenville Front and other major tectonic boundaries in the western Grenville Province. Box indicates position of Figure 2. AB – Allochthon Boundary, CAB – Composite Arc Belt, CABb – Composite Arc Belt boundary, F-AB – Frontenac-Adirondack Belt. AP – Algonquin Park; BB – Barry's Bay, Mi – Minden, NB – North Bay; NY – New York, ON – Ontario, QC – Québec.

100x71mm (300 x 300 DPI)

Figure 2

Fig. 2: (a) Structural domain map for the western ORGC showing structural levels 1-5 of the thrust-sheet stack; (b) Sketch map showing km-scale cross-fold system in the western ORGC. The three groups of cross-folds are separated by fold abutments (see text for explanation); (c) NW-SE crustal-scale cross-section showing thrust stacking and crustal levels; (d) Crustal section reconfigured as a metamorphic core complex showing extensional structures. (a) and (c) redrawn and slightly modified from Slagstad et al. (2004), after Culshaw et al. (1997), Ketchum and Davidson (2000), and Jamieson et al. (2010); (b) modified from Davidson et al. 1982, their Fig. 30.2; (d) modified from Rivers and Schwerdtner (2015). AB – Allochthon Boundary, BA – Bracebridge antiform, BNA – Britt-Nobel antiform, GFTZ – Grenville Front Tectonic Zone, GVA – Golden Valley antiform, MRS – Moon River synform, PSD – Parry Sound domain, SS – Seguin synform, TMBSZ – Twelve Mile Bay shear zone, WALA – Wawashkesh-Ahmic Lakes antiform.

177x206mm (300 x 300 DPI)

Figure 3

Fig. 3: Examples of mesoscopic cross-folds (photographed approximately parallel to their hinge-lines) and shape fabrics in the ORGC; (a) Nearly symmetrical upright cross-folds, Moon River synform near MacTier, exit ramp from Hwy. 69 to Hwy 400; (b) S-asymmetric inclined cross-folds at Georgian Bay shore, northwest of Parry Sound; (c) Inclined cross-fold, Hwy 69 north of Pointe au Baril Station; (d) Recumbent cross-folds at southeast termination of the Britt-Nobel antiform; (e) Sub-horizontal elongation lineation in grey gneiss defined by stretched, primary mineral constituents (total-strain indicators), hanging wall of the Allochthon Boundary near Shawanaga, northeast side of Hwy. 559; (f) Heterogeneously deformed, An-rich plagioclase pseudomorphs after Grs-rich garnet, Hwy. 118 east of Bracebridge. Low strain of pseudomorphs documents strain accumulated during exhumation and retrogression. Note xenoblastic garnet relics in the cores of some pseudomorphs. (a) and (c) from Rivers and Schwerdtner (2015).

176x218mm (300 x 300 DPI)

Draft

Fig. 4: 'Cartoon' block diagram illustrating the geometry and bulk kinematic setup used in modelling ductile transtension, and orientation of resultant transtensional folds and triaxial strain ellipsoid (redrawn and modified from Fossen et al. 2013, their Fig. 1).

173x103mm (300 x 300 DPI)

Figure 5

Fig. 5: Down-plunge projection of the northwest Moon River synform and vicinity (slightly modified from Gower 1992, his Fig. 11c, with the publisher’s permission). Outcrop-scale parasitic folds on the larger structure are not shown.

112x79mm (300 x 300 DPI)

Figure 6

Fig. 6: (a) Buckle cross-folds of the Allochthon Boundary and its hanging wall and footwall southwest of the Britt-Nobel antiform, Pointe au Baril area (modified from Dickinson et al. 2008, their Fig. 5, after Culshaw et al. 1994, their Figs. 1, 2); (b) Buckle cross-folds in the Key River area, western Britt domain (after Davidson et al. 1982, their Fig. 30.2).

106x185mm (300 x 300 DPI)

Figure 7

Fig. 7: Sketch maps showing structures relating to the km-scale Grundy Lake synform and Pakesley antiform, Key River area, western Britt domain: (a) Distribution and attitudes of mesoscopic cross-folds and associated hinge-parallel elongation lineations; (b) Variation in M-, S- and Z-style in outcrop-scale parasitic cross-folds.

125x216mm (300 x 300 DPI)

Figure 8

Fig. 8: Structural maps of the central Moon River synform (a) Foliation trend map (data from Lumbers and Vertolli, 2000a, b; Lumbers et al., 2000a, b, used with permission; domain boundary traces after Davidson et al. 1982); (b) Lineation trend-line map showing changes in plunge magnitude and reversals in plunge direction along strike (redrawn from Schwerdtner and van Berkel 1991, their Fig. 10, with the publisher's permission); (c) Sketch map showing the variation in M-S-Z style of mesoscopic buckle folds. Stereoplot shows poles to the folded fabric; vertical NE-SW-trending great circle is profile plane for this part of the regional Moon River synform; black square labelled CA is the cylindrical fold axis.

185x198mm (300 x 300 DPI)

Figure 9

Fig. 9: Elongation lineations and mesoscopic cross-folds, central Moon River synform: (a) SE-trending, subhorizontal lineation defined by mafic-mineral streaks in a cross-fold hinge zone, central Bala; (b, c) Horizontal SE-trending quasi-cylindrical segments of upright cross-folds, same locality; (d) Steeply NW-plunging cross-fold (white arrow), northbound lanes of Hwy. 400, just south of exit to Hwy 69 near MacTier; (e) Oval cross-sections of sheath-like folds, same locality as (d); (f) Stretched plagioclase pseudomorphs after garnet define subhorizontal elongation lineation in SE-trending foliation, central Bala.

144x218mm (300 x 300 DPI)

Figure 10

Fig. 10: Field photographs of the attenuated, leucosome-rich, retrogressed straight-gneiss foliation deformed by cross-folds, Muskoka domain: (a) High-strain monzonitic straight gneiss with amphibolite layers containing Hbl±Bt; (b-c) Granodioritic to monzonitic straight gneiss with amphibolite layers, with tadpole-shaped amphibolite inclusion in foliation in b, and folded clinopyroxenite layers partially retrogressed to Hbl in Bt-bearing straight-gneiss fabric in c; (d) Metre-scale boudinage of amphibolite in straight-gneiss layering; (e) Dismembered amphibolite layers exhibiting cm-scale boudinage with leucosome in boudin necks, indicating formation under suprasolidus conditions; (f) Opx boudin surrounded by Hbl in straight-gneiss fabric. From Rivers and Schwerdtner (2015).

173x199mm (300 x 300 DPI)

Figure 11

Fig. 11: Lower-hemisphere stereoplots (redrawn from Waddington 1973) showing angular relationships between mesoscopic structural elements and the principal directions of magnetic-susceptibility anisotropy, determined in small cylinders with a torque meter (King and Rees 1962). (a) Gently-plunging fold in hornblende-bearing gneiss near junction of MR 18 and 189, Bala; (b) NW-plunging, mesoscopic fold, same locality as (a); (c) NW-plunging mesoscopic fold near the junction between Hwys. 400 and 69, MacTier. The stereo-plot in (a) is referred to mesoscopic fabric directions, those in (b-c) are projected onto a geographic reference frame.

148x146mm (300 x 300 DPI)

Figure 12

Fig. 12: (a) Sketch map of the western ORGC and adjacent Composite Arc Belt (modified from Davidson 1984a, his Fig. 1) showing regional cross-folds. Thick dashed blue arrows show inferred limits and orientations of the pure shear and simple shear parts of the transtension tensor. The Muskoka domain is highlighted in yellow, and the Moon River, Seguin and Wallace synforms (MRS, SS, WS) and Bracebridge antiform (BA) are shown. The Wallace synform is defined in this paper. Foliation mega-boudins (dark yellow; CL – Clear Lake, G – Germania, RL – Rockaway Lake) are weakly strained granulite-facies remnants within the Muskoka domain that escaped pervasive amphibolite-facies retrogression (Schwerdtner et al. 2014; Rivers and Schwerdtner 2015). AB – Allochthon Boundary, BSZ – Bancroft shear zone, CAB – Composite Arc Belt, CABb – Composite Arc Belt boundary zone, GF – Grenville Front, GFTZ – Grenville Front Tectonic Zone, MB – Mulock batholith (the northern part of which is deflected into a NE-trend in the GFTZ; see text for discussion). (b) Two-dimensional geometric sketch showing orientations and relative magnitudes of the pure-shear and simple-shear parts of the transtension tensor (dashed blue arrows), the divergence angle α and the divergence vector parallel to the axial trace of the Moon River synform.

141x171mm (300 x 300 DPI)

Draft

Figure 13

Fig. 13: Sketches of vertical magnetic-gradient patterns of first-order cross-folds; (a) Moon River synform; (b) Wallace synform and other large cross-folds near Whitney. The intensity of the vertical magnetic-gradient ranges from $+3$ to -3 nT.km^{-1} . Simplified from Gupta (1991b), Ontario Geological Survey Map 2591; reproduced with permission.

133x201mm (300 x 300 DPI)

Figure 14

Fig. 14: Mesoscopic structures and fabrics in cross-folded orthogneiss, (a-f) in northwest Wallace synform along Hwy. 60, and (g) in Key River area. (a) NE-dipping L>S fabric in grey orthogneiss with stretched leucosome; (b) Profile view of recumbent cross-folded foliation in orthogneiss with coeval leucosome; (c) Straight traces of deformed leucosome in orthogneiss on a horizontal surface parallel to upright SE-striking foliation; (d) Straight traces of leucosome veins on a foliation surface in orthogneiss showing the mineral-streak lineation; (e) Recumbent cross-folds in grey orthogneiss cut by a diffuse pegmatite sheet; (f) Recumbent cross-folds in grey orthogneiss with axial-planar compaction melt bands and hinge-parallel elongation lineations in grey gneiss; (g) SE-trending L-fabric parallel to cross-fold hinges, Key River area, Britt domain (note lack of leucosome).

151x207mm (300 x 300 DPI)

Figure 15

Fig. 15: Structural sketch map of the West Smith antiform showing hinge-line trace and gentle plunge of fold closure parallel to local elongation lineations (for location, see Fig. 13). Basemap redrawn from Google Maps.

114x117mm (300 x 300 DPI)

Figure 16

Fig. 16: Mesoscopic structures and fabrics from hinge zone of West Smith antiform, Hwy. 60 northwest of Whitney. (a) Gently SE-plunging, highly strained Pl pseudomorphs after Grt porphyroblasts in mafic gneiss, 50 m west of Pinetree Lake portage, east side of Hwy 60; (b) SW-dipping, foliated and veined monzodiorite gneiss with Hbl-Bt (amphibolite-facies; pink unit at top) in contact with lineated grey straight gneiss (grey unit at bottom), west side of Hwy. 60 near Pinetree Lake portage; (c-d) Gently SE-plunging lineation defined by stretched megacrysts in monzonitic orthogneiss, hinge zone of West Smith antiform; (e) Folded contact between purely lineated monzodioritic orthogneiss (light grey) and amphibolite (dark grey), cut by late-tectonic pegmatite with pinch-and-swell structures (below scale bar) indicating subhorizontal extension; same locality as b; (f) Gently SE-plunging elongate Pl pseudomorphs after Grs-rich garnet porphyroblasts in metagabbro, ~500 m south of Brewer Lake, hinge zone of West Smith antiform.

169x201mm (300 x 300 DPI)

Draft

Figure 17

Fig. 17: Modifications to standard transtension model (a; after Sanderson and Marchini 1984; Jones et al. 2004; Fossen et al. 2013) in which transtensional folds are upright cylindrical structures with horizontal hinge lines; (b-c) applications to the ORGC. (b) Inclined zone of ductile transtension with 'regional' heterogeneous strain due to variable vertical thinning, strike-slip displacement and horizontal extension within ductile zone (greatest in centre and declining towards margins), and 'local' heterogeneous strain due to viscosity contrasts in deformed multilayer sequence, illustrated by pinch-and-swell structures; transtensional folds have plunging hinge lines and non-cylindrical shapes. (c) Wedge-shaped zone of inclined ductile transtension with volume increase due to incorporation of material from dynamically retrogressed footwall due to strain softening and possible syntectonic magma addition. An additional source of strain heterogeneity in (b) and (c) would be present if the base of the ductile zone did not remain horizontal. Note that transtensional strain in the middle of the ductile zone in (b) and (c) may be approximately represented by the standard model. AD - Algonquin domain, BD - Britt domain, CAB - Composite Arc Belt, MD - Muskoka domain.

105x205mm (300 x 300 DPI)

Draft

Figure 18

Fig. 18: (a) Map of part of ORGC and adjacent Composite Arc Belt configured in terms of the core, detachment and cover of a metamorphic core complex, with at right oriented sketches showing inferred geometry of oblique extension in the detachment zone, Composite Arc Belt boundary, and Composite Arc Belt. Variations in divergence angle α , orientation of divergence vector, and attitudes and magnitudes of the pure-shear and simple-shear parts of the transtension tensor are shown. In the detachment zone of the ORGC, the resolved vector of pure-shear-dominated ductile transtension is towards the southeast (large red arrow), in the Composite Arc Belt boundary transtensional deformation is distributed in discrete extensional and simple shear parts (black arrow and shear couple respectively), and in the extensional Bancroft shear zone in the northern Composite Arc Belt the displacement vector is approximately dip-slip (black arrows on shear plane). (b) Crustal-scale cross-section showing the location, surface extent and projection of the detachment zone to depth. Axial traces of most cross-folds are approximately parallel to the line of section. (c) Block diagram showing trace of Composite Arc Belt boundary (CABb) around the Big Bend, orientations of isolated reclined S-shaped cross-folds, and their relation to the resolved vector of pure-shear-dominated

transtension in the underlying detachment zone of the ORGC. Map and cross-section in (a) and (b) modified from Slagstad et al. (2004), (c) modified from Schwerdtner et al. (2010b).

154x222mm (300 x 300 DPI)

Draft

Appendix 1

Ottawa River Gneiss Complex

Rivers and Schwerdtner (2015) introduced the name *Ottawa River Gneiss Complex* (ORGC) for a large area of gneissic rocks in the Grenville Province of Ontario and Quebec comprising the western part of the Central Gneiss Belt of Wynne-Edwards (1972). More specifically, the ORGC is situated between the Grenville Front Tectonic Zone and the Composite Arc and Frontenac–Adirondack belts (see Fig. 1, this paper). In their brief explanation for the introduction of the new name, Rivers and Schwerdtner (2015) referred to the present paper, in preparation at the time, for a more complete justification.

Meaning of the term Central Gneiss Belt no longer clear

In the original definition (Wynne-Edwards 1972, his Fig. 1), the Central Gneiss Belt, comprising Quebec and Ontario gneiss segments, encompassed rocks extending approximately 1000 km from 68°W in central Quebec to 80°W at Georgian Bay in Ontario, and was bounded to the northwest by the Grenville Front Tectonic Zone and to the southeast by the Central Metasedimentary Belt and Central Granulite Terrain. However, in recent usage its geographic extent has generally been limited to gneisses west of the Central Metasedimentary Belt (~76°W), i.e., excluding its ~500 km eastern half. Moreover, as discussed by Rivers (2014), for Wynne-Edwards (1972) the high-grade gneisses of the Central Gneiss Belt *were inferred to represent the depositional basement to the lower-grade, non-gneissic rocks in the adjacent Central*

Metasedimentary Belt (see his Fig. 4). During the 1980s to 1990s, the tectonic nature of the contact between the Central Gneiss Belt and the Central Metasedimentary Belt was recognized, their arc and back-arc tectonic settings deduced, and documentation of lithological and stratigraphic contrasts within the Central Metasedimentary Belt led to its tectonic subdivision into a collage of fault-bounded domains representing components of an accreted pericratonic arc system, and eventually to its renaming as the Composite Arc Belt and the western part of the Frontenac–Adirondack Belt (Easton 1992; Carr et al. 2000; Rivers et al. 2012).

Beginning in the early 1980s, mapping of the Central Gneiss Belt led to its subdivision into imbricated domains interpreted as remnants of a mid-crustal thrust stack (Davidson and Morgan 1981; Davidson et al. 1982; Culshaw et al. 1983; Davidson 1984a, 1984b; Indares and Martignole 1989), and subsequently, following modifications to their boundaries, the domains were grouped into ‘decks’ or ‘structural levels’ (e.g., Culshaw et al. 1983; 1994, 1997; Timmermann et al. 1997, 2002; Carr et al. 2000; Ketchum and Davidson 2000; White et al. 2000; Slagstad et al. 2004a; Jamieson et al. 2007, 2010). In addition, it is transected by a first-order tectonic boundary (Allochthon Boundary Thrust of Rivers et al. 1989; Allochthon Boundary of Carr et al. 2000; this paper), implying it consists of allochthonous and paraautochthonous parts that were later found to have different ages of prograde metamorphism and compressional deformation.

Evidence for extensional reworking in the Central Gneiss Belt was described by Culshaw et al. (1994, 1997) and extensional reworking of the originally thrust-sense Allochthon Boundary was confirmed by Ketchum et al. (1998). The collapsed crustal-scale architecture of the hinterland of the Grenville Orogen as a whole was recognized by Rivers (2008, 2012), and in the ORGC the large NW-trending lobes of the Muskoka domain, originally interpreted as thrust

structures (e.g., Davidson 1984a, 1984b), were re-interpreted as the axial traces of post-peak synformal cross-folds that developed during exhumation and retrogression (Rivers and Schwerdtner 2015; this paper), supporting interpretation of the Central Gneiss Belt as the exhumed core of a large metamorphic core complex in which the Muskoka domain forms the ductile detachment zone separating the core and carapace.

This summary of the evolving understanding of the western Grenville Province shows that both the geological significance and geographic boundaries of the term *Central Gneiss Belt* have fundamentally changed to such an extent that it now bears scant resemblance to the original definition. For these reasons, usage of the term is no longer clear, presenting two options for the way forward: redefinition, or proposal of a new name that fits current understanding. We have chosen the latter, given the questionable use of the adjective *Central* for a feature in the southwest Grenville Province, the requirement for the name to be based on a geographic feature, and considering that its overall shape is an elongate dome rather than a belt. The *Ottawa River Gneiss Complex* (ORGC) is named after the river that transects it, subdividing it into Ontario and Québec segments of approximately equal size (Fig. 1; this paper).

Use of the term ‘gneiss complex’

Proper usage of the term ‘complex’ is discussed in Article 37 of the North American Stratigraphic Code (North American Commission on Stratigraphic Nomenclature 2004). The article states that a complex is “*An assemblage or mixture of rocks of two or more genetic classes, i.e., igneous, sedimentary, or metamorphic, with or without highly complicated structure...*”, and continues “*Identification of an assemblage of diverse rocks as a complex is useful where mapping of each separate lithic unit is impractical at ordinary mapping scales...*”

(p. 1572). We contend that since the ORGC consists of a wide range of intimately intermingled metamorphic rocks of igneous and sedimentary origin, as well as a minority of undeformed cross-cutting units, it is an appropriate candidate for the term. Article 37 also states: “*Complex*” is unranked but commonly comparable to a suite or supersuite; therefore the term may be retained if subsequent detailed mapping distinguishes some or all of the component lithodemes or lithostratigraphic units.” (p. 1572). This statement is also applicable inasmuch as large parts of the ORGC remain only crudely mapped, but in areas of abundant exposure there has been considerable success in mapping coherent tectonic units such as domains and gneiss associations. Article 37 then proceeds to discuss *volcanic complex* and *structural complex* as examples. Gneiss complex is not mentioned, but examples of recent usage were obtained by searching the World Wide Web; e.g., Acasta Gneiss Complex in the Slave Province (Iizuka et al. 2007; Bowring et al. 2014), and Istaq Gneiss Complex in West Greenland (Nutman et al. 2013). The term appears to be in common usage when referring to Eoarchean / Hadean examples, but since it carries no age connotation there is no reason to exclude its application to younger examples. Finally, following the North American Stratigraphic Code, the first letters of all the words are capitalised (Article 7; p. 1562).

Present understanding of the Ottawa River Gneiss Complex

The Ottawa River Gneiss Complex is a large (~400 x 250 km) metamorphic core complex, the core principally consisting of polydeformed granulite-facies orthogneiss and subordinate paragneiss lithologies that is overlain across a melt-weakened amphibolite-facies ductile detachment zone by a lower grade cover divided into the Composite Arc and Frontenac-Adirondack belts (Rivers 2012; Rivers and Schwerdtner 2015). Integration of field mapping and

seismic imaging suggests the detachment zone is several kilometres thick – and that it is pervasively cross-folded (this paper). The present crustal architecture is interpreted in terms of gravitationally-driven collapse of a crustal-scale thrust stack in which the crust may originally have been of double thickness.

In terms of its composition, much of the ORGC is composed of heterogeneous 'grey' tonalitic, dioritic, and granodioritic gneiss with abundant amphibolite and more homogeneous 'pink' monzonitic gneiss. From the numerous geochemical studies, U-Pb ages, and Nd depleted-mantle (T_{DM}) model-age determinations, these orthogneisses are deduced to have originated in a continental arc on the southeast margin of Laurentia that existed from the late Paleoproterozoic until the onset of the Grenvillian Orogeny (Carr et al. 2000; Rivers and Corrigan 2000; Blein et al. 2003; Slagstad et al. 2004a, 2004b; Rivers et al. 2012). T_{DM} model ages range from Archean to Mesoproterozoic, indicating derivation from reworked crust of the adjacent Laurentian margin and juvenile arc / back-arc magmatism (e.g., Indares and Martignole 1989; Easton 1992; Carr et al. 2000; Dickin 2000; Dickin and Guo 2001; Herrell et al. 2006; Jamieson et al. 2007; Dickin et al. 2008; Slagstad et al. 2009; Dickin and North 2015).

Most structural and metamorphic studies of the ORGC have focussed on the prograde development of the Ottawa thrust stack (stacking sequence, P - T - t estimates of peak metamorphic conditions, location of the Allochthon Boundary etc.) that pre-dated its extensional collapse as a metamorphic core complex. On the basis of available data, the thrusting was early to mid Ottawa (~1080-1050 Ma) in much of the ORGC, but was of Rigolet age (~1000 Ma) near the Grenville Front (e.g., Carr et al. 2000; Rivers et al. 2012). Timing of gravitationally-driven extensional collapse and development of the metamorphic core complex is less well

constrained, but probably began at ~1060 Ma and continued until at least 1020 Ma (Rivers and Schwerdtner 2015; this paper).

Overlapping terminologies

Previous work on the high-grade gneisses in the ORGC has attempted to rationalise them in terms of criteria such as geochemical signature, crystallization and Nd model age, crustal architecture, stacking sequence, peak metamorphic grade etc., which have been used, individually or in tandem, as the basis for a subdivision. Each approach has spawned its own terminology designed to shed light on particular aspects of the ORGC. For instance, one is based on the locations of *Grenvillian* shear zones, another on the regional distribution of the *pre-Grenvillian* magmatic protoliths, and yet another combines elements of both, involving *Grenvillian* shear zones that controlled the distribution of *pre-Grenvillian* units. All provide insight into the origin and evolution of the ORGC. However, there is no a priori reason why boundaries based on different metrics should coincide, explaining the plethora of overlapping units and multiple levels of nomenclature.

Recognition of the ORGC as a metamorphic core complex, while opening up exciting new avenues of research (e.g., Whitney et al. 2013; Platt et al. 2015), undoubtedly adds to this confusion of overlapping nomenclature. For instance, we have argued that the division into allochthonous and parautochthonous belts, although a first-order tectonic feature of the orogen during thrusting and prograde metamorphism, was not important during orogenic collapse since the Allochthon Boundary was mechanically inactive at that time (Fig. 6, this paper). On the other hand, we have defined three new tectonic units, the core, detachment zone, and cover of the metamorphic core complex, all of which encompass one or more units previously defined on the

basis of other criteria. The detachment zone, for instance, along which ductile transtensional strain led to cross-folding during orogenic collapse, comprises the Muskoka, Shawanaga, and Britt domains, and to a lesser degree the Parry Sound domain, entities previously distinguished on the basis of a range of criteria such as lithology, structure, Nd model age etc., and that occupied different structural levels of the mid-crustal thrust stack during prograde metamorphism (Figs. 2 and 18; this paper). The deduction that they later all formed part of the detachment zone of the metamorphic core complex thus reinforces the general point that boundaries that were active at one stage of crustal evolution may have become inactive at a later stage.

In summary, the growing diversity of databases and interpretations has resulted in a collage of overlapping subdivisions and nomenclatures, each justified in its own right but based on limited criteria, scale, or stage of orogenic development. This inevitably presents a challenge to newcomers to Grenvillian geology, but appears to be the price of improved understanding in an increasingly data-rich world.

References:

- Blein, O., LaFlèche, M.R., and Corriveau, L. 2003. Geochemistry of the granulitic Bondy gneiss complex: a 1.5 Ga arc in the Central Metasedimentary Belt, Grenville Province, Canada. *Precambrian Research*, **120**: 193-217.
- Bowring, S., Chacko, T., Heaman, L., Reimink, J. 2014. Acasta Gneiss Complex. *Encyclopedia of Scientific Dating Methods*, Springer Science+Business Media, Dordrecht. 4p. DOI:10.1007/978-94-007-6326-5_91-1
- Carr, S.D., Easton, R.M., Jamieson, R.A., and Culshaw, N.G. 2000. Geologic transect across the Grenville orogen of Ontario and New York. *Canadian Journal of Earth Sciences*, **37**: 193-216.

- Culshaw, N.G., Davidson, A., and Nadeau, L. 1983. Structural subdivisions in the Parry Sound – Algonquin region, Ontario. *In* Current Research, Part B. Geological Survey of Canada, Paper 83 – 1B, 243-252.
- Culshaw, N G., Ketchum, J.W.F., Wodicka, N., and Wallace, P. 1994. Deep-crustal ductile extension following thrusting in the southwestern Grenville Province, Ontario. *Canadian Journal of Earth Sciences*, **31**: 160-175.
- Culshaw, N.G., Jamieson, R.A., Ketchum, J.W.F., Wodicka, N., Corrigan, D., and Reynolds, P.H. 1997. Transect across the southwestern Grenville orogen, Georgian Bay, Ontario: polystage convergence and extension in the lower orogenic crust. *Tectonics*, **16**: 966-982.
- Davidson, A., 1984a. Identification of ductile shear zones in the southwestern Grenville Province of the Canadian Shield. *In* Precambrian Tectonics Illustrated. *Edited by* A. Kröner and R. Greiling. E. Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, Germany, p. 263-279.
- Davidson, A., 1984b. Tectonic boundaries in the Grenville Province of the Canadian Shield. *Journal of Geodynamics*, **1**: 433-444.
- Davidson, A., and Morgan, W.C. 1981. Preliminary notes on the geology east of Georgian Bay, Grenville Structural Province, Ontario. *In* Current Research, Part A. Geological Survey of Canada, Paper 81-1A, p. 291-298.
- Davidson, A., Culshaw, N.G., Nadeau, L. 1982. A tectono-metamorphic framework for part of the Grenville Province, Ontario. *In* Current Research, Part A. Geological Survey of Canada, Paper 82-1A, p. 275-290.
- Dickin, A.P. 2000. Crustal formation in the Grenville Province: Nd-isotopic evidence. *Canadian Journal of Earth Sciences*, **37**: 165-181.
- Dickin, A.P., and Guo, A. 2001. The location of the Allochthon Boundary Thrust and the Archean-Proterozoic suture in the Mattawa area of the Grenville Province: Nd isotopic evidence. *Precambrian Research*, **107**: 31-43.
- Dickin, A.P., and North, R. 2015. Nd isotope mapping of the Allochthon Boundary Thrust on the shores of Georgian Bay, Ontario: significance for Grenvillian crustal evolution. *Geological Magazine*, **152**: 993-1008.
- Dickin A. P., Moretta, K. and North, R. 2008. Isotope mapping of the Allochthon Boundary Thrust in the Grenville Province of Ontario, Canada. *Precambrian Research*, **167**: 260-266.

- Easton, R.M. 1992. The Grenville Province and the Proterozoic history of central and southern Ontario. *In* *Geology of Ontario. Edited by* P.C. Thurston, H.R. Williams, R.H. Sutcliffe, and G.M. Stott: Ontario Geological Survey. Special Volume 4, p. 715-790.
- Herrell, M.K., Dickin, A.P., and Morris, W.A. 2006. A test of detailed Nd isotopic mapping in the Grenville Province: delineating a duplex thrust sheet in the Kipawa–Mattawa region. *Canadian Journal of Earth Sciences*, **43**: 421-432.
- Iizuka, T., Komiya, T., Ueno, Y., Katayama, I., Uehara, Y., Maruyama, S., Hirata, T., Johnson, S.P., Dunkley, D.J. 2007. Geology and zircon geochronology of the Acasta Gneiss Complex, northwestern Canada: New constraints on its tectonothermal history. *Precambrian Research*, **153**, 179-208.
- Indares, A. and Martignole, J. 1989. The Grenville Front south of Val d'Or. *Tectonophysics*, **157**: 199-228.
- Jamieson, R.A., Beaumont, C., Nguyen, M.H., and Culshaw, N.G. 2007. Synconvergent ductile flow in variable-strength ductile crust: Numerical models with application to the western Grenville orogen. *Tectonics* 26, TC5005, 1-23.
- Jamieson, R.A., Beaumont, C., Warren, C.J., and Nguyen, M.H. 2010. The Grenville Orogen explained? Applications and limitations of integrating numerical models with geological and geophysical data. *Canadian Journal of Earth Sciences*, **47**: 517-539.
- Ketchum, J.W.F., and Davidson, A. 2000. Crustal architecture and tectonic assembly of the Central Gneiss Belt, southwestern Grenville Province, Canada: a new interpretation. *Canadian Journal of Earth Sciences*, **37**: 217-234.
- Ketchum, J.W.F., Heaman, L.M., Krogh, T.E., Culshaw, N.G., and Jamieson, R.A. 1998. Timing and thermal influence of late orogenic extension in the lower crust: a U-Pb geochronological study from the southwest Grenville orogen, Canada. *Precambrian Research*, **89**: 25-45.
- North American Commission on Stratigraphic Nomenclature 2004. North American Stratigraphic Code. *AAPG Bulletin*, **89**, 1547-1591. DOI:10.1306/07050504129
- Nutman, A.P., Bennett, V.C., Friend, C.R.L., Hidaka, H., Yi, K., Lee, S.R., Kamiichi, T. 2013. The Itsaq Gneiss Complex of Greenland: Episodic 3900 to 3660 Ma juvenile crust formation and recycling in the 3660 to 3600 Ma Isukasian orogeny. *American Journal of Science*, **313**, 877-911. doi:10.2475/09.2013.03

- Platt, J. P., Behr, W.M., and Cooper, F.J., 2015. Metamorphic core complexes: windows into the mechanics and rheology of the crust. *Journal of the Geological Society, London*, **172**, 9-27. doi:10.1144/jgs2014-036
- Rivers, T. 2012. Upper-crustal orogenic lid and mid-crustal core complexes: signature of a collapsed orogenic plateau in the hinterland of the Grenville Province. *Canadian Journal of Earth Sciences* **49**: 1-42. doi: 10.1139/E11-114
- Rivers, T. 2014. Tectonic setting and evolution of the Grenville Orogen: an assessment of progress over the last 40 years. *Geoscience Canada*, **42**: 77-124
- Rivers, T., and Schwerdtner, W.S. 2015. Post-peak evolution of the Muskoka domain, western Grenville Province: ductile detachment in a crustal-scale metamorphic core complex. *Geoscience Canada*, 42: 403-436. doi.org/10.12789/geocanj.2015.42.080
- Rivers, T., Martignole, J., Gower, C.F., Davidson, A. 1989. New tectonic divisions of the Grenville Province, southeast Canadian Shield. *Tectonics*, **8**: 63-84. doi: 10.1029/TC008i001p00063.
- Rivers, T., Culshaw, N.G., Hynes, A., Indares, A. Jamieson, R., and Martignole, J. 2012. The Grenville Orogen – a post-Lithoprobe perspective. *In* *Tectonic Styles in Canada: the Lithoprobe perspective. Edited by J.A. Percival, F.A. Cook, and R.M. Cloves. Geological Association of Canada, Special Paper 49, p. 97-238.*
- Slagstad, T., Hamilton, M.A., Jamieson, R.A., and Culshaw, N.G. 2004a. Timing and duration of melting in the mid orogenic crust: constraints from Pb-U (SHRIMP) data, Muskoka and Shawanaga domains, Grenville Province, Ontario. *Canadian Journal of Earth Sciences* **41**: 1339-1365.
- Slagstad, T., Culshaw, N.G., Jamieson, R.A., and Ketchum, J.W.F. 2004b. Early Proterozoic tectonic history of the southwestern Grenville Province, Ontario: Constraints from geochemistry and geochronology of high-grade gneisses. *In* *Proterozoic Tectonic Evolution of the Grenville Orogen in North America. Edited by R.P. Tollo, L. Corriveau, J. McLelland, and M.J. Bartholemew, Geological Society of America, Memoir 197, p. 209-241.*
- Slagstad, T., Culshaw, N.G., Daly, J.S., and Jamieson, R.A. 2009. Western Grenville Province holds key to midcontinental Granite-Rhyolite Province enigma. *Terra Nova*, **21**: 181-187.

- Timmermann, H., Parrish, R.R., Jamieson, R.A., and Culshaw, N.G. 1997. Time of metamorphism beneath the Central Metasedimentary Belt boundary thrust zone, Grenville Orogen, Ontario: accretion at 1080 Ma? *Canadian Journal of Earth Sciences*, **34**: 1023-1029.
- Timmermann, H., Jamieson, R.A., Parrish, R.R., and Culshaw, N.G. 2002. Coeval migmatites and granulites, Muskoka domain, southwestern Grenville Province, Ontario. *Canadian Journal of Earth Sciences*, **39**: 239-258.
- White, D.J., Forsyth, A.D., Asudeh, I., Carr, S.D., Wu, H., Easton, R.M., and Mereu, R.F. 2000. A seismic-based cross-section of the Grenville orogen in southern Ontario and western Québec. *Canadian Journal of Earth Sciences*, **37**: 183-192.
- Whitney, D.L., Teyssier, C., Rey, P., and Buck, W.R. 2013. Continental and oceanic core complexes. *Geological Society of America Bulletin*, **125**: 273-298. <http://dx.doi.org/10.1130/B30754.1>
- Wynne-Edwards, H.R. 1972. The Grenville Province. *In* *Variations in Tectonic Styles in Canada*. Edited by R.A. Price and R.J.W. Douglas, Geological Association of Canada, Special Paper 11, p. 263-334.