Upper-crustal orogenic lid and mid-crustal core complexes: signature of a collapsed orogenic plateau in the hinterland of the Grenville Province [1]

Toby Rivers

Abstract: This paper provides a re-interpretation of the crustal architecture of the Ottawan hinterland of the Grenville Province in light of published empirical and numerical models of orogenic collapse. It is now seen as a series of high-grade, mid-crustal core complexes from tens to hundreds of kilometres across that are juxtaposed against segments of the lower grade upper and uppermost crust including the orogenic lid. Juxtaposition of such contrasting crustal levels, which exhibit decoupled tectonic styles corresponding to the orogenic infrastructure and suprastructure, respectively, is interpreted as a signature of the foundering of an orogenic plateau into a mid-crustal channel. Ottawan metamorphism progressed from granulite-facies in the mid crust at ~1090–1050 Ma, through amphibolite-facies in the upper crust at ~1050–1020 Ma, to heating to ≤500 °C in the uppermost crust at ~1020–980 Ma. This temporal progression is interpreted to reflect conductive heat transfer during collapse, as hot mid-crustal core complexes were exhumed against successively higher crustal levels. Exhumation was facilitated by substantial thinning and lengthening of the mid crust by simple- and pure-shear mechanisms. This was accompanied by wholesale boudinage of the brittle uppermost crust. Moreover, it may have resulted in excision of part of the ductile upper crust, which appears under-represented. Collapse was accompanied by diverse magmatic and hydrothermal products, their range of structural states implying that high-strain Ottawan deformation in the mid crust took place beneath an orogenic lid that was not penetratively deformed. Preliminary analysis indicates the Grenvillian inliers exhibit a comparable range of crustal levels to the Grenville Province, suggesting the orogenic plateau may have extended ~5000 km along strike from Labrador to Texas.

Résumé : Le présent article fournit une réinterprétation de l'architecture de la croûte dans l'arrière-pays ottavien de la province de Grenville à la lumière de modèles publiés, empiriques et numériques, de l'effondrement orogénique. Cette architecture est maintenant perçue comme une série de complexes noyaux du milieu de la croûte et à un degré élevé de métamorphisme; ces complexes, atteignant des dizaines, voire des centaines, de kilomètres de longueur, seraient juxtaposés à des segments de la croûte supérieur et sommitale, à métamorphisme moins élevé, incluant la couche orogénique de couverture. La juxtaposition de tels niveaux contrastants de la croûte, qui montrent des styles tectoniques découplés correspondant respectivement à l'infrastructure et à la superstructure, est interprétée comme une signature de l'effondrement d'un plateau orogénique dans un chenal du milieu de la croûte. Le métamorphisme ottavien a progressé du faciès des granulites dans le centre de la croûte vers ~1090–1050 Ma, au faciès des amphibolites dans la croûte supérieure vers ~1050–1020 Ma, puis à un échauffement ≤500 °C dans la plus haute partie de la croûte vers ~1020–980 Ma. Cette progression dans le temps est interprétée comme le reflet d'un transfert de chaleur conductrice durant l'effondrement, alors que des complexes chauds du milieu de la croûte étaient exhumés par rapport à des niveaux successivement plus élevés de la croûte. L'exhumation a été facilitée par une élongation et un amincissement importants de la croûte moyenne par de simples mécanismes de cisaillement; elle était accompagnée par le boudinage complet du sommet cassant de la croûte. De plus, cela peut avoir conduit à l'excision d'une partie de la croûte supérieure ductile, laquelle semble sous-représentée. L'effondrement a été accompagné de divers produits magmatiques et hydrothermaux, l'étendue de leurs états structuraux impliquant que la déformation ottavienne sous de grandes contraintes dans la croûte moyenne a eu lieu sous une couche de couverture orogénique qui n'a pas été déformée en profondeur. Une analyse préliminaire indique que les boutonnières grenvilliennes montrent une plage comparable de niveaux de la croûte dans la province de Grenville, suggérant que le plateau orogénique puisse s'être étendu ~5000 km le long de la direction, du Labrador au Texas.

[Traduit par la Rédaction]

Received 7 October 2010. Accepted 4 February 2011. Published at www.nrcresearchpress.com/cjes on 20 December 2011.

Corresponding Editor: Brendan Murphy.

T. Rivers. Department of Earth Sciences, Memorial University, St. John's, NL A1B 3X5, Canada.

E-mail for correspondence: trivers@mun.ca.

[1] This article is one of a series of papers published in *CJES Special Issue: In honour of Ward Neale* on the theme of Appalachian and Grenvillian geology.

Introduction

Extensional collapse of orogens

In a classic paper entitled “Extensional Collapse of Orogens”, Dewey (1988) used examples from several Mesozoic orogens that exhibit different stages of collapse after the formation of an orogenic plateau to empirically evaluate the mechanical processes leading to the demise of plateau topography in mountain belts. In part based on data from the Tibetan Plateau and the North American Basin and Range Province, in which collapse is at early and more advanced stages, respectively, he concluded that extensional collapse comprised the final phase of a three-phase orogenic evolution. In this scenario, the first phase involves lithospheric shortening and thickening over a period of ~15–20 Ma, leading to the development of a plateau with an elevation of ~3 km above sea level. A second phase of similar duration is dominated by broadening of the plateau and development of a system of conjugate strike-slip faults reflecting initiation of lateral tectonic extrusion. The third phase, of short (≥5 Ma) duration, involves rapid uplift to an elevation of 4–5 km above sea level, crustal heating, and the initiation of radial extensional collapse. In the Tibetan case, third-phase uplift began ~5 Ma ago at rates of 3–6 $\mathrm{mm \cdot a^{-1}}$ leading to a plateau elevation of ~5 km, and the kinematic velocities of its ongoing collapse were recently elegantly quantified by Zhang et al. (2004) using data from a network of global positioning system (GPS) receivers. The more advanced stage of collapse in the Basin and Range Province illustrates the reduction of local topography from ~4 to 1.5 km, dramatic thinning of the crust associated with the formation of metamorphic core complexes, and concomitant thinning of the underlying mantle lithosphere (see Mix et al. 2011 for details of timing and kinematics). Dewey (1988) argued that the maximum topographic elevation above local base level resulting from thrusting was ~3 km (i.e., that achieved in phase 1), and hence that the rapid uplift and heating in phase 3 had another cause. For the latter, he invoked convective thinning of the mantle lithosphere (e.g., Bird 1979; England et al. 1988). In the Himalaya–Tibet Orogen, where the subcontinental mantle lithosphere of the Eurasian and Indian plates is subvertical beneath the orogenic suture (e.g., DeCelles et al. 2002; Shi et al. 2004), slab break-off is another possibility (e.g., Gerya et al. 2004). In some respects the distinction may be moot, however, since both processes result in removal of mantle lithosphere and are followed by isostatic rebound, thereby providing a plausible explanation for the abrupt uplift of the plateau and the accompanying high geothermal gradient because of the rise of the asthenosphere (e.g., Klemperer 2006). These concepts are illustrated schematically in Fig. 1, which is adapted from a figure in Dewey’s paper.

Another important source of conceptual information for this study comes from the results of numerical modelling. Of particular relevance are the results of Rey et al. (2001, 2009), Vanderhaeghe and Teyssier (2001) and Teyssier and Whitney (2002) derived for an orogenic plateau with a weak mid-crustal rheology, and the more empirical study of ultra-hot orogens by Chardon et al. (2009). As discussed later, such models provide important constraints for the Grenville Orogen.

Gravitational versus tensional forces

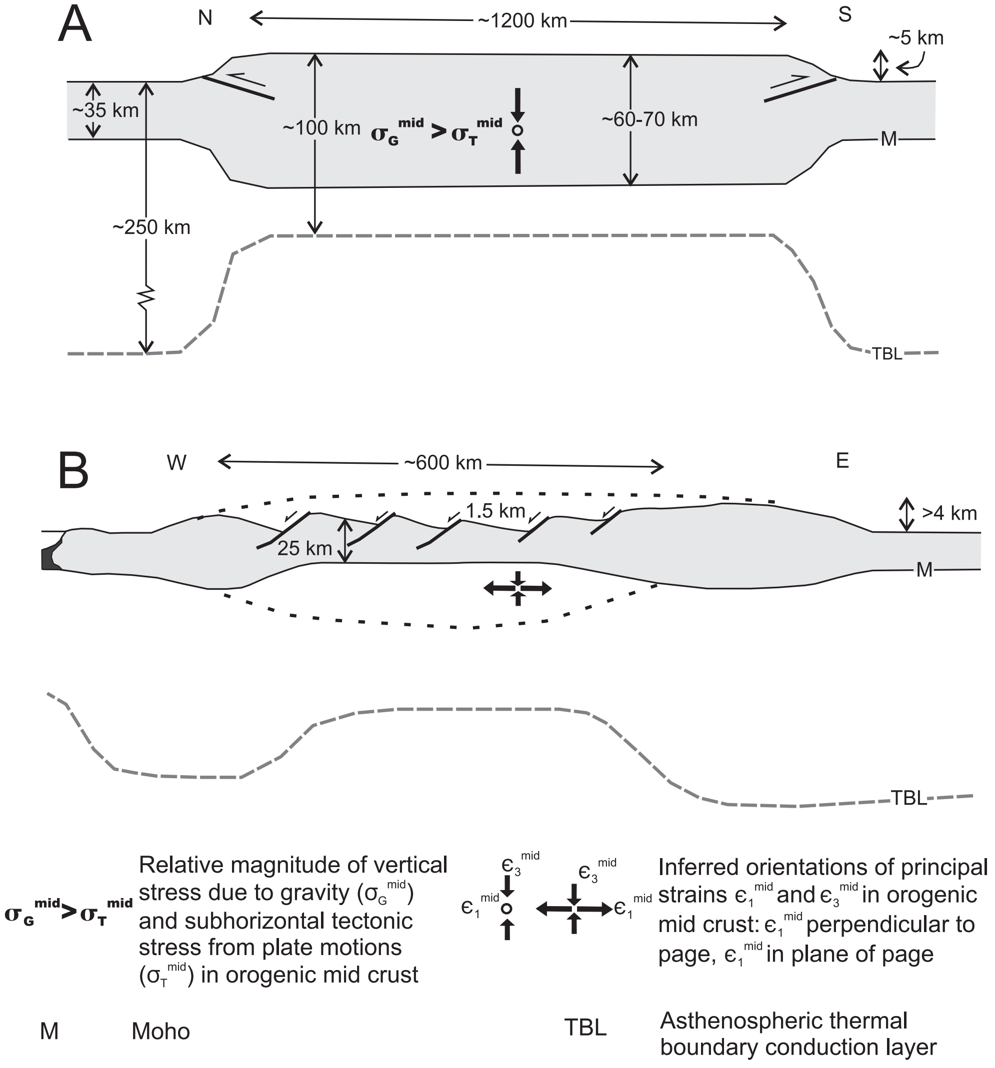
In general, the transition from crustal thickening to collapse in an orogenic belt is ascribed to three mechanisms: (*i*) an abrupt increase in topography, and hence gravitational potential energy, related to removal of the lithospheric mantle root; (*ii*) a decrease in the strength of the mid and (or) lower segments of the thickened crust because of thermal weakening; and (*iii*) a change in the regional plate tectonic regime from compression to tension (e.g., Molnar et al. 1993; Vanderhaeghe and Teyssier 2001). In this study, the subvertical downward stress resulting from gravitational potential energy acting on the weak mid crust, which is balanced by an equal upward stress exerted by the underlying stronger lower crust and mantle, is referred to as σ_G, and the subhorizontal stress resulting from plate-tectonic traction forces is referred to as σ_T. Dewey (1988) argued that when the topographic height of a plateau exceeds ~3 km, σ_G becomes the dominant force acting on the mid crust, i.e., $\sigma_G^{\mathrm{mid}} > \sigma_T^{\mathrm{mid}}$. In this situation, the plateau will begin to collapse by outward radial flow over time scales of a few tens of Ma. This concept was quantified with respect to the Tibetan Plateau by Clark and Royden (2000), who used topographic data to show that plateau elevation and local topography could be modelled as functions of the viscosity of the underlying mid crust.

In the discussion that follows, it will become desirable to distinguish between gravity-driven and tectonically driven strain of the mid crust. Because gravity is a body force (proportional to length^3), it is an order of magnitude larger than deviatoric stress (proportional to length^2) at any point in the Earth’s crust. Consequently, orogenic collapse is driven mainly by gravity, and tectonically induced tensional forces will be responsible for only modest amounts of crustal thinning. Nonetheless, inasmuch as such forces may contribute to the collapse process, it is appropriate to consider criteria to distinguish between the two. In this respect, timing and orientation of the normal-sense shear zones are crucial. Specifically, if the contractional structures driving crustal thickening are closely related in time to the normal-sense structures that cause crustal thinning, it is likely the latter are a result of gravitational forces. On the other hand, if the normal-sense structures significantly postdate compressional tectonics, they are more likely tensional in origin. In the former case, the normal-sense structures are typically restricted to the upper part of an overall compressional orogen and sole in the weak mid crust, implying that extension is limited to the overlying crust. Moreover, as noted, they ideally exhibit a radial pattern. On the other hand, where normal-sense shear zones form in a tensional setting, they may reach the Moho and they typically exhibit a preferred orientation reflecting the regional stress field.

Terminology

In this paper, the non-genetic term *normal-sense* rather than “extensional” is used to describe the kinematics of individual faults or shear zones that lengthen the crust in cases where the overall stress setting is unclear. Similarly, the non-genetic term *orogenic collapse* is used in preference to “gravitational collapse” and “extensional collapse” in the absence of precise knowledge of the regional stress field (while acknowledging that all extension is a function of gravitational forces acting on potential-energy gradients at some scale). Rey et al. (2001) proposed a distinction between gravity-

Fig. 1. Schematic cross-sections showing two stages of orogenic collapse in an overall convergent tectonic setting, modelled on the Tibetan Plateau and the Basin and Range Province of the North American Cordillera, respectively (after Dewey 1988). (A) Early phase of collapse illustrating double thickness orogenic crust, uplift of the plateau, and generation of ~5 km of local relief following the rise of the asthenospheric thermal boundary layer because of removal of lower lithospheric mantle by delamination or slab break-off. (B) Collapsing orogen exhibiting lower topographic relief and substantially thinner crust. σ_T^{mid} and σ_G^{mid} are inferred principal stresses in the mid crust owing to tectonic tractional forces and gravity, respectively; $\varepsilon_1{}^{mid}$ and $\varepsilon_3{}^{mid}$ are inferred principal strains in the mid crust.



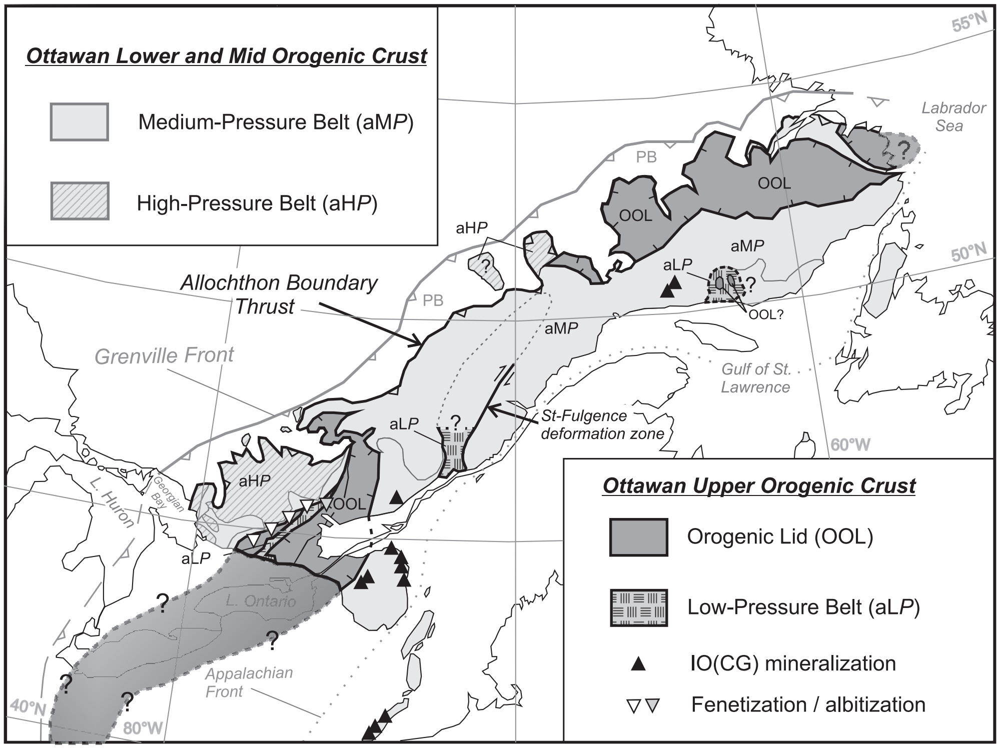
driven collapse and combined gravity- and tension-driven collapse in terms of their boundary conditions, referring to them as fixed-boundary and free-boundary collapse, respectively.

The term *orogenic lid* was introduced into the geological lexicon by Laubscher (1983) to describe topographically high areas of the eastern Alps that lack evidence for the main Tertiary Alpine metamorphism. Although metamorphic, the rocks making up the Alpine orogenic lid were metamorphosed in the Late Cretaceous (Eo-alpine stage), and in the Tertiary they formed the exhumed orogenic suprastructure above the hot metamorphic core. This concept was recently applied to the Grenville Province by Rivers (2008, 2009), who identified an orogenic lid in the Grenvillian hinterland that, although surrounded by upper amphibolite- to granulite-facies rocks, lacks evidence for coeval high-grade metamorphism and, hence, was interpreted to be part of the contemporary upper crust.

Aims of paper

This paper seeks to improve understanding of the geometry, kinematics, cooling history, and tectonic setting of orogenic collapse in the Ottawan hinterland of the Grenville Province. A substantial database exists on which to build, including $^{40}Ar/^{39}Ar$ geochronology, U–Pb geochronology of zircon, monazite, titanite, and rutile, pressure–temperature (*P–T*) estimations pertaining to Ottawan metmorphism, temperature–time (*T–t*) cooling curves, studies of deformation and magmatism associated with the Ottawan phase, crustal-scale reflection seismic experiments, and recognition of remnants of the upper orogenic crust (orogenic lid), but they have not been integrated into single consilient narrative. Following a brief review of geological evidence for the former existence of an Ottawan orogenic plateau, structural and metamorphic data and cooling histories from different crustal levels are described. Special attention is paid to the contrast-

Fig. 2. Map of the Grenville Province showing the distribution of Ottawan high-pressure rocks (aH*P* Belt), medium-pressure rocks (aM*P* Belt), low-pressure rocks (aL*P* Belt), and the Ottawan Orogenic Lid (OOL), with possible extensions of the OOL to the northeast and southwest (modified after Rivers 2008). Note the inferred extent of the OOL in relation to the mid crust exposed in the Grenville Province. IO (CG), iron ore – copper – gold type mineralization.



ing geological signals preserved in the cool uppermost orogenic crust comprising the orogenic lid, the L*P* (low pressure) upper crust, and the subjacent high-grade, M*P* (medium pressure) mid crust. Results from the three crustal levels are then integrated with the crustal architecture and concurrent magmatic and hydrothermal activity to provide a more nuanced model of orogenic collapse in the Grenville Province and to suggest possible extensions elsewhere in the Grenville Orogen. Finally, since the Grenville Orogen exhibits a more deeply exhumed section than is exposed in Mesozoic orogens, including those studied by Dewey (1988), the perspective gained is used to provide information on the deep levels of collapsed orogens.

The Grenville Orogen

The Grenville Orogen developed on southeast Laurentia during the late Mesoproterozoic to earliest Neoproterozoic Grenvillian Orogeny (~1090–980 Ma; e.g., Rivers et al. 1989, 2002; McLelland et al. 1996, 2010; Rivers 1997, 2008; Carr et al. 2000). It is widely believed to have been a collisional orogen at the centre of the supercontinent Rodinia (e.g., Li et al. 2008), although this has been disputed on paleomagnetic grounds (Evans 2009). The orogen subsequently became the site of late Neoproterozoic rifting that led to the formation of the Iapetus Ocean, and only part of it is preserved in North America. Several independent lines of evidence have been used to suggest that Amazonia was the rifted conjugate margin of the Grenville Orogen (e.g., Loewy et al. 2003; Tohver et al. 2004; Fisher et al. 2010), but an unambiguous fit currently remains out of reach rendering tectonic reconstructions open to debate.

On the basis of the spatial and temporal distribution of metamorphic rocks, the Grenvillian Orogeny has been subdivided into two orogenic phases: the ~1090–1020 Ma Ottawan phase that is principally exposed in the hinterland of the Grenville Orogen in the hanging wall of the Allochthon Boundary Thrust, and the ~1005–980 Ma Rigolet phase that occurs near the northwest margin in the hanging wall of the Grenville Front (e.g., Rivers et al. 2002; Rivers 2008, 2009). Emphasis in this paper is on the rocks in the orogenic hinterland in the hanging wall of the Allochthon Boundary Thrust, with special focus on the variations in their Ottawan structural and metamorphic signatures (Fig. 2). Following Rivers (2008), the orogenic hinterland is subdivided into four tectonic units reflecting their contrasting Ottawan metamorphic evolution, known as the allochthonous high-pressure belt (aH*P* Belt), the allochthonous medium-pressure belt (aM*P* Belt), the allochthonous low-pressure belt (aL*P* Belt), and the Ottawan Orogenic Lid (OOL). The term allochthonous is used to contrast with parautochthonous units in the northwestern Grenville Province and to imply that they were far-travelled during the Ottawan orogenic phase, but does not imply they were exotic to Laurentia. Brief descriptions of the tectonic characteristics of each unit are given after reviewing the evidence for a plateau in the orogenic hinterland during the Ottawan phase.

Orogenic plateau during the Ottawan phase

Evidence for the former existence — and subsequent collapse — of an orogenic plateau surmounting double thickness crust in the hinterland of the Grenville Orogen during the Ottawan orogenic phase remains circumstantial, but it has become much more compelling in the last two decades on the basis of a wide range of independent observations. Evidence includes (*i*) the width of the Grenville Province preserved in North America (≥600 km), all of which was derived from Laurentian crust, implying the full width of the Grenville Orogen was probably ≥ 1000 km; (*ii*) the long (~70 Ma) duration of the Ottawan orogenic phase; (*iii*) the identification of a first-order, crustal-scale shear zone along the length of the Grenville Province, the Allochthon Boundary Thrust (Rivers et al. 1989), which was active as a contractional structure in early Ottawan time (~1090–1040 Ma) and reworked as a normal-sense structure at ~1020 Ma (Ketchum et al. 1998); (*iv*) pressure estimates of ca. 1700 MPa (~50–60 km depth) for Ottawan eclogite-facies rocks, interpreted to be samples of the orogenic lower crust exhumed by thrusting along the Allochthon Boundary Thrust (e.g., Indares et al. 1998); (*v*) recognition of a transpressional shear zone system in the central Grenville Province that was active in Ottawan time and in part coeval with the Allochthon Boundary Thrust (Hébert et al. 2009); (*vi*) field evidence for widespread subhorizontal, normal-sense flow of the mid crust following Ottawan crustal thickening (e.g., Culshaw et al. 1994); (*vii*) quantification of the effect of melt weakening on the rheology of high-grade metamorphic rocks (e.g., Rosenberg and Handy 2005); (*viii*) improved understanding of the role played by a weak mid crust in defining the geometry of orogenic collapse (e.g., Rey et al. 2001, 2009; Vanderhaeghe and Teyssier 2001); (*ix*) recognition that peak Ottawan temperatures in the mid crust attained 750–850 °C throughout the hinterland of the Grenville Orogen, sufficient to induce widespread partial melting of metapelitic and quartzofeldspathic lithologies (e.g., Rivers 2008); (*x*) improved numerical models of orogenesis that shed light on possible boundary conditions of the Grenville Orogen (e.g., Jamieson et al. 2007, 2010); and (*xi*) recognition that the architecture of the exposed Grenvillian crust consists of mid-crustal horsts with the geometry of core complexes (e.g., Selleck et al. 2005; McLelland et al. 2009) juxtaposed against the Ottawan Orogenic Lid (Rivers 2008, 2009).

Taken together, these data and concepts support the existence of an orogenic plateau several hundred kilometres wide that developed on crust ≥60 km thick in early Ottawan time as a result of the imbrication of large thrust slices, mapped as terranes and domains, on and above the Allochthon Boundary Thrust. Thrust-sense displacements on this and other NE-striking, gently SE-dipping structures resulted in the tectonic transport of hot rocks from the lower and mid orogenic crust towards the orogen margin (Rivers 2009).

The presence of a major, coeval, NE-striking transpressional shear zone system located southeast of the Allochthon Boundary Thrust was recently reported by Hébert et al. (2009). Known as the St-Fulgence deformation zone (Fig. 2), this anastomosing network of shear zones is up to 50 km wide and can be traced for several hundred kilometres. Apart from its size, its importance can be gauged from its association with voluminous Ottawan intrusions. Hébert et al. (2009) determined that penetrative strain on the St-Fulgence deformation zone took place between ~1080 and 1030 Ma, but it remained a site of minor displacement until ~1010 Ma. Other evidence for approximately strike-parallel displacement comes from the Manicouagan Imbricate Zone, where high-*T* ductile gneisses with orogen-parallel lineations were interpreted as a lateral tectonic escape zone by Indares et al. (2000). Together they constitute evidence for phase 2 of Dewey's (1988) scenario for orogenic collapse, i.e., the development of a crustal-scale system of orogen-parallel transcurrent structures reflecting lateral tectonic extrusion beneath the orogenic plateau. This is discussed further in a later section.

The high temperature of the mid crust in Ottawan time, for which evidence is given by Rivers (2008), is inferred to be a result of the long duration of collision that permitted effective contributions to heating by conduction and radioactivity, in addition to viscous heating (Burg and Gerya 2005) and advected heat associated with the transport of hot thrust slices and emplacement of magmatic bodies. As a result, it is reasonable to infer that the observed widespread migmatization of mid-crustal quartzofeldspathic lithologies was a result of in situ dehydration-melting reactions. Such reactions would have induced profound rheological weakening of the mid crust, with the result that the Ottawan hinterland of the Grenville Orogen resembled a large hot long-duration orogen in the sense of Beaumont et al. (2004, 2006). The question of whether the orogenic plateau that formed during the Ottawan phase involved some form of channel flow has been discussed elsewhere (e.g., Jamieson et al. 2007, 2010; Rivers 2008, 2009) and is not specifically addressed here. Orogenic collapse, the topic of this paper, is inferred to have been a direct result of the formation of a plateau surmounting doubled crust and the concomitant melt-weakening of the mid crust.

Ottawan mid crust — aM*P* Belt

Exposed along the length of the Grenville Province for ~2000 km from Labrador through Quebec to Ontario and New York State (Fig. 2), the orogenic mid crust in the aM*P* Belt is characterized by Ottawan upper amphibolite- to granulite-facies rocks with gneissic fabrics, typified by the widespread subassemblages: Grt–Sil–Kfs–L in quartzofeldspathic–metapelitic lithologies and Cpx–Opx–Pl ± Hbl ± L in metamafic lithologies (mineral abbreviations after Kretz 1983). Peak *P*–*T* estimates (i.e., *P* at $T_{\max}$) for Ottawan metamorphism along the length of the belt cluster around 800–1100 MPa and 750–850 °C and are dated from ~1090–1050 Ma. Details of assemblages, estimated *P*–*T* conditions, and the time of metamorphism are given in Rivers (2008). Quartzofeldspathic and metapelitic lithologies are typically pervasively migmatitic, with penetrative L–S fabrics defined in part by strongly deformed melt pods (e.g., Culshaw et al. 1997). These fabrics provide evidence for very high strains (e.g., Schwerdtner et al. 1977; Davidson 1983) and point to the low viscosity of the mid crust during the Ottawan orogenic phase. In the two major areas of the Grenville Province where eclogite-facies rocks are preserved (aH*P* Belt; Fig. 2), the high-*P* rocks are in thrust contact with Ottawan mid-crustal rocks and variably overprinted by granulite- to upper amphibolite-facies assemblages, indicating they were exhumed from the lower orogenic crust and, thereafter, made up part of the Ottawan mid crust.

Ottawan upper crust — aL*P* Belt

The aL*P* Belt is a minor component of the exposed Ottawan crust, consisting of three discrete segments at the scale of Fig. 2, all of which are composed of medium- to high-grade, schistose to locally gneissic Ottawan metamorphic rocks, in which the peak pressures were significantly lower than in the aM*P* Belt (≤600 MPa). The boundaries of two of the three segments remain poorly mapped so the overall extent of the aL*P* Belt remains undefined, and a fourth segment along the northwestern margin of the orogenic lid in the southwest Grenville Province is too small to show on Fig. 2. On the basis of available data, the aL*P* Belt is characterized by peak pressures in the range 350–600 MPa and peak temperatures from ~500–750 °C. Contacts between the aM*P* and aL*P* belts are tectonic and exhibit normal-sense displacement where observed (details in Rivers 2008), whereas in the southwest Grenville Province, one segment of the aL*P* Belt is surrounded by the Ottawan Orogenic Lid (Fig. 2).

Uppermost Ottawan crust — Ottawan Orogenic Lid

As explained previously, Rivers (2008, 2009) identified an orogenic lid of Ottawan age (herein Ottawan Orogenic Lid (OOL)) to account for the occurrence of metamorphic rocks in the interior Grenville Province that lack evidence for high-grade Ottawan deformation and metamorphism. Figure 2 shows that the OOL comprises two large regions in the northeast and southwest Grenville Province, each covering a surface area of >50 000 km^2. In both regions, the lid is composed of several discrete domains. A third smaller region of the OOL is probable, comprising greenschist-facies rocks in the vicinity of Wakeham Bay in eastern Quebec within the aL*P* segment noted previously, but at present, there are insufficient data to define the boundaries with precision. The large region of the OOL in the northeast Grenville Province is principally underlain by high-grade gneisses with Paleoproterozoic metamorphic ages (~1650 Ma, i.e., Labradorian), whereas that in the southwest is composed of rocks with late Mesoproterozoic metamorphic ages reflecting accretion during the Elzevirian and Shawinigan orogenies (~1245–1220 and ~1190–1140 Ma, respectively). As documented later, although surrounded by Ottawan upper amphibolite- to granulite-facies gneisses, the rocks composing the lid apparently escaped penetrative Ottawan effects. In both areas, the lithologic units forming the OOL can be linked to those in the adjacent aM*P* Belt, implying it is not exotic. As with the aL*P* Belt, contacts between the OOL and surrounding belts exhibit normal-sense displacement, suggesting it was juxtaposed against higher grade rocks after the peak of Ottawan metamorphism.

Figure 2 shows that the maximum extent of the OOL is unconstrained because of the limits of exposure. Possible continuations beyond the northeast and southwest ends of the Grenville Province are shown based on deep-crustal seismic reflection data. They suggest that the lid is extensive and comprises a significant proportion of the preserved Grenvillian crust.

Crustal structure of the Grenville Orogen

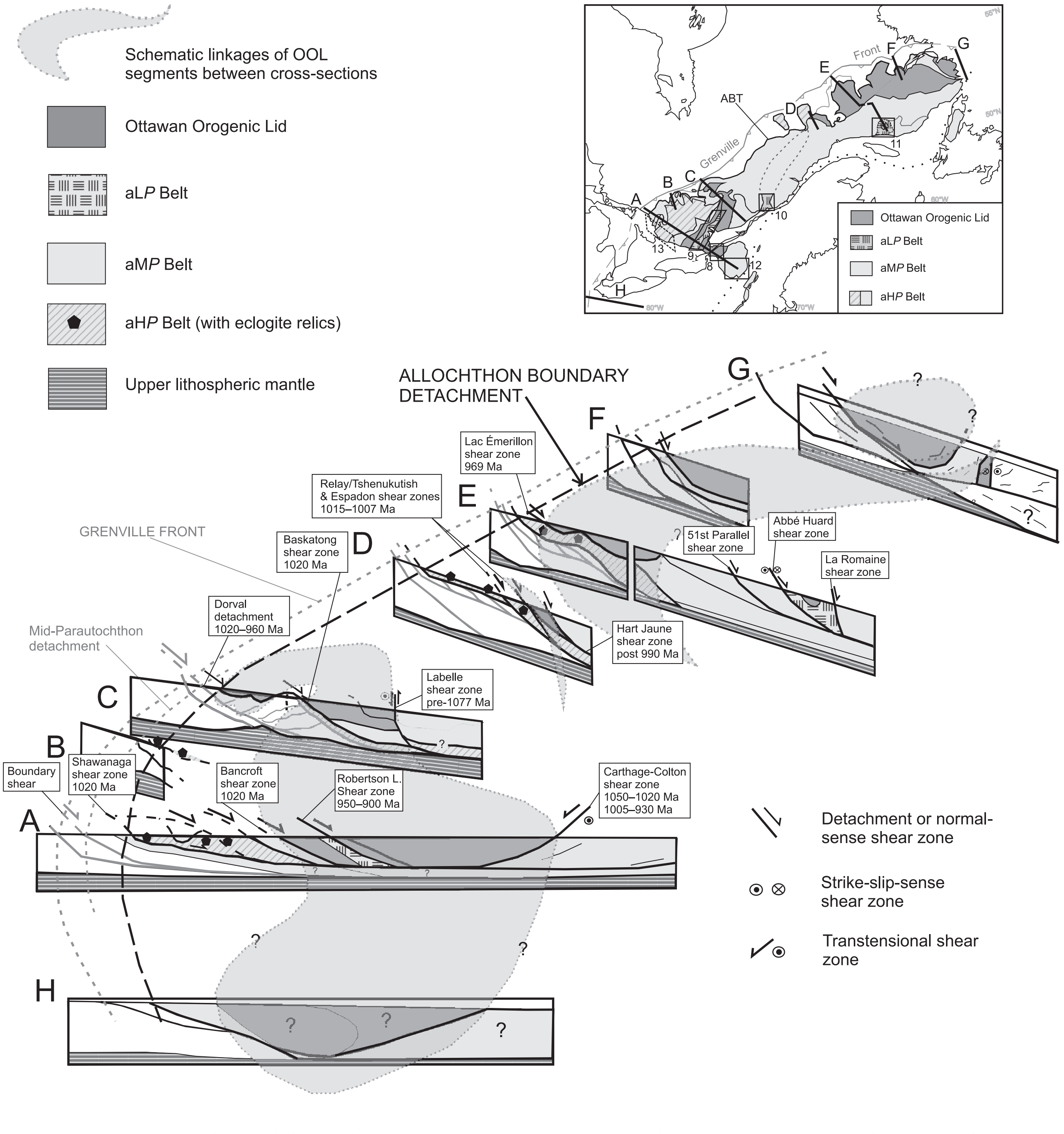
The results of individual deep-seismic transects of the Grenville Orogen undertaken under the auspices of Lithoprobe and the Consortium for Continental Reflection Profiling (COCORP) were reported by Green et al. (1988), Pratt et al. (1989), Kellett et al. (1994), Gower et al. (1997), Hynes et al. (2000), Martignole et al. (2000), and White et al. (2000), and were subsequently assembled into composite figures to show along-strike variations in crustal architecture by Ludden and Hynes (2000), Rivers et al. (2002), Rivers (2008, 2009), and Hynes and Rivers (2010). Early seismic interpretations were principally concerned with the crustal-scale architecture and did not incorporate the signatures of different orogenic levels (i.e., aH*P*, aM*P*, and aL*P* belts) nor did they recognize the OOL. The focus here is on integrating the different crustal levels with the effects of the major, post-peak, normal-sense shear zones into a coherent picture of the collapsed Grenville Orogen. Culshaw et al. (1994, 1997) and Carr et al. (2000) described field evidence for the reworking of several thrust-sense shear zones by normal-sense structures in the Georgian Bay area, and similar observations have been made elsewhere in the Grenville Province (e.g., van der Pluijm et al. 1994; Johnson et al. 2004). The straight gneiss fabrics making up these shear zones range from several hundred metres to over a kilometre wide and are (sub)parallel to the dominant foliation in adjacent rocks rendering offsets cryptic. Kinematic indicators in the shear zones commonly indicate both thrust- and normal-sense displacements, the latter generally in discrete segments of the shear zone. In an important study of the Allochthon Boundary Thrust in the western Grenville Province, Ketchum et al. (1998) described structural, metamorphic, and geochronologic evidence for its reworking as a normal-sense structure in late Ottawan time (ca. 1020 Ma). Similar evidence has subsequently been reported from elsewhere along its length, and Rivers et al. (2002) suggested the name Allochthon Boundary Detachment to describe the shear zone at this time.

The distribution of late- to post-Ottawan normal-sense structures on crustal-scale sections of the Grenville Orogen is shown in Fig. 3 (after Rivers 2008). Although the shear zones can commonly be identified on seismic sections, their normal sense of displacement is determined from surface data. Also shown on this figure are the inferred extent of the OOL and aL*P* Belt, which were not distinguished in the original seismic interpretations. As discussed earlier in the text, many normal-sense shear zones follow older contractional structures and most are moderately southeast dipping and penetrate to mid-crustal depths. However, important NW-dipping examples were imaged in the Adirondack region, in the buried southwest extension of the Grenville Orogen under Ohio, and in the Labrador offshore. Taken together, these oppositely dipping pairs of detachment structures define the limits of the OOL and indicate that it has the geometric form of several crustal-scale graben. Moreover, the architecture of the intervening horsts, which are composed of the mid-crustal aM*P* Belt, resembles a series of crustal-scale metamorphic core complexes (Cosca et al. 1995; Selleck et al. 2005; Bickford et al. 2008; Rivers 2008; McLelland et al. 2009, 2010). It is a conclusion of this paper that juxtaposition of the exhumed mid crust with the uppermost crust constitutes a signal of orogenic collapse. Additional support for this contention comes from the upper-crustal aL*P* Belt, which, when considered with the OOL, implies that collapse resulted in the juxtaposition of a range of upper-crustal levels with mid-crustal rocks. Elaboration of this general model is the subject of the remaining sections of this paper.

P–T signature of orogenic collapse

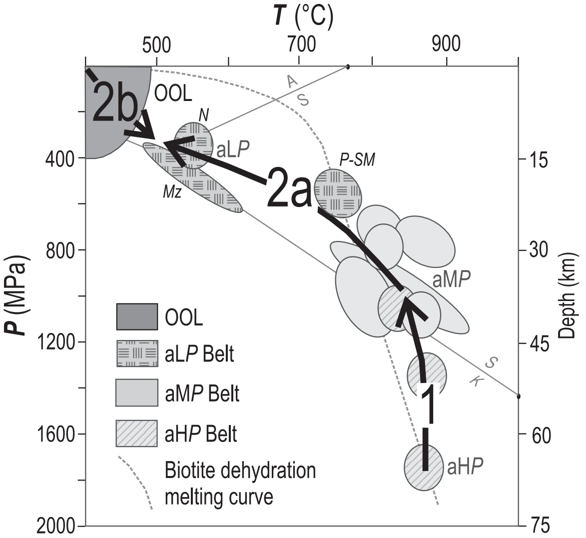
Details of Ottawan mineral assemblages and references to estimated *P–T* conditions in the aH*P* Belt, aM*P* Belt, aL*P*

Fig. 3. Crustal-scale cross-sections of the Grenville Orogen showing normal-sense structures and their relationship to the Ottawan Orogenic Lid (OOL; based on Lithoprobe data; modified after Rivers 2008). Transparent overlays show linkage of OOL between cross-sections. Ages in boxes refer to the times of major normal- and (or) strike-slip-sense displacements on shear zones. Inset map of Grenville Province shows locations of cross-sections and Figs. 8–13.



Belt, and OOL, summarized briefly earlier in the text, are given in Rivers (2008). Ottawan metamorphic conditions in the OOL have not been determined directly, but a maximum depth of ≤15 km was estimated from *T* constraints imposed by ^{40}Ar/^{39}Ar data and the estimated geothermal gradient. Figure 4 is a compilation of peak Ottawan *P–T* data from across the Grenville Province (adapted from Rivers 2008). The large arrows superimposed on the data indicate the inferred *P–T* trajectories of crustal segments after peak Ottawan metamorphism. Exhumation of the aH*P* Belt into the mid crust was approximately isothermal (arrow labelled 1 in Fig. 4), suggesting it accompanied thrusting and crustal thickening when the temperature was increasing. This is supported by geochronological evidence that the peak *P–T* conditions, or their

Fig. 4. *P*–*T* diagram showing estimated peak metamorphic conditions for the aH*P*, aM*P*, and aL*P* belts, and the OOL (modified after Rivers 2008). Arrow 1 shows early, thrust-sense, isothermal uplift of aH*P* Belt to mid crust. Arrows 2a and 2b show inferred path of exhumation and cooling of mid crust and its juxtaposition with the OOL resulting from crustal-scale, normal-sense displacements during orogenic collapse. Location of biotite dehydration-melting curve from Teyssier and Whitney (2002). Mazinaw (*Mz*), Natashquan (*N*), and Portneuf – St-Maurice (*P-SM*) are aL*P* domains discussed in text. Depths (in km) were calculated assuming an average crustal density of 2800 kg·m^{-3} and are approximate.



subsequent mid-crustal overprint, are the earliest recorded Ottawan event in the H*P* rocks (Rivers 2008). On the other hand, exhumation of the mid crust and its juxtaposition with the upper crust involved cooling from ~800 to ≤500 °C and is interpreted to be a result of orogenic collapse that postdated peak Ottawan metamorphism (arrows labelled 2a and 2b in Fig. 4). Since the Ottawan temperature of the OOL constrains its depth to ≤15 km, the data suggest the mid crust was exhumed from ~30 to ≤15 km depth by this process.

Thermochron signature of the Ottawan Orogenic Lid

The Ottawan Orogenic Lid was initially identified by its thermochron signature, specifically the "old" ^{40}Ar/^{39}Ar hornblende apparent ages of ≥1100 Ma that predate the Ottawan metamorphism (Rivers 2008). Since the closure temperature (T_C) for argon loss from the hornblende lattice is ~480–500 °C (McDougall and Harrison 1999), the old apparent ages were interpreted to imply that the rocks forming the lid were not heated above the hornblende T_C during the Ottawan orogenic phase. In contrast, as discussed previously, the surrounding mid-crustal rocks attained peak Ottawan metamorphic temperatures of ≥750–850 °C, and they typically yield ^{40}Ar/^{39}Ar hornblende cooling ages between 980 and 900 Ma (e.g., Anderson 1988). Since the units making up the lid exhibit lithologic linkages with the surrounding mid crust, they are interpreted as part of the cool (≤500 °C) orogenic upper crust that was juxtaposed with the hot (≥750–850 °C) mid crust after the peak of Ottawan metamorphism.

Although the focus of the discussion here is on ^{40}Ar/^{39}Ar hornblende data, the maximum temperature of the OOL in Ottawan times is also constrained by U–Pb data for titanite and rutile, for which a T_C of ~650 °C was estimated (Cherniak 1993, 2000). The higher T_C of these minerals provides a less rigorous constraint than the Ar results, but the data still imply the maximum temperature in the OOL was ≥100 °C, less than that of the surrounding mid-crustal rocks. After taking the inferred geothermal gradient into account, this implies derivation from a significantly shallower crustal depth.

Compilation of ^{40}Ar/^{39}Ar hornblende data from the SW Grenville Province

Published ^{40}Ar/^{39}Ar analyses of hornblende and other minerals in the hinterland of the Grenville Province have yielded important contributions to understanding post-orogenic cooling, including (*i*) that it was unusually slow, and (*ii*) that the distribution of ^{40}Ar/^{39}Ar hornblende apparent ages in the southwest Grenville Province is a signal of orogenic collapse (e.g., Cosca et al. 1991, 1995; van der Pluijm et al. 1994; Streepey et al. 2000, 2001, 2004). However, the data have not been re-examined in light of the orogenic crustal level and orogenic lid concepts. Since the most complete data sets are in the southwest Grenville Province, this area was selected for further interpretation in this study. Figure 5A is a map showing the distribution of almost 100 ^{40}Ar/^{39}Ar hornblende apparent ages from the literature — only plateau and a few total-gas ages are included; analyses exhibiting excess argon are omitted. Colder colours indicate apparent ages predating or coeval with the Ottawan orogenic phase, whereas warmer colours indicate apparent ages postdating it. Authors of the original data are listed in the figure caption. In Fig. 5C, the data have been projected orthogonally on to two NW–SE transects, X–X′, Y–Y′. Several authors (e.g., Cosca et al. 1991, 1995; van der Pluijm et al. 1994; Streepey et al. 2000, 2004) have pointed out that significant changes in apparent age, as much as ~200 Ma locally, take place at or close to domain and terrane boundaries, implying that they were tectonically active during cooling. This is especially evident in transect Y–Y′, which exhibits abrupt discontinuities in apparent age across the boundaries of the Frontenac–Adirondack Lowlands and Elzevir terranes.

To facilitate the interpretation of Fig. 5C, the data are grouped into four apparent-age bands labelled *AA-1* to *AA-4*; attention is initially focussed on the rocks in the hanging wall of the Allochthon Boundary Thrust (*ABT*, Fig. 5B). Interpretation of bands *AA-1* and *AA-4* is relatively straightforward. With respect to band *AA-1*, which includes apparent ages ≥1100 Ma, the data are restricted to two locations in central Elzevir and Frontenac–Adirondack Lowlands terranes in transect Y–Y′ (delimited by dotted lines in Fig. 5C). In the context of the Grenville Province, these are anomalously old apparent ages (the oldest of which are ≥1200 Ma), and they imply the rocks in question were not heated above ~500 °C during the Ottawan phase, one of the defining characteristics of the OOL. The apparent ages are, therefore, interpreted as cooling ages following the Elzevirian or Shawinigan metamorphisms, or in one case, as the time of cooling in the thermal aureole of a post-Elzevirian intrusion. The Ar data from the Elzevir terrane are supported by U–Pb titanite ages of ~1070–1050 Ma that are similarly interpreted as cooling ages (Davis and Bartlett 1988; Mezger et al. 1993).

At the other end of the spectrum, samples in band *AA-4*,

Fig. 5. (A) Map showing distribution of ^{40}Ar/^{39}Ar hornblende apparent ages in the southwest Grenville Province (only plateau ages (circles) and a few total-gas ages (squares) are plotted; analyses exhibiting excess argon are omitted). Symbols are colour-coded by apparent age into bins of 20 Ma duration (from 920–1020 Ma) and 30 Ma duration (from 1030 to ≥1090 Ma). Cold colours indicate "old" apparent ages that predate or are coeval with the Ottawan orogenic phase, warmer colours indicate younger apparent ages that postdate the Ottawan orogenic phase. The 2σ uncertainties of most plateau ages are ≤5 Ma (except ≤10 Ma for some older data), hence, smaller than bin sizes. Symbol sizes in (C) are schematic and larger than the measured uncertainties. X–X′ and Y–Y′ are locations of NW–SE-trending transects in western Quebec and SW Ontario. Data sources: Berger and York 1981; Lopez-Martinez and York, 1983; Onstott and Peacock, 1987; Cosca et al. 1991, 1992, 1995; Haggart et al. 1993; Reynolds et al. 1995; Busch et al. 1996*a*, 1996*b*, 1997; Cureton et al. 1997; Martignole and Reynolds 1997; Streepey et al. 2000, 2001, 2004; Dahl et al. 2004. (B) Sketch showing boundaries of aM*P* Belt, aL*P* Belt, and the Ottawan Orogenic Lid (OOL) along the two transects. (C) ^{40}Ar/^{39}Ar hornblende apparent ages projected onto X–X′ and Y–Y′ (modified after Martignole and Reynolds 1997; Carr et al. 2000; Dahl et al. 2004). Clusters of "old" apparent ages in Frontenac–Adirondack Lowlands (F-AL) terrane, Elzevir terrane, and Mont Laurier terrane are bounded by dotted bell-shaped curves. Durations of Grenvillian and pre-Grenvillian metamorphisms in central box are after Rivers (2008). *AA-1* to *AA-4* are apparent age bands referred to in the text. Domains and terranes: AH, Adirondack Highlands terrane; AL, Adirondack Lowlands; A-LD, Algonquin – Lac Dumoine; B, Belmont domain; BR, Britt domain; BT, Bancroft terrane; C, Cabonga domain; CAB, Composite Arc Belt; F, Frontenac domain; G, Grimsthorpe domain; GFTZ, Grenville Front Tectonic Zone; HC, Harvey–Cardiff Arch; M, Morin terrane; ML, Mont-Laurier domain; Mz, Mazinaw domain; PS, Parry Sound domain; RD, Réservoir Dozois; SL, Sharbot Lake; T, Tomiko; X, X terrane. Shear zones: *ABT*, Allochthon Boundary Thrust; *BSZ*, Bancroft; *BLSZ*, Black Lake; *BySZ*, Boundary; *CABSZ*, Composite Arc Belt; *CCSZ*, Carthage–Colton; *CSZ*, Cayamant; *HSZ*, Heaney; *LSZ*, Labelle; *MSZ*, Maberly; *MoSZ*, Mooroton; *NCSZ*, Nominingue–Chénéville; *PRSZ*, Perth Road; *PSSZ*, Parry Sound; *RSZ*, Renzy; *RLSZ*, Robertson Lake; *SSZ*, Shawanaga. GF, Grenville Front.

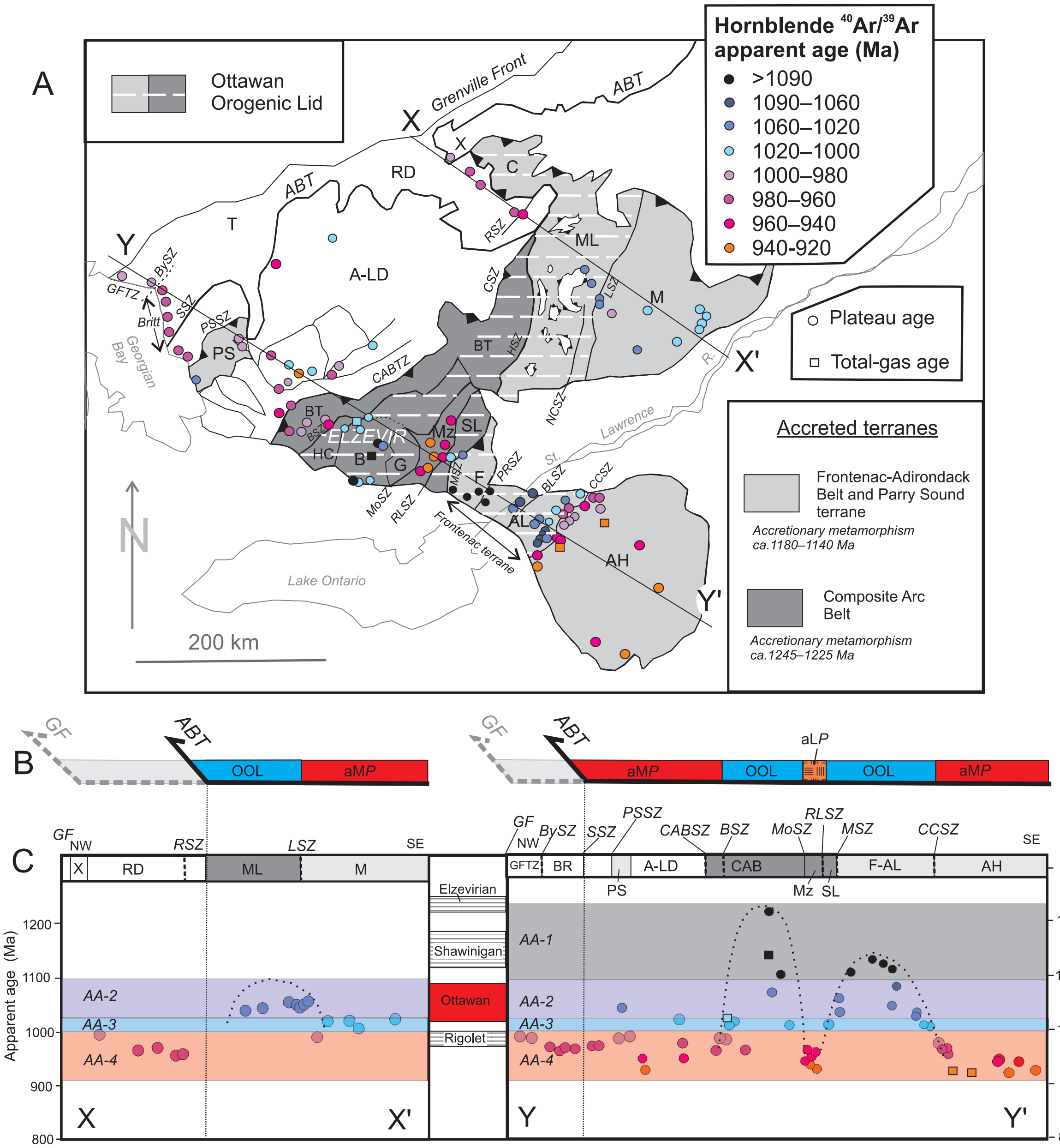
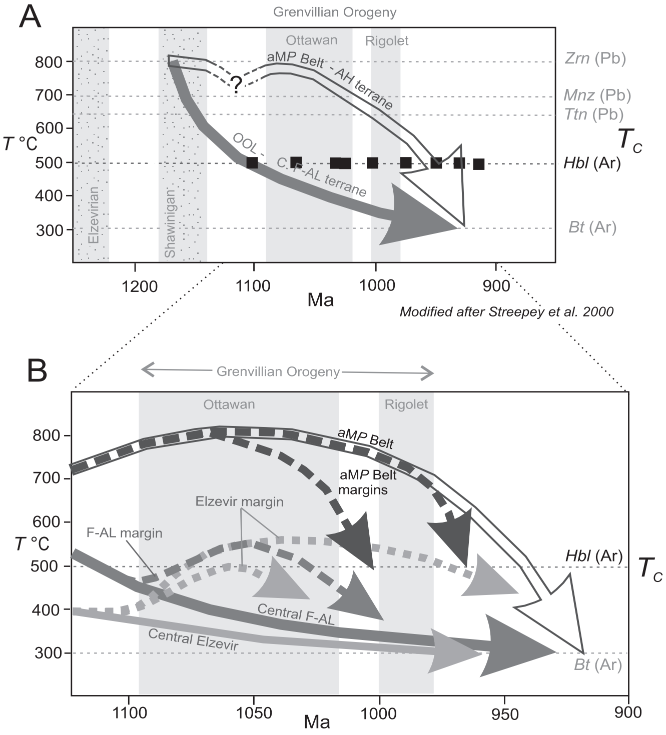


Fig. 6. Representative *T–t* paths for aM*P* Belt and Ottawan Orogenic Lid (OOL). (A) Arrows represent measured *T–t* paths for Adirondack Highlands (AH) terrane and central Frontenac–Adirondack Lowlands (C. F-AL) terrane using data from U–Pb zircon (*Zrn* (Pb)), monazite (*Mnz* (Pb)), and titanite (*Ttn* (Pb)), and ^{40}Ar/^{39}Ar data from hornblende (*Hbl* (Ar)) and biotite (*Bt* (Ar)) — modified from Streepey et al. (2000). Black boxes indicate the range of ^{40}Ar/^{39}Ar *Hbl* apparent ages determined for the Composite Arc Belt by Cosca et al. (1991). (B) Enlargement of part of figure shown in (A), showing explanation for the range of apparent hornblende ages and schematically illustrating cooling of parts of the aM*P* Belt and heating of the margins of the OOL and their subsequent common *T–t* path after ~950 Ma, when they come into contact during orogenic collapse (dashed curves).



Modified after Streepey et al. 2000

which principally come from the aM*P* Belt and occur in both transects, yield apparent ages of ∼1000–900 Ma. Since the aM*P* Belt experienced high-grade Ottawan metamorphism from ∼1090–1020 Ma, the apparent ages are interpreted as the time of cooling through the T_C for hornblende after this event. Possible explanations for their spread over ∼100 Ma are discussed later.

The samples in band *AA-3* yield apparent ages of 1100–1020 Ma, i.e., coeval with the Ottawan orogenic phase outside the lid, and most occur in the OOL. The preferred interpretation of these data is that they reflect the time of heating and subsequent cooling of the OOL after it came into contact with the hot mid crust during orogenic collapse, i.e., they are the thermal signature of synorogenic collapse in the upper crust. The lack of age plateaux for some samples in this group, and hence the data being reported as total-gas ages, indicates that the maximum temperature achieved was locally insufficient, or was maintained for insufficient time, to completely release all previously accumulated radiogenic Ar from the hornblende lattice. This suggests that it was close to the T_C and, therefore, probably did not greatly exceed 500 °C.

Finally, band *AA-2* is characterized by apparent ages of ∼1020 to 1000 Ma. The samples in this band come from both the OOL and the aM*P* Belt and include several from close to their mutual boundary. This distribution suggests convergence of cooling paths for the OOL and aM*P* Belt during the latter stages of the Ottawan orogenic phase. These interpretations are summarized schematically in *T–t* space in Fig. 6B.

With respect to Fig. 6A (modified from Streepey et al. 2000, 2004), the contrasting cooling paths for samples from the central Frontenac–Adirondack Lowlands terrane (OOL) and the structurally underlying Adirondack Highlands terrane (aM*P* Belt) are tightly constrained, being based on many data acquired using several geochronometers with a range of closure temperatures noted on the right side of the figure. They imply that, during the Ottawan orogenic phase, the hanging-wall rocks in the lid were ∼250 °C cooler than those in the underlying aM*P* Belt footwall. Data for the Composite Arc Belt are superimposed on the figure and exhibit a spread of apparent ages from ∼1100 to 940 Ma (data from Cosca et al. 1991, 1995; for location of the Composite Arc Belt, see Fig. 5A).

Figure 6B illustrates the interpretation that the lid was partially heated and the adjacent footwall locally cooled through the 500 °C isotherm as a result of their tectonic juxtaposition during orogenic collapse, the hornblende apparent ages preserved in bands *AA-2* and *AA-3* providing a record of when this occurred. Apparent ages of ∼1099–1020 Ma in band *AA-2*, principally restricted to the OOL, as noted, but also occurring in the allochthonous Parry Sound terrane (Fig. 5A), are interpreted to indicate heating slightly above the 500 °C isotherm during the Ottawan phase before subsequent cooling through this temperature later during the Ottawan phase. Note that in contrast to the Frontenac–Adirondack Lowlands and Elzevir terranes, no pre-Ottawan apparent ages are preserved in the Mont-Laurier terrane (transect X–X′, Fig. 5), implying it was heated above the 500 °C isotherm during the Ottawan phase. Finally, the apparent ages of ∼1020–1000 Ma in band *AA-3*, which occur in both the OOL and its mid-crustal footwall, are interpreted as a signature of their common cooling path following tectonic juxtaposition in late to post-Ottawan time.

In summary, integration of the orogenic lid concept with the hornblende ^{40}Ar/^{39}Ar data provides a framework in which to interpret the apparent ages and leads to an improved understanding of the heat transfer between hot footwall and cool hanging wall during orogenic collapse. Specifically, it yields the crucial insight that their tectonic juxtaposition resulted in modification of the *T–t* paths of both. Hence what formerly appeared as a data cloud, such as the results for the Composite Arc Belt shown in Fig. 6A, is seen to have a coherent explanation. Other examples of Ar data clouds include the ≥100 Ma spread of apparent ages in the Carthage–Colton and Robertson Lake shear zones, which may not be a result of prolonged displacement on the shear zones as previously interpreted, but rather reflect convergence of contrasting *T–t* paths in the footwalls and hanging walls. It is in this context, and noting that some of the *T–t* paths are inferred to have involved heating as well as cooling, the term *apparent age* is preferred over the more widely used “cooling age.”

Relationship between Ottawan Orogenic Lid and the L*P* Mazinaw terrane

Figure 5 shows that the OOL in transect Y–Y′ is divided into two segments by the Mazinaw terrane, part of the aL*P* Belt. Hornblende ^{40}Ar/^{39}Ar apparent ages in the Mazinaw terrane are ~940–960 Ma (band *AA-4*), indicating it followed a *T–t* path similar to the aM*P* Adirondack Highlands terrane, remaining above ~500 °C until that time. The occurrence of L*P* Ottawan metamorphism in the Mazinaw terrane, and its location between the Frontenac–Adirondack Lowlands and Elzevir terranes of the OOL (Fig. 4), defines a horst and graben architecture, in which the Mazinaw terrane forms a "hot" horst bounded on both sides by "cool" graben. Peak temperatures of ~550 to 650 °C across the Mazinaw metamorphic field gradient (Rivers 1976; Easton 1992) are interpreted as a frozen-in record of the thermal structure of the orogenic crust during collapse.

Unusually slow cooling in the Adirondack Highlands terrane?

Hornblende ^{40}Ar/^{39}Ar apparent ages in the Adirondack Highlands terrane are as young as ~920 Ma (Fig. 5), significantly younger than those measured elsewhere in the aM*P* Belt, which are mostly in the range 970–950 Ma (Anderson 1988). Given the robust evidence for Ottawan granulite-facies metamorphism in the Adirondack Highlands terrane from ~1090 to 1040 Ma, with a probable peak at ≥1060 Ma (e. g., Mezger et al. 1991; McLelland et al. 1996, 2001; Spear and Markussen 1997; Darling et al. 2004; Heumann et al. 2006; Bickford et al. 2008), and the fact that the terrane represents the most internal exposed part of the Grenvillian hinterland raises an important question — Are the young apparent ages a regional feature of the internal hinterland reflecting slow cooling (i.e., ~2.5 °C·Ma^{-1} over the temperature interval 850–500 °C)? This issue is germane to the discussion of orogenic collapse because, if the inference of slow cooling is taken at face value as a regional signal of slow exhumation, it would imply that this part of the aM*P* Belt remained hot and, therefore, assuming a reasonable geothermal gradient, at considerable depth for over 100 Ma after peak Ottawan metamorphism. Such an evolution would be incompatible with the rapid exhumation of the mid crust expected as a consequence of orogenic collapse. Here, we discuss some data on the age and character of hydrothermal alteration processes associated with the extensive ~1050 Ma Lyon Mountain leucogranite in the Adirondack Highlands terrane that suggest an alternative interpretation.

Hydrothermal alteration in the Lyon Mountain granite is of the iron ore – copper – gold (IO(CG)) type, in which iron ore (magnetite) mineralization was dominant (Foose and McLelland 1995; McLelland et al. 2002). This type of alteration, which is characterized by mineralogically diverse products, is inferred to result from the crustal-scale circulation of chloride-rich, low sulphur hydrothermal fluids in terranes undergoing extension, with the Cl^{-} ions facilitating transport of metals as chloride complexes (e.g., Hitzman et al. 1992; McLelland et al. 2002; Corriveau and Mumin 2010). Thus, the very existence of this type of mineralization suggests an extensional setting. Secondly, the temperature of the fluids that caused related alteration in quartz–sillimanite nodules has been estimated from partitioning of δ^{18}O between Qtz–Mt pairs to be ~565–675 °C (McLelland et al. 2002), which is above the T_C for Ar in hornblende. Thirdly, since the earliest petrographic studies (e.g., Postel 1952), the great width of the alteration zones, up to hundreds of metres wide, characterized by the breakdown of mafic silicate phases, such as pyroxene and amphibole, and the presence of magnetite mineralization, has been recognized (see also e.g., Foose and McLelland 1995; Valley et al. 2009). Fourthly, evidence was presented by McLelland et al. (2002) that the Lyon Mountain granite was intruded at a considerably shallower depth than that recorded by geobarometry of the surrounding country rocks, compatible with the inference of Selleck et al. (2005) that emplacement occurred during exhumation of the Adirondack Highlands terrane. Finally, dating of hydrothermal zircon from several of the magnetite deposits associated with the Lyon Mountain granite has yielded ages of ~1040 and ~1015–1000 Ma (Valley et al. 2009), indicating that hydrothermal alteration occurred some 20–60 million years after crystallization of the host granite at different localities. Coupled with petrographic evidence for different mineralogical styles of alteration (potassic, sodic, Fe-rich) in several deposits, noted by many workers, and the existence of a related swarm of undeformed, rutilated quartz–albite (Ab$_{98}$) dykes and veins in the northwest Adirondack Lowlands (Brown 1983), this implies a prolonged episode of fluid activity.

Collectively, these results suggest the possibility that the young hornblende ^{40}Ar/^{39}Ar apparent ages in the Adirondack Highlands terrane may be a result of resetting of Ottawan hornblende by hot fluids circulating along normal-sense shear zones during orogenic collapse, the proposed setting for the IO(CG) mineralization in several recent studies (e.g., McLelland et al. 2002; Selleck et al. 2005; Valley et al. 2009). This interpretation is also supported by the observation that several of the Ar analyses are total-gas ages rather than plateau ages (Fig. 5), implying incomplete resetting. Moreover, the zircon population from one of the samples analysed by Valley et al. (2009) yielded a spectrum of age determinations from ~1020 to 950 Ma, supporting prolonged hydrothermal fluid activity and multiple episodes of zircon growth and (or) Pb loss. If this interpretation is correct and applicable throughout the Adirondack Highlands terrane where IO(CG) deposits are widespread (McLelland et al. 2002; Gauthier and Chartrand 2005), the unusually young hornblende apparent ages may not be a signal of slow regional cooling (e.g., Onstott and Peacock 1987), but rather reflect an unusually high geothermal gradient driven by circulating hydrothermal fluids in an extensional setting. As such, they would be a manifestation of orogenic collapse rather than evidence against it.

^{40}Ar/^{39}Ar hornblende apparent ages in the Parautochthonous Belt

The distribution of ^{40}Ar/^{39}Ar hornblende apparent ages in the Parautochthonous Belt in the northwest of the Grenville Orogen is also shown in Fig. 5. Although structurally beneath the Allochthon Boundary Thrust, on which Ottawa collapse took place, the data for the Parautochthonous Belt provide constraints on the tectonic setting at the time collapse occurred in the hinterland. As discussed previously, the Parautochthonous Belt is situated in the hanging wall of the Grenville Front and is characterized by Rigolet metamorphism from ~1005 to 980 Ma. $^{40}Ar/^{39}Ar$ hornblende analyses determined from the Parautochthonous Belt in transect Y–Y′ indicate apparent ages of ~995–980 Ma in the Grenville Front Tectonic Zone, declining to ~980–960 Ma farther southeast. These data lead to several conclusions: (*i*) cooling from peak metamorphic Rigolet temperatures (of ~750 °C; Rivers 2008) through the 500 °C isotherm was rapid, occurring within ~20–40 million years after peak metamorphism; i.e., cooling rates were several times more rapid than in the orogenic hinterland; and (*ii*) peak Rigolet metamorphism in the Parautochthonous Belt was coeval with cooling in the hanging wall of the Allochthon Boundary Thrust. This latter observation is significant as it provides an important constraint on the driving force for collapse in the orogenic interior, specifically that (*iii*) collapse took place in an overall compressional orogen and, hence, as surmised from theoretical considerations, was principally driven by the excess potential energy of the orogenic plateau.

Structural signature of orogenic collapse

In this section, we examine the structural signatures of three levels of Ottawan orogenic crust, the OOL, aL*P* Belt, and aM*P* Belt. Seismic imaging of the normal-sense shear zones linking these crustal levels was discussed previously (Fig. 3). The approach here is to review the structural evolution of selected well-studied areas using maps, cross-sections, and structural analysis, focussing on geological evidence related to orogenic collapse. After identification of characteristic features of each crustal level, they are integrated into a crustal-scale model for orogenic collapse of the Grenville Orogen.

Structures in Ottawan Orogenic Lid

It was shown in the previous section that the OOL was not heated above ~500 °C during the Ottawan orogenic phase. Since this temperature approximately coincides with the boundary between the greenschist and amphibolite facies, and most rocks in the OOL exhibit amphibolite-facies assemblages, this implies the ductile amphibolite-facies structures that characterize much of the OOL are pre-Ottawan — although the possibility exists that they may have undergone some Ottawan reworking below the T_C of Ar in hornblende. Using geon-11 (~1170 Ma) and geon-10 (~1090–1050 Ma) intrusions in the OOL and aM*P* Belt as regional strain markers permits a comparison of deformation between the two crustal levels. This is illustrated in cartoon form in Fig. 7A. In the OOL, undeformed geon-11 and -10 intrusions that crosscut the regional foliation, such as the Kingston dykes (ca. 1160; Davidson 1998) and the subcircular composite bodies of the ~1090–1060 Ma potassic Kensington–Skootamatta suite (Fig. 7B; Corriveau 1990; Easton 1992; Corriveau and Gorton 1993) in the southwest Grenville Province, are characterized as post-tectonic in the field. This implies that the regional foliation in the OOL is pre-Ottawan, as argued previously on the basis of the Ar data. Moreover, some ~1070 Ma granitoid bodies in the Frontenac domain exhibit granophyric groundmass, miarolitic cavities, and narrow contact aureoles (Davidson 2001), indicating emplacement at very high levels into cool crust. In contrast, mid-crustal intrusions of this age in the adjacent aM*P* Belt are strongly recrystallized and exhibit foliated, augen, and gneissic fabrics in Ottawan shear zones, e.g., ~1050 Ma Lyon Mountain granite in the Adirondack Highlands terrane (Fig. 7C; McLelland et al. 2001; Selleck et al. 2005; Valley et al. 2009), several ~1080–1030 Ma granitoid bodies in the aM*P* Belt in central Quebec (Hébert et al. 2009). Collectively, these observations indicate not only the coeval emplacement of granitoid rocks into the mid and upper orogenic crust, but also the contrasting tectonic regimes of the two crustal levels — specifically ductile strain of the hot mid crust beneath a cool, rheologically strong, upper-crustal carapace that remained undeformed.

A map and cross-section of part of the Adirondack Lowlands domain and adjacent Highlands terrane is shown as an example of the structural style of the OOL and its relationship with the adjacent aM*P* Belt (Fig. 8; after Wiener et al. 1984). In this domain, there are no geon-10 intrusions to act as strain markers, so other data are discussed. The Adirondack Lowlands domain, part of the OOL, is principally underlain by a geon-12 supracrustal sequence comprising metapelite, metapsammite, marble, and calc-silicate rocks, and Qtz–Bt–Pl gneiss of possible metavolcanic origin (metadacite? Carl 1988). The adjacent Adirondack Highlands terrane, part of the aM*P* Belt, is largely underlain by metaplutonic units associated with a geon-11 anorthosite–mangerite–charnockite–granite (AMCG) complex. Both the Adirondack Lowlands and Highlands are polydeformed, but the folds are of different age. Of the five fold generations recognized in the Adirondack Lowlands domain, only the youngest (F5) is Ottawan. It consists of NW-trending cross-folds that give rise to the F3 and F5 dome-and-basin interference pattern in the Lowlands that are continuous into the Highlands (Fig. 8; Wiener 1983; Wiener et al. 1984; Baird and MacDonald 2004; McLelland et al. 2010). According to Selleck et al. (2005), the late cross-folds affected the ca. 1050 Ma Lyon Mountain leucogranite (Fig. 7C; east of the area shown in Fig. 8), and hence, they are inferred to be of mid to late Ottawan age.

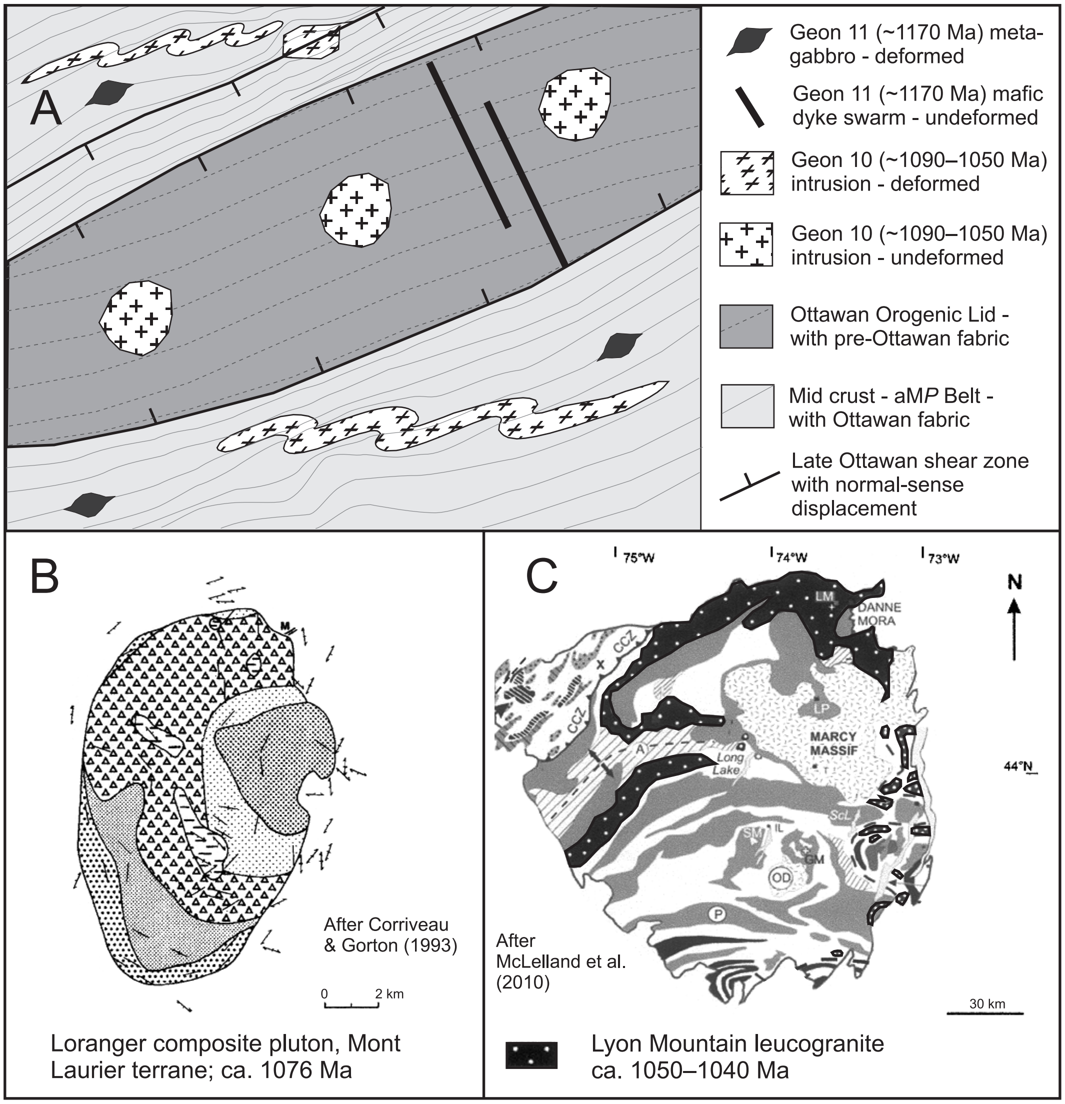
On the other hand, the NW-dipping Carthage–Colton shear zone at the southern margin of the Adirondack Lowlands domain has yielded evidence for protracted Ottawan deformation involving oblique-reverse displacement from ~1090 to 1030 Ma and oblique-normal displacement from ~1040 to 1020 Ma (Streepey et al. 2001; Johnson et al. 2004; Selleck et al. 2005; Baird et al. 2008). Ottawan granulite-facies metamorphism was pervasive in the Adirondack Highlands terrane in the footwall of the shear zone, and heating to ≥500 °C up to a few kilometres into the Lowlands domain in its immediate hanging wall is indicated by “young” $^{40}Ar/^{39}Ar$ hornblende apparent ages (~1020–1000 Ma; Figs. 5, 6B).

Structures in allochthonous Low-Pressure Belt

Although united by their low-pressure Ottawan metamorphic signature, structures in the three segments of the aL*P* Belt shown on Fig. 2 differ in ways that provide critical insight into the evolution of the orogenic crust. As a result, each is discussed briefly in the following paragraphs.

Mazinaw terrane: Fig. 9 is a sketch map of the Mazinaw terrane (after Easton 1992), a structural horst that divides the OOL in the southwest Grenville Province into two parts (Fig. 2). Ottawan structure and metamorphism are recorded by the monocyclic Flinton Group that was deposited uncon-

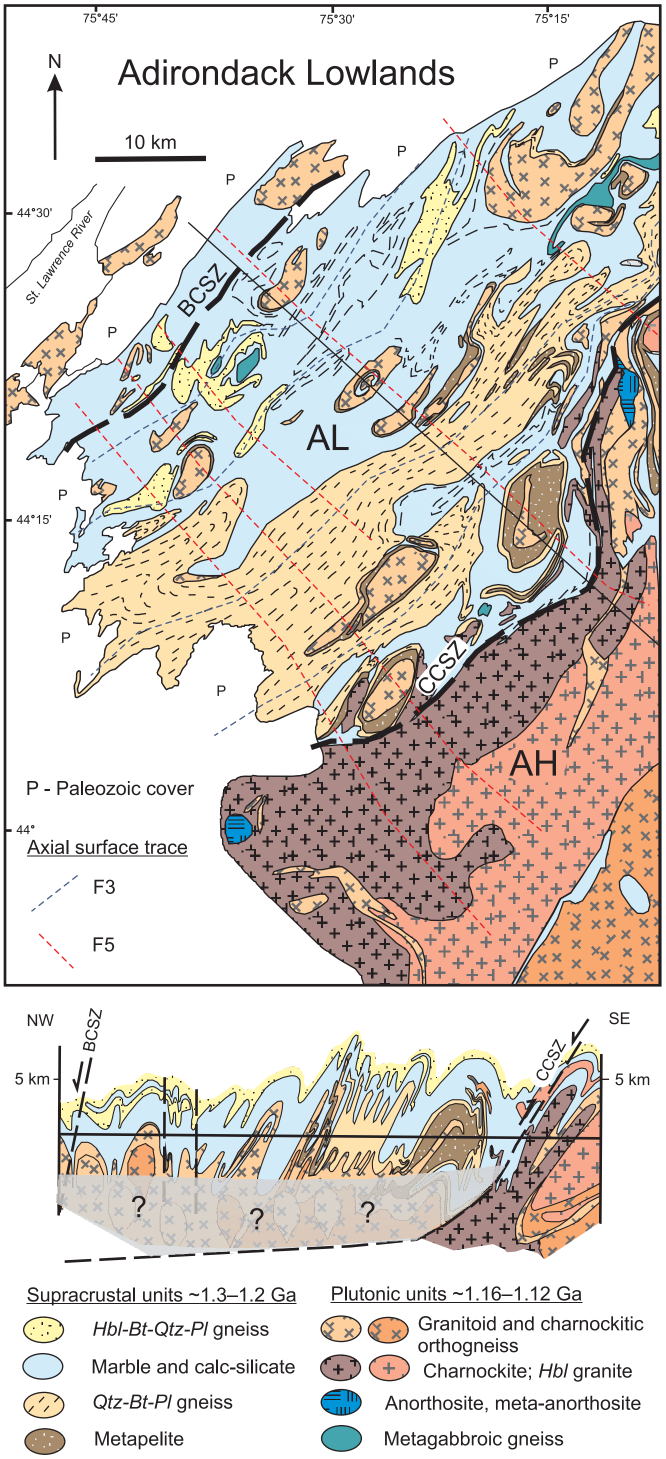
Fig. 7. (A) Cartoon illustrating the contrasting deformation states of geon-11 and geon-10 intrusions in the mid-crustal aM*P* Belt and the upper-crustal Ottawan Orogenic Lid. Note contrasting ages of fabrics in the two crustal levels. (B) Map of the undeformed ~1076 Ma Loranger composite pluton, part of Kensington–Skootamatta suite, emplaced into OOL (from Corriveau and Gorton 1993). (C) Map of Adirondack Highlands terrane, part of aM*P* Belt, emphasising folded bodies of the ~1050–1040 Ma Lyon Mountain leucogranite (from McLelland et al. 2010).



formably on igneous and sedimentary rocks metamorphosed during the ~1245–1225 Ma Elzevirian Orogeny. Deposition of the Flinton Group is, thus, traditionally bracketed by the end of Elzevirian and beginning of the Ottawan orogenies (i.e., ~1225–1090 Ma; Moore and Thompson 1980), although earlier deposition as an Elzevirian molasse unit cannot be excluded. The Flinton Group is preserved in a series of tight, NE-trending cuspate synforms separated by wider lobate antiforms underlain by its Elzevirian basement. Three episodes of deformation have been identified (Rivers 1976), of which at least the last two were Ottawan. The observation that individual units cannot be traced around map-scale F1 fold hinges suggests the existence of cryptic D1 shear zones that Schwerdtner et al. (2004) inferred were stretching thrusts. D1 structures were refolded by coaxial F2 folds with steep axial surfaces that developed under peak Ottawan metamorphic conditions and by more open post-peak F3 folds with gently dipping axial surfaces, resulting in two generations of hook-shaped fold interference patterns. From a regional perspective, the dominant F2 structures imply vertical thickening resulting from subhorizontal shortening, and the subordinate F3 structures, which are especially developed in weak schistose units, indicate this was followed by limited vertical flattening, the sequence implying that crustal loading owing to thickening locally exceeded the rock strength and led to minor collapse.

An Ottawan metamorphic field gradient, defined by Cld–St, St–And, St–Ky, Sil–Ms, and Sil–Kfs zones in Flinton Group metapelite is at high angles to the NE-trending structures (Fig. 9 inset). Peak *P*–*T* conditions ranged from ~350 MPa at 500 °C for the Cld–St zone to ~500 MPa at 650 °C for the Sil–Kfs zone (Fig. 4; Rivers 1976). Geochronological data indicate the timing of Ottawan metamorphism was different in the northwest and central parts of the Mazi-

Fig. 8. Map and schematic cross-section showing the geology of part of the Adirondack Lowlands domain (AL; part of the OOL) and adjacent Adirondack Highlands terrane (AH; part of the aM*P* Belt) separated by the Carthage–Colton shear zone (CCSZ) (modified from Wiener et al. 1984; see Fig. 3 for location). Grey transparent overlay in cross-section below ∼5 km depth indicates uncertainty concerning nature of basement. Note NW-trending F5 folds cross the CCSZ, implying they formed after juxtaposition of AL and AH. Cross-section shows pre-Ottawan structure in the Adirondack Lowlands domain. BCSZ, Black Creek shear zone.



naw terrane. In the former, it is constrained by ∼1090 Ma metamorphic zircon and monazite from pelitic schist and by a ∼1065 Ma crystallization age for a sheared pegmatite, whereas in the latter, metamorphic zircon and monazite yielded ages of ∼1035 and ∼1025 Ma (Corfu and Easton 1995). It is noted here that the ∼1090–1065 Ma age range overlaps with the time of thrusting in the adjacent aM*P* Belt (e.g., ∼1080 Ma in Bancroft terrane; McEachern and van Breemen 1993), whereas the ∼1035–1025 Ma ages coincide with normal-sense displacement and orogenic collapse. The peak pressure estimates indicate metamorphic burial depths ranged from ∼13 to 20 km along the Ottawan metamorphic field gradient (Fig. 4), the exposure of a range of depths at the erosion surface being attributed to tilting and differential exhumation of the Mazinaw horst within the larger OOL graben (Rivers 2008). The high angle between isograds and the Ottawan structural grain, the location of the low-grade rocks at the margin of the "Hastings metamorphic low," and the zircon and monazite metamorphic ages of ∼1035–1025 Ma are all consistent with prograde L*P* metamorphism in late Ottawan time by heating from below, rather than as a result of thermal relaxation in a thick imbricated crustal section.

The shear zones bounding the Mazinaw terrane are amphibolite-facies structures that dip steeply ESE (Fig. 9). Kinematic analysis of the Mooroton shear zone has revealed early reverse displacement overprinted by normal-sense microstructures, whereas the Robertson Lake shear zone only exhibits normal-sense kinematics. In the latter, brittle fabrics overprint mylonitic microstructures, indicating displacement continued to low temperature, and the shear zone separates greenschist-facies rocks in its hanging wall from amphibolite-facies rocks in the Mazinaw terrane footwall, implying it is the site of significant displacement. Time of displacement on the Mooroton shear zone is bracketed by ^{40}Ar/^{39}Ar hornblende apparent ages of ca. 1020 Ma in the Elzevir terrane and 930 Ma in the Mazinaw terrane (Cureton et al. 1997), whereas, in the Robertson Lake shear zone, it is bracketed by ^{40}Ar/^{39}Ar hornblende ages of ∼950–900 Ma (Busch and van der Pluijm 1996).

Portneuf–St-Maurice domain: The Mauricie region, of which the L*P* Portneuf–St-Maurice domain is a part, comprises four stacked terranes and domains (Fig. 10; Bernier and MacLean 1993; Corrigan and van Breemen 1997). At the bottom of the stack in the west the Mékinac–Taureau terrane is principally composed of ∼1.37 Ga orthogneiss, overlain by the Shawinigan domain composed of younger (≤1.3 Ga) ortho- and paragneiss, including the St-Boniface metasediments that are provisionally dated at ≤1.18 Ga. The central Portneuf–St-Maurice domain comprises remnants of the accreted ca. 1.45 Ga Montauban island arc intruded by ∼1.37 Ga orthogneiss, and the Parc des Laurentides domain in the east is dominated by a geon-10 AMCG suite. The ∼1.37 Ga orthogneisses are calc-alkaline and comprise part of the continental-margin arc that developed on southeast Laurentia during the Mesoproterozoic (Nadeau et al. 2006). Recent work suggests the Montauban island arc is a major feature of the central Grenville Province that continues ∼400 km farther towards the northeast (Fig. 2; Dunning and Indares 2010).

The Ottawan structural architecture of the Mauricie region is dominated by the Tawachiche shear zone (Fig. 10), an anasto-

Fig. 9. Sketch map of the aL*P* Mazinaw terrane (after Easton 1992; see Fig. 3 for location); cross-section X–X′ modified after Rivers (1976). BD, Belmont domain; GD, Grimsthorpe domain; SLD, Sharbot Lake domain. Inset figure shows simplified distribution of Ottawan isograds and metamorphic zones (A to D) determined from metapelitic assemblages in the Flinton Group.

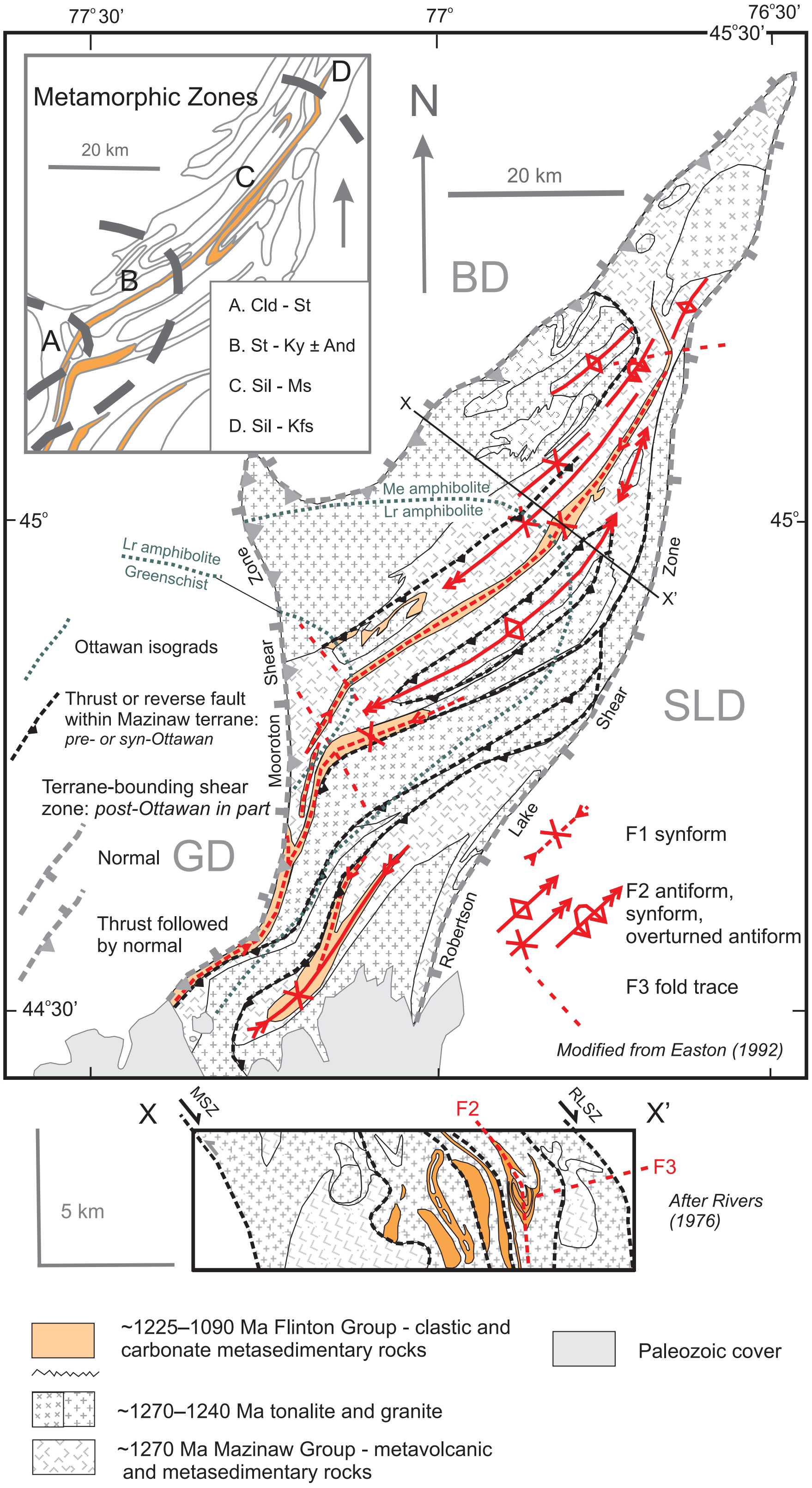
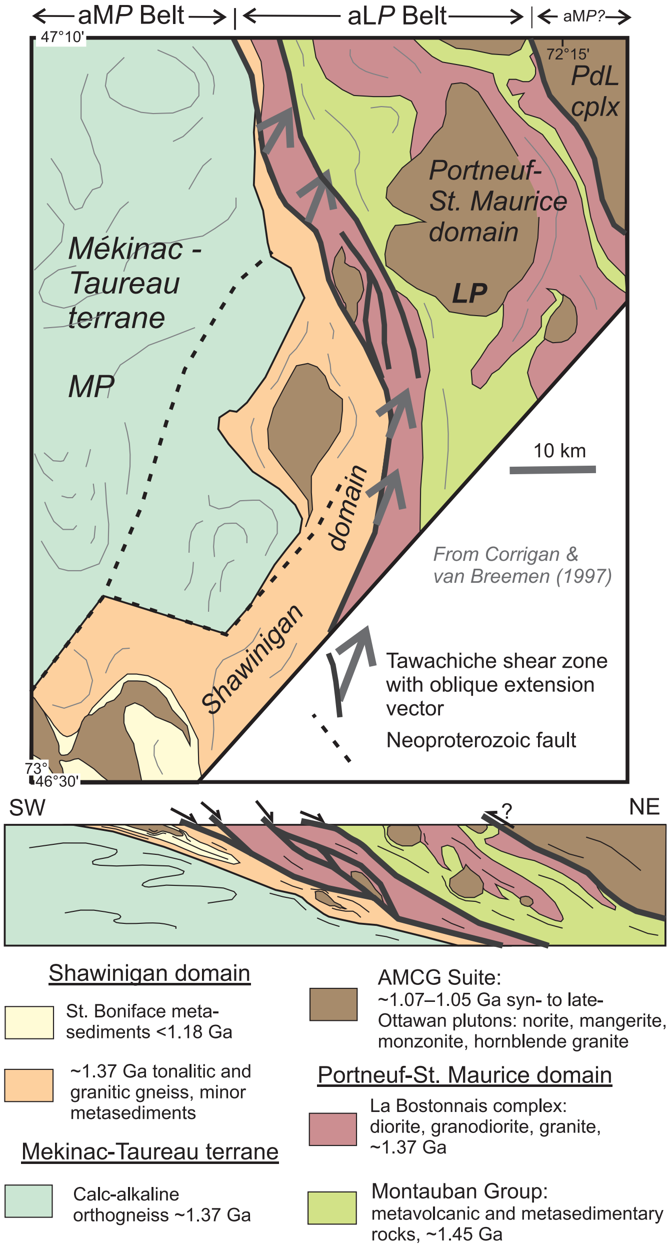


Fig. 10. Sketch map of the aL*P* Portneuf – St-Maurice domain in the hanging wall of the oblique-normal Tawachiche shear zone (after Corrigan and van Breemen 1997; see Fig. 3 for location). Schematic cross-section is after Rivers et al. (in press). The footwall of the shear zone is composed of the aM*P* Mékinak–Taureau terrane and overlying Shawinigan domain.

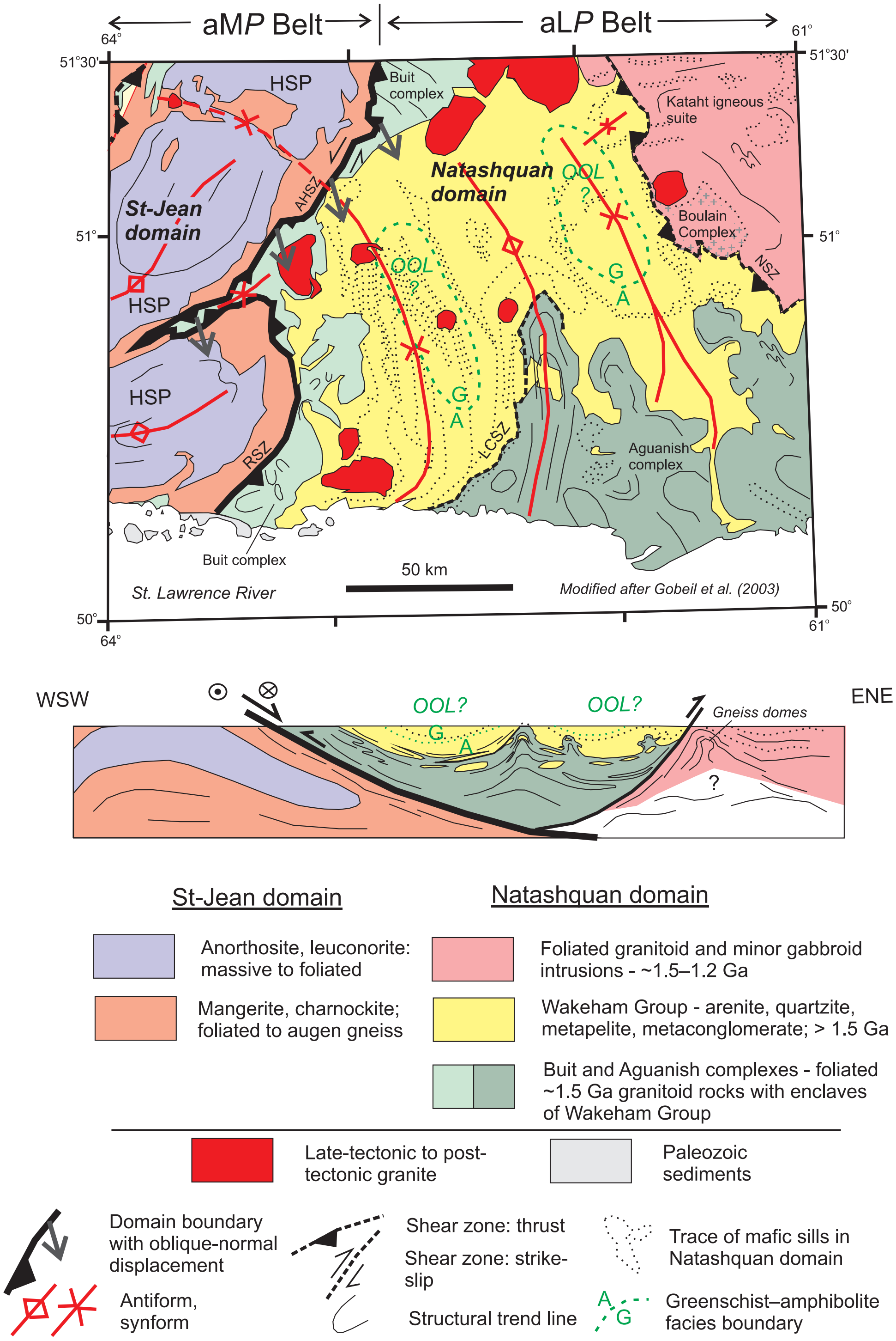


mosing NNW-trending structure a few kilometres wide that exhibits oblique-normal displacement and divides the region into footwall and hanging wall terranes. The domical footwall, composed of the mid-crustal Mékinac–Taureau terrane and Shawinigan domain, both part of the aM*P* Belt, exhibits recumbent structures, whereas the hanging wall, consisting of the Portneuf–St-Maurice domain, part of the aL*P* Belt, exhibits NE-dipping structures. High-grade metamorphism in the M*P* footwall of the shear zone took place in the early Ottawan (between ~1.12 and 1.09 Ga according to available age determinations) and is recorded by granulite-facies assemblages in the Mékinac–Taureau terrane that locally contain sapphirine (*P–T* conditions ~900 MPa at 820 °C; Herd et al. 1986), and by upper amphibolite-facies assemblages in the overlying Shawinigan domain (*P–T* conditions ~700–1000 MPa at 725–825 °C; Corrigan and van Breemen 1997). Metamorphism in the hanging wall of the Tawachiche shear zone is constrained by assemblages in the Montauban Group, which outcrop in two strands. The western strand exhibits amphibolite-facies assemblages and locally well-preserved primary structures, whereas the eastern strand is pervasively migmatitic and exhibits low-pressure granulite-facies assemblages (e.g., Opx–Ath–Crd–Bt–Pl–L; Crd–Spl–Grt–Sil–Qtz–L). Peak *P–T* conditions are estimated as 300–600 MPa at 550–625 °C for the western strand and 500–600 MPa at 750 °C for the eastern strand, with the metamorphism being dated at ~1056 Ma based on the age of a syntectonic pegmatite (Corrigan and van Breemen 1997 and references therein). The estimated difference in Ottawan metamorphic pressures between the footwall and hanging wall of ca. 200–300 MPa indicates the Tawachiche shear zone accommodated ~6–10 km of vertical displacement. The geochronological data imply peak metamorphism in the hanging wall took place some 40–50 million years after that in the footwall. Voluminous geon-10 AMCG magmatism in the vicinity of the Tawachiche shear zone, including within the Parc des Laurentides domain, was approximately coeval with oblique-normal displacement, both predating and postdating it locally.

Natashquan domain, Côte Nord: The Côte Nord region of the Grenville Province comprises several stacked domains, of which the two shown in Fig. 11 are separated by the Abbé Huard shear zone, a thrust structure that was reworked by oblique-normal displacement (Martignole et al. 1994; Gobeil et al. 2003). The structurally lower of the two domains, the mid-crustal St-Jean domain, is dominated by several thick thrust slices of the voluminous, geon-11, Havre St-Pierre AMCG complex, in which anorthosite displays Opx–Cpx–Grt coronas. This, together with Qtz–Kfs–Grt–Sil–Bt ± Gr and Opx–Cpx–Pl–Grt assemblages in metapelite and metabasite, respectively, elsewhere in the domain imply the St-Jean domain underwent granulite-facies metamorphism (Gobeil et al. 2003). Metamorphic zircon and monazite from these rocks has yielded ages between ~1100 and 1050 Ma, i.e., early Ottawan (Wodicka et al. 2003). In terms of structural style, NW-directed deformation of the Havre St-Pierre AMCG complex was strongly heterogeneous, leading to anorthosite sheets forming relatively open, upright antiforms, and the associated quartz-bearing lithologies, such as mangerite forming NW-verging, tight synforms. The latter are also the sites of high-strain zones, such as the Abbé Huard shear zone, which exhibits SE-plunging lineations defined by ribbon quartz fabrics (Gobeil et al. 2003).

The L*P* Natashquan domain in the hanging wall of the Abbé Huard shear zone is largely underlain by the >1.5 Ga Wakeham Group composed of meta-arenite, quartzite, metapelite, subordinate metaconglomerate, and felsic metavolcanic rocks. The group is injected by a suite of geon-11 mafic dykes and sills, and both can be traced into several marginal orthogneissic complexes, in which they occur as ductilely deformed rafts (e.g., Buit and Aguanish complexes in Fig. 11, La Romaine complex farther east; Gobeil et al. 2003; Gervais

Fig. 11. Map of the Côte Nord region (after Gobeil et al. 2003; see Fig. 3 for location) showing the aL*P* Natashquan domain in the hanging wall of the oblique-normal Abbé Huard shear zone (AHSZ). Schematic cross-section is after Rivers et al. (in press). The interpretation of gneiss domes in the Aguanish complex is after Gervais et al. (2004). The aM*P* St-Jean domain in the footwall of the shear zone consists of the Havre St-Pierre AMCG complex (HSP). The centres of two northwest-trending upright synclines in the Natashquan domain are underlain by greenschist-facies rocks and may comprise part of the Ottawan Orogenic Lid. LCSZ, Lac Caron shear zone; NSZ, Natashquan shear zone; RSZ, Romaine shear zone.



et al. 2004; Bonnet et al. 2005). In central Natashquan domain, the Wakeham Group is preserved at low grade (coexisting Chl–Ms–Bt ± Stp; Martignole et al. 1994), suggesting these areas may constitute part of the OOL. Although containing evidence for small-scale F1 isoclinal folds, the Wakeham Group in this area is essentially a stratigraphically upward-facing sequence deformed into regional, NNW-trending upright open F2 folds. Structural complexity increases in the marginal gneiss complexes, where grade of metamorphism is upper amphibolite facies (Gervais et al. 2004). In the Buit Complex, Wakeham metapelite exhibits the assemblages Crd–Grt–Ath and Qtz–Ms–Grt–Bt–And ± St ± Sil, the latter yielding peak *P*–*T* conditions of 350 MPa at 550 °C (Madore et al. 1999). U–Pb ages of metamorphic zircon, monazite, titanite, and rutile from the marginal amphibolite-facies complexes indicate a complex history. Firstly, some enclaves of the Wakeham Group in the La Romaine complex have yielded evidence for ~1.5 Ga granulite-facies metamorphism overprinted by ~1.0 Ga amphibolite-facies assemblages (van Breemen and Corriveau 2005). Secondly, Wodicka et al. (2003) reported four multigrain rutile ages of ~1060–1045 Ma from the Buit Complex, whereas metamorphic zircon and monazite ages from the Boulain and La Romaine complexes and Kataht intrusive suite are all younger than ~1030 Ma (~1030–1020 Ma concordant zircon and monazite ages or lower intercepts; ~1010–1000 Ma and ~966 Ma concordant monazite ages; Wodicka et al. 2003; van Breemen and Corriveau 2005). Finally, Wodicka et al. (2003) reported four single grain titanite ages ranging from ~970 to 940 Ma from the Aguanish complex. The preferred interpretation of these data is that the old rutile ages from the Buit Complex record cooling following early Ottawan M*P* metamorphism in the adjacent St-Jean domain, whereas the more widespread younger ages record a later episode of prograde metamorphism and subsequent protracted cooling in the marginal complexes of the L*P* Natashquan domain. This would imply prograde L*P* Ottawan metamorphism in much of the Natashquan domain took place some 40–50 million years after peak Ottawan metamorphism in the underlying M*P* St-Jean domain and was coeval with orogenic collapse elsewhere in the Grenville Province. Finally, the prominent NW-trending upright open folds in the Natashquan domain also affect the St-Jean domain (Fig. 11), implying they are late structures that developed after oblique-normal displacement on the Abbé Huard shear zone brought the two domains into tectonic contact.

Structures in allochthonous Medium-Pressure Belt

After a brief discussion of the rheology of the Ottawan mid crust, this section presents structural information from two areas of the aM*P* Belt in the southwest Grenville Province that provide insight into different aspects and scales of orogenic collapse in the mid crust: the regional structure of the Adirondack Highlands terrane and outcrop-scale structures in the Algonquin – Lac Dumoine terrane.

Rheology of the Ottawan mid crust

As noted previously, Ottawan crustal thickening was accompanied by peak metamorphic assemblages in uppermost amphibolite to granulite facies throughout the aM*P* Belt, for which a temperature range of ~800 ± 50 °C is estimated (Fig. 4). Under these conditions, pelitic and quartzofeldspathic rocks undergo dehydration melting of micas, compatible with the ubiquitous presence of leucosome in these lithologies. Since as little as 7% leucosome causes a dramatic reduction in viscosity (Rosenberg and Handy 2005; Rey et al. 2009), this implies that much of the mid crust was rheologically weak in early to mid Ottawan time, and it has been argued that it underwent some form of channel flow or hot nappe transport during the thickening phase (Jamieson et al. 2007, 2010; Rivers 2008, 2009).

Although orogenic collapse occurred on the retrograde *P*–*T* path, available data indicate it was initiated while the rocks were still hot. For instance, Culshaw et al. (1994) reported that sillimanite in some Sil–Kfs-bearing rocks near Georgian Bay locally defined the axial planar fabrics of folds developed during collapse. Moreover, microstructures in deformed leucosomes in these rocks exhibit normal-sense kinematics, indicating melt was still present and, hence, that the viscosity of the mid crust remained low. These observations imply that collapse was initiated under upper amphibolite-facies conditions, probably in the temperature range ~650–750 °C.

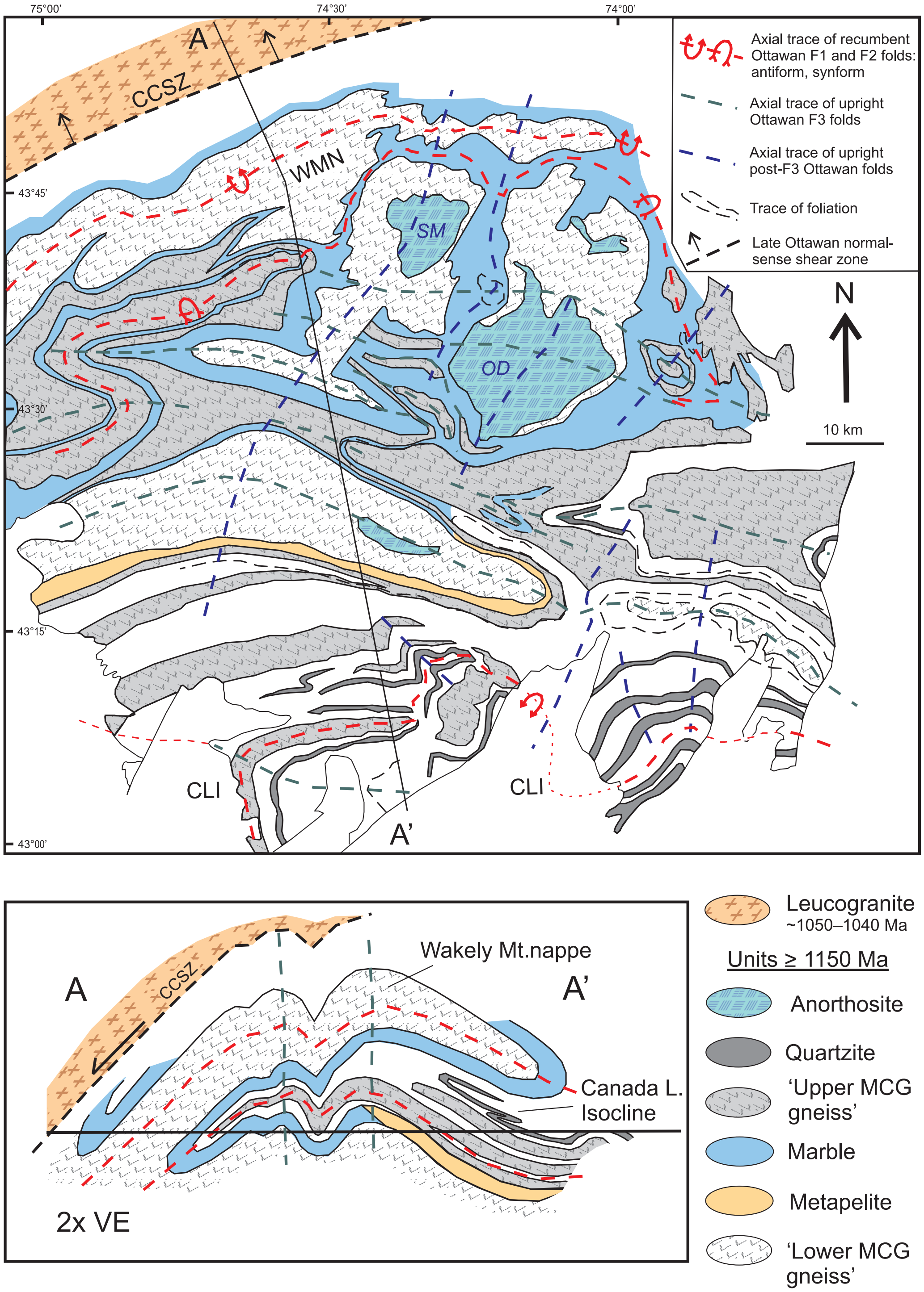
Orogenic collapse was locally accompanied by retrogression. Evidence for the former presence of hydrous fluid, possibly in part derived from the crystallization of leucosomes in gneissic units as the temperature waned, is indicated by the widespread presence of granitic pegmatite dykes and veins with amphibolite-facies reaction rims and by partial replacement of garnet by biotite in metapelitic rocks and pyroxene by hornblende and biotite in mafic rocks, respectively. These mid-amphibolite assemblages suggest retrogression occurred after significant cooling from peak temperatures. Crystallization of leucosome would have increased the viscosity of the mid crust, although this may have been partly offset by reaction softening associated with retrogression as hydrous fluid previously dissolved in leucosomes was released into the surroundings.

These observations suggest that although collapse occurred after the metamorphic peak, it was initiated while the temperature was still high and the mid-crustal rocks had a relatively low viscosity because of the presence of melt. Petrographic and mineral chemical observations, such as a decline in the grossular component in retrograde garnet and the formation of An-rich plagioclase pseudomorphs after garnet, suggest cooling was accompanied by decompression, an inference supported by several geothermobarometric studies (e.g., Annovitz and Essene 1990; Indares and Martignole 1990; Tuccillo et al. 1990, 1992; Jamieson et al. 1995). These *P*–*T* results are consistent with exhumation-driven cooling, as indicated by path 2a in Fig. 4.

Adirondack Highlands terrane

The regional structure of the Adirondack Highlands terrane was a topic of active research in the 1980s (e.g., Wiener et al. 1984; McLelland and Isachsen 1986), but more recent work has focussed on detailed studies of small critical areas. Figure 12 is a simplified map and cross-section of the central-southern part of the terrane after Wiener et al. (1984), schematically incorporating the data of Selleck et al. (2005) and McLelland et al. (2010) for the northwestern margin. Much of the terrane is underlain by deformed metaplutonic units of the geon-11 Marcy AMCG suite infolded with thin metasedi-

Fig. 12. Sketch map and cross-section of part of the Adirondack Highlands terrane showing superposed fold pattern (modified from Wiener et al. 1984; see Fig. 3 for location). Unpatterned areas are mostly metasediments. CLI, Canada Lake isocline; WMN, Wakely Mountain nappe; OD and SM, Oregon Dome and Snowy Mountain anorthosite massifs, respectively. Northwestern part of map and cross-section (area to the northwest of and including the Carthage–Colton shear zone (CCSZ)) are schematic, based on information in Selleck et al. (2005). VE, vertical exaggeration.



mentary units that together compose a tectonostratigraphy. Ottawan metamorphism reached granulite facies (McLelland et al. 2001), and the terrane is characterized by polyphase ductile deformation involving two early phases of recumbent folding and associated thrusting followed by several phases of more upright folding (McLelland and Isachsen 1986). First-generation recumbent folds predated the ~1155 Ma emplacement of the AMCG suite (McLelland et al. 1988), suggesting they are Shawinigan structures. The earliest Ottawan structures are regional isoclinal recumbent folds with east-trending axial traces, such as the Wakeley Mt. nappe and Canada Lake isocline (Fig. 12; F2 of McLelland and Isachsen 1986; F1$_{\text{Ottawan}}$ of McLelland et al. 2010), that are overturned to the southeast and accompanied by a penetrative subhorizontal axial planar foliation defined by granulite-facies mineral assemblages. The lower limbs of some of these structures are highly attenuated and may mark the locations of thrusts. In addition to folding gneisses derived from the ca. 1155 Ma AMCG suite, the F1$_{\text{Ottawan}}$ folds also affect the ca. 1095 Ma Hawkeye granite (McLelland and Isachsen 1986), which provides an upper age limit for these structures. This polydeformed package, with Shawinigan and F1$_{\text{Ottawan}}$ folds, was intruded by the ~1050 Ma Lyon Mountain leucogranite and refolded by large-scale, upright, coaxial, F2$_{\text{Ottawan}}$ folds that become tighter towards the north. According to Selleck et al. (2005, p. 781), "Field relationships unequivocally demonstrate granite emplacement [was] coeval with mylonitic deformation in a NW-SE–directed extensional strain field", an observation they interpreted as evidence for emplacement during orogenic collapse. On this basis, McLelland et al. (2010) argued that F2$_{\text{Ottawan}}$ folds formed at ca. 1050 Ma and were related to orogenic collapse. F3$_{\text{Ottawan}}$ folds are open, upright NNE-trending structures that fold the Lyon Mountain granite and give rise to dome and basin structures where they intersect F2$_{\text{Ottawan}}$ axes (Fig. 12). They were also inferred to be related to orogenic collapse by McLelland et al. (2010). The NW-trending folds in the northwest of the Highlands domain (shown in Fig. 8; F5 of Wiener et al. 1984) affect the high-strain rocks of the Carthage–Colton shear zone as noted previously, implying they are high-level structures that postdated juxtaposition of the Highlands and Lowlands domains. They also continue into the southern Highlands domain (McLelland and Isachsen 1986). McLelland et al. (2010) considered that the F2$_{\text{Ottawan}}$ and F3$_{\text{Ottawan}}$ structures formed in a manner comparable to the "a" and "b" folds that develop parallel (F2) and perpendicular (F3) to extension in a core complex terrane (e.g., Jolivet et al. 2004). This interpretation is discussed further in a later section.

Algonquin – Lac Dumoine terrane

Crustal thickening of the aM*P* Belt in the Algonquin – Lac Dumoine terrane has similarly been interpreted as a result of stacking of large thrust sheets and fold nappes followed by ductile extensional flow (e.g., Gower 1992; Jamieson et al. 1992; Culshaw et al. 1994, 1997), although an alternative model involving heterogeneous flow and indentor-driven expulsion of assembled nappes has also been advocated (Jamieson et al. 2007; Culshaw et al. 2010; see later discussion). The focus here is on the mechanisms of late- to post-Ottawan orogenic collapse inferred from detailed studies of the abundant shoreline and road outcrops around Georgian Bay, as reported by several authors (e.g., Schwerdtner 1987; Culshaw et al. 1994, 1997, 2010; Ketchum et al. 1998; Klemens and Schwerdtner 1997; Schwerdtner et al. 1998, 2005, 2010). On the basis of their results, the mechanics of crustal thinning in mid to post Ottawan time principally involved three processes: (*i*) normal-sense ductile shearing in zones several hundred metres to a kilometre or more wide at the boundaries of terranes and domains and also in narrower internal shear zones; (*ii*) distributed regional subvertical thinning and associated subhorizontal extension within the terranes and domains; and (*iii*) normal-sense displacement on narrow brittle–ductile shear zones that transect both the thickened crust and the ductile shear zones. These broadly correspond to simple shear (*i* and *iii*) and pure shear (*ii*) strain end members, respectively, and are discussed under these headings in the text that follows.

Normal-sense, simple-shear-dominated deformation

In a study of two ductile, SE-dipping, normal-sense, shear zones in the southwest Grenville Province, Culshaw et al. (1994) concluded that the straight-gneiss fabrics in the shear zones were formed by transposition of early Ottawan gneissic foliations developed during thrusting. They noted the southeast-plunging folds within the shear zones are parallel to the local stretching direction, and they inferred an origin either as sheath folds that preserve early Ottawan fabrics in their hinges or as more open structures formed by shearing of layers oblique to the shear plane ("a-folds"; Malavieille 1987; Jolivet et al. 2004). On the basis of dated dyke swarms, they concluded that the normal-sense shear zones were active between ~1040–1020 Ma, an interpretation subsequently confirmed for the Shawanaga shear zone (the local name for the Allochthon Boundary Thrust) by Ketchum et al. (1998), who documented normal-sense displacement at ~1020 Ma (Fig. 13C).

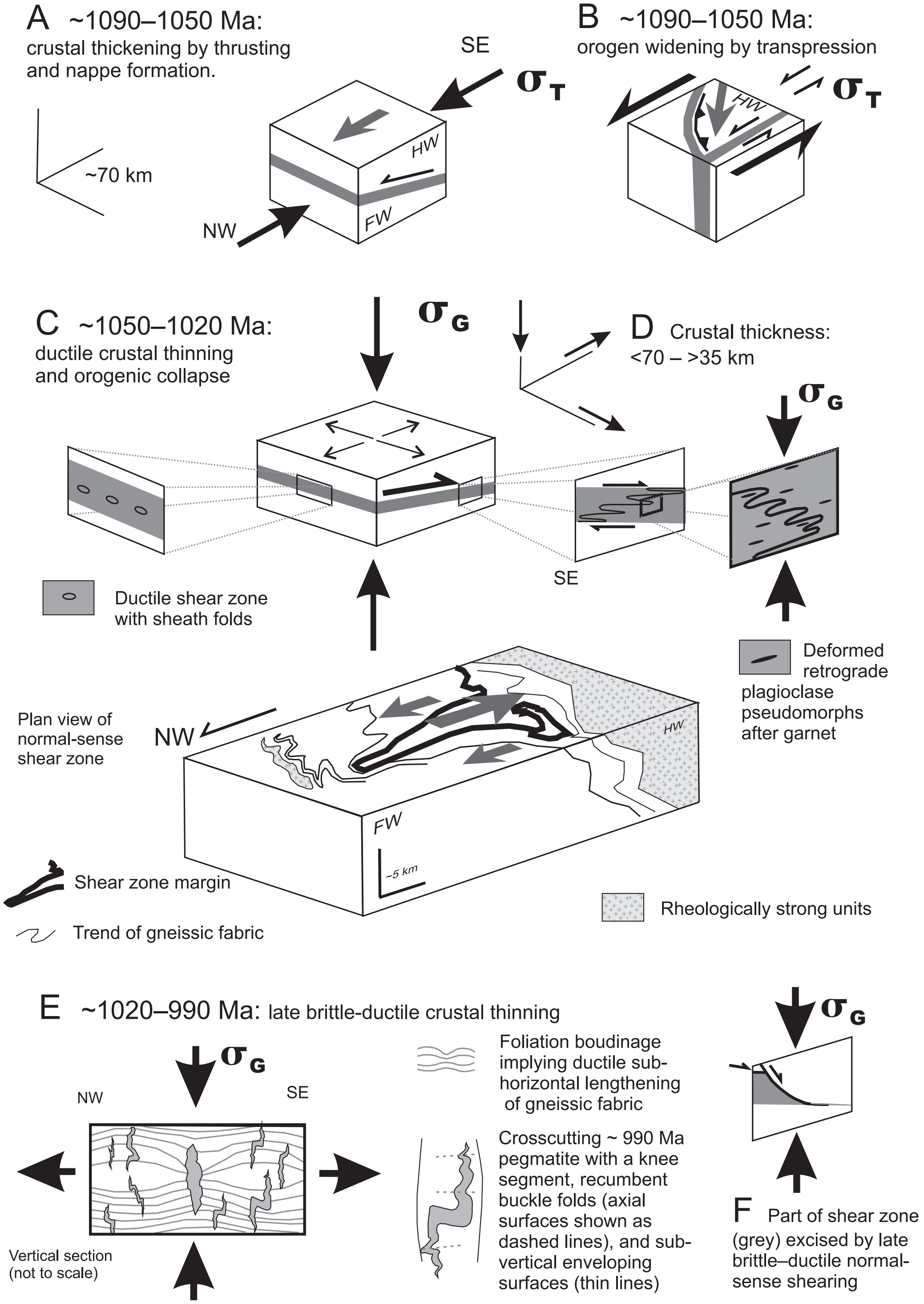
Normal-sense, simple-shear-dominated deformation was not uniquely ductile, however, and Schwerdtner et al. (2010) described a case in which a ductile strain gradient adjacent to a major thrust-sense shear zone is absent. They ascribed this geometry to excision of a large segment of the sole thrust and adjacent high-strain rocks from the observed section by normal-sense, brittle–ductile shear zones and brittle faults, the latter indicated by pseudotachylite. They inferred the excised segment may have been up to ~10–20 km long and presumably ≥1 km thick, this being the approximate thickness of the shear zone. This is illustrated schematically in Fig. 13F.

Normal-sense, pure-shear-dominated ductile deformation

A component of within-domain vertical thinning of the crust in the aM*P* Belt by pure shear has been inferred by several authors (e.g., Culshaw et al. 1994; Klemens and Schwerdtner 1997), but the significance of its contribution to orogenic collapse remains uncertain. However, it is suggested here that its importance may be qualitatively gauged from the dimensions of stacked thrust sheets (terranes and domains) in the southwest Grenville Province, some of which are ≥300 km long but only ~12–15 km thick. Thrust stacking is contingent on individual thrust slices having sufficient rheological strength to retain coherence during transport, but this seems improbable for sheets with a length-to-thickness ratio of ≥20 and evidence they were ductile at the time of thrusting (e.g., abundant leucosome; Culshaw et al. 1997). Hence, it is qualitatively concluded on rheological grounds that at the time of their emplacement the thrust sheets were considerably thicker, their present dimensions constituting a cryptic record of important vertical thinning, an inference supported by their penetrative S–L fabrics (Fig. 13C; Schwerdtner et al. 1977; Davidson 1983).

That such vertical thinning in gneisses of the Algonquin –

Fig. 13. Sketches illustrating structural evolution of mid crust in aM*P* Algonquin – Lac Dumoine terrane as determined from field studies in the area outlined by the box in Fig. 3. Note inferred change in orientation of maximum principal stress in the mid crust from subhorizontal ($\sigma_T^{mid} > \sigma_G^{mid}$) in A and B, to subvertical ($\sigma_G^{mid} > \sigma_T^{mid}$) in C to E (modified from Culshaw et al. 1994; Klemens and Schwerdtner 1997; Schwerdtner et al. 2005, 2010). (A–B) ~1090–1050 Ma: Crustal thickening and orogen widening by thrusting, nappe formation, and transpression. (C–D) ~1050–1020 Ma: initiation of orogenic collapse leading to thinning and lengthening of mid crust by ductile pure and simple shear mechanisms. (E–F) ~1020–990 Ma: continued late-stage orogenic collapse by brittle–ductile mechanisms, leading to minor vertical thinning and horizontal lengthening of mid crust.



Lac Dumoine terrane continued during and (or) after the Ottawan orogenic phase was demonstrated by Klemens and Schwerdtner (1997). Firstly, the subhorizontal gneisses exhibit foliation boudinage, implying horizontal extension. Secondly, they observed that the boudin necks are commonly the sites of ~990 Ma pegmatite dykes. Although individual dykes of this swarm are typically <1–2 m in width, their overall density suggests they represent a minor but significant mechanism of post-Ottawan subhorizontal crustal dilation. In addition, the common presence within the pegmatite dykes of short-wavelength buckle folds with subhorizontal axial surfaces and subvertical enveloping surfaces implies a later component of vertical shortening of the dykes and thinning of their gneissic envelope. Schwerdtner and Klemens (2008) noted that in one area the pegmatite dyke swarm was oriented approximately perpendicular to the subhorizontal axis of a late upright cross-fold that can be traced for at least 50 km. This suggests that by ~990 Ma, cooling of the gneissic mid-crust had proceeded to a stage where minor horizontal extension parallel to the axis of a large-scale upright fold led to opening of brittle fractures, thereby generating dilational sites for pegmatite emplacement. Although the pegmatites crosscut the gneissic fabric, many also exhibit short foliation-parallel segments ("knees"), suggesting their intrusion exploited local weaknesses in the subhorizontal gneissic layering and lifted the overlying crust. With respect to the later folding, since buckle folds develop in strong layers in a rheologically weaker matrix, they must have developed during crystallization of the pegmatites at a stage when, although still ductile, their strength was greater than that of the finer grained gneissic envelope (Klemens and Schwerdtner 1997). In any case, they document an increment of minor, but widespread vertical thinning of the crust after ~990 Ma, and they appear to represent the final stage of ductile collapse recorded in the mid crustal rocks. These features are illustrated in Fig. 13E.

Combined normal-sense, pure, and simple shear deformation

Despite their appearance as 2-dimensional (2-D) planar features on small-scale maps and cross-sections, the substantial width of some ductile normal-sense shear zones (≥1 km) implies they may be the sites of significant triaxial strain. Schwerdtner et al. (2005) demonstrated this to be the case at one location by recording a horizontal flattening fabric axial planar to buckle folding of lithotectonic layering at an angle to a gently-dipping shear zone boundary. This fabric, defined by the long axes of deformed, retrograde, plagioclase pseudomorphs after garnet that are axial planar to the buckle folds (Fig. 13D) provides evidence for an increment of subvertical flattening accompanying retrogression within the shear zone. Assuming this result is representative, it suggests that normal-sense shear zones were not only sites of simple shear displacement between the hanging wall and footwall, but also locations where the crust was thinned vertically and extended horizontally by pure shear. Moreover, inasmuch as the boundaries of such shear zones typically exhibit diffuse strain gradients into adjacent domains, this result is compatible with the inference of within-domain ductile thinning discussed previously.

Integration of Ottawan structural evolution in OOL, aL*P* Belt, and aM*P* Belt

Timing of Ottawan metamorphism in the aMP and aLP belts

The data presented earlier, summarized in Table 1, show that in all cases examined peak Ottawan metamorphism in the M*P* belt predated or was coeval with that in the adjacent L*P* segment. Moreover, the timing of L*P* metamorphism postdated initiation of normal-sense displacement on the shear zones bounding the L*P* segments. Assuming it is justifiable to infer that the few older (i.e., ≥1050 Ma) ages in the L*P* segments are related to metamorphism in the underlying M*P* footwalls, the remaining younger metamorphic ages in the L*P* segments (~1035–1020 and 1010–1000 Ma) can be interpreted as a signal of a second pulse of heating at a higher crustal level. Such an interpretation is reinforced by Table 2, which provides a summary of the Ottawan and post-Ottawan metamorphic and structural histories of the aM*P* Belt, aL*P* Belt, and OOL during the intervals ~1090–1050, 1050–1020, and 1020–990 Ma, as determined from the areas described in the previous sections. The time of peak temperature and the character of deformation at each crustal level are shown in the shaded boxes, highlighting the conclusion that metamorphism migrated progressively from the mid crust (aM*P* Belt), through the upper crust (aL*P* Belt) to the uppermost crust (OOL) over a period of ~100 Ma. Specifically, the data indicate that granulite-facies metamorphism in the aM*P* Belt took place from ~1090 to 1050 Ma at ~30 km depth and peak temperatures of ~850 °C; amphibolite-facies metamorphism in the aL*P* Belt took place from ~1030 to 1020 Ma (locally ~1000 Ma) at ~12–20 km depth under peak temperatures of ~500–750 °C; and heating and subsequent cooling in and adjacent to the OOL, which mostly occurred from ~1020 to 990 Ma or later, took place at ≤12 km depth and locally exceeded 500 °C, the T_C for Ar in hornblende. The author's preferred interpretation is to ascribe this pattern of heating to conduction as exhumed hot mid crust came into contact with successively higher crustal levels during orogenic collapse. In other words, the principal driving force for metamorphism in the aL*P* segments and subsequently for heating of the uppermost crust adjacent to the OOL was conductive heating associated with cooling of the orogenic mid crust during exhumation. This scenario is compatible with several other features discussed earlier, including the range of $^{40}Ar/^{39}Ar$ hornblende apparent ages in and adjacent to the OOL, the occurrence of isograds at high angles to structural trends in the Mazinaw terrane, and the presence of late transverse folds that cross the boundaries of the L*P* Natashquan and Adirondack Lowlands domains into adjacent M*P* domains. All these features are consistent with the inference that the late metamorphic and ductile structural evolution of the M*P* terranes took place in the upper crust after their juxtaposition with upper-crustal terranes.

Amount of displacement on shear zones bounding aLP segments

All three aL*P* segments are bounded by normal-sense shear zones that were active during and (or) after the L*P* metamorphism, the data implying displacement was initiated in the mid to late Ottawan. The amount of displacement is

Table 1. Summary of peak* P–T conditions and age data for Ottawan metamorphism in adjacent terranes of the aMP and aLP belts (*except the P–T estimate for the Composite Arc Belt (CAB) boundary shear zone, which is a minimum according to Hanmer and McEachern 1992) .

aMP Belt	aLP Belt	ΔP (MPa), Δd (km) , Δt (Ma)	References
Muskoka domain P: 1000–1150 MPa; T: 750–850 °C; Age: ~1080–1050 Ma	**Bancroft terrane – CAB boundary shear zone** P: ~600 MPa*; T: 600–650 °C*; Age: ~1080 Ma	**Muskoka–Bancroft** ΔP: 400–550 MPa; Δd: 14–18 km; Δt: approximately coeval	Annovitz and Essene 1990; Cosca et al. 1991; Hanmer and McEachern 1992; McEachern and van Breemen 1993; Corfu and Easton 1995; Culshaw et al. 1997; Timmermann et al. 1997, 2002; Slagstad et al. 2004
	Mazinaw terrane P: ~350–500 MPa; T: ~500–650 °C; Age: ~1090–1065; 1035–1025 Ma	**Bancroft–Mazinaw** ΔP: 100–250 MPa; Δd: 3.5–8.5 km; Δt: 45–55 Ma	
M-T P: ~1000–1100 MPa; T: 800–900 °C; Age: ~1120–1090 Ma	**PSM(w)** P: 300–600 MPa; T: 550–625 °C; Age: no data below	**M-T – Shawinigan** ΔP: 1000 MPa; Δd: 3 km; Δt: approximately coeval?	Corrigan and van Breemen 1997
Shawinigan domain P: ~700–1000 MPa; T: 725–825 °C; Age: no data; same as M-T?	**PSM(e)** P: 500–600 MPa; T: 750 °C; Age: ~1056 Ma	**Shawinigan – PSM(w)** ΔP: ~300 MPa; Δd: 10 km; Δt: no estimate	
		Shawinigan – PSM(e) ΔP: ~400 MPa; Δd: 14 km; Δt: 44 Ma	
St-Jean domain Granulite facies P: no data; T: no data; Age: ~1100–1050 Ma	**Natashquan domain** P: ≤350 MPa; T: 550 °C; Age: pre-1050 Ma, ~1030–1020 Ma, and ~1010–1000 Ma	**St-Jean – Natashquan** ΔP: ≥500 MPa ?; Δd: ≥16.5 km ?; Δt: ≥50 Ma	Madore et al. 1999; Gobeil et al. 2003; Wodicka et al. 2003; van Breemen and Corriveau 2005

Notes: (*i*) ? indicates no quantitative estimate. (*ii*) Where more than one age range for Ottawan metamorphism has been determined in the aLP Belt, those associated with orogenic collapse are underlined. (*iii*) ΔP, Δd, and Δt are differences in estimated pressure, depth, and time of peak metamorphism between adjacent MP and LP segments; where a range of P and t estimates is given in the first two columns, the median value was used to calculate the values in the third column. (*iv*) Estimated depths were calculated assuming an average density of 2800 kg.m^{-3} and results were rounded. (*v*) M-T, Mékinak–Taureau terrane; PSM(e) or (w), Porteneuf–St-Maurice domain (east strand) or (west strand).

Table 2. Summary of Ottawan and post-Ottawan metamorphic and structural evolution of a*MP* Belt, a*LP* Belt, and Ottawan Orogenic Lid (OOL) in the intervals ~1090–1050, ~1050–1020, and ~1020–990 Ma.

a*MP* Belt	a*LP* Belt	OOL
~1090–1050 Ma		
Metamorphism: Prograde granulite facies; widespread growth of metamorphic Zrn and Mnz	*Metamorphism*: Local growth of metamorphic Rt and Ttn adjacent to a*MP* Belt	*Metamorphism*: Predominantly pre-Grenvillian: Ottawan $T < T_c$ for Ar in Hbl
Structure: Thrust imbrication, recumbent nappes (dominant), upright refolding (minor)	*Structure*: Local ductile deformation adjacent to a*MP* Belt	*Structure*: Predominantly inherited pre-Grenvillian structures
~1050–1020 Ma		
Metamorphism: Decompression and retrogression to amphibolite facies	*Metamorphism*: Prograde amphibolite facies; growth of metamorphic Zrn and Mnz	*Metamorphism*: Predominantly pre-Grenvillian: Ottawan $T < T_c$ for Ar in Hbl
Structure: Initiation of orogenic collapse – formation of subhoriontal fabrics by vertical thinning and displacement on normal-sense shear zones	*Structure*: Thrust imbrication (minor), upright refolding (dominant), late recumbent folding (minor)	*Metamorphism*: Predominantly inherited pre-Grenvillian structures
~1020–990 Ma		
Metamorphism: Cooling through T_c for Ar in Hbl	*Metamorphism*: Amphibolite facies (retrograde); growth of metamorphic Zrn and Mnz; cooling through T_c for Ar in Hbl	*Metamorphism*: $<T_c$ for Ar in Hbl; margins of OOL heated to $\geq T_c$ for Ar in Hbl because of juxtaposition with a*MP* Belt
Structure: Late upright cross-folds; minor vertical thinning of crust; normal-sense brittle–ductile displacement on bounding shear zones	*Structure*: Late upright cross-folds that postdate normal-sense displacement on bounding shear zones	*Structure*: Normal-sense displacement on bounding shear zones, late upright cross-folds

Note: Peak Ottawan *P–T* conditions at each structural level are indicated in the dark grey cells, post-peak and retrograde conditions in the light grey cells. Minor effects predating the peak metamorphism are indicated in the unshaded cells.

constrained by *P* estimates for the Ottawan metamorphism in adjacent footwall and hanging wall pairs, which translate into depth differences of 14–18 km between the Muskoka domain and the Bancroft terrane – Composite Arc Belt boundary shear zone, ~3.5–8.5 km between the Bancroft terrane and the Mazinaw terrane, and ~11–22 km between the Mékinak–Taureau terrane and the Portneuf–St-Maurice domain (Table 1). Although uncertainties on the estimates are large, the data nonetheless indicate that the (oblique)-normal shear zones bordering the a*LP* Belt were the sites of major vertical displacement, and hence, they were first-order features of the mid and upper crust in late Ottawan time.

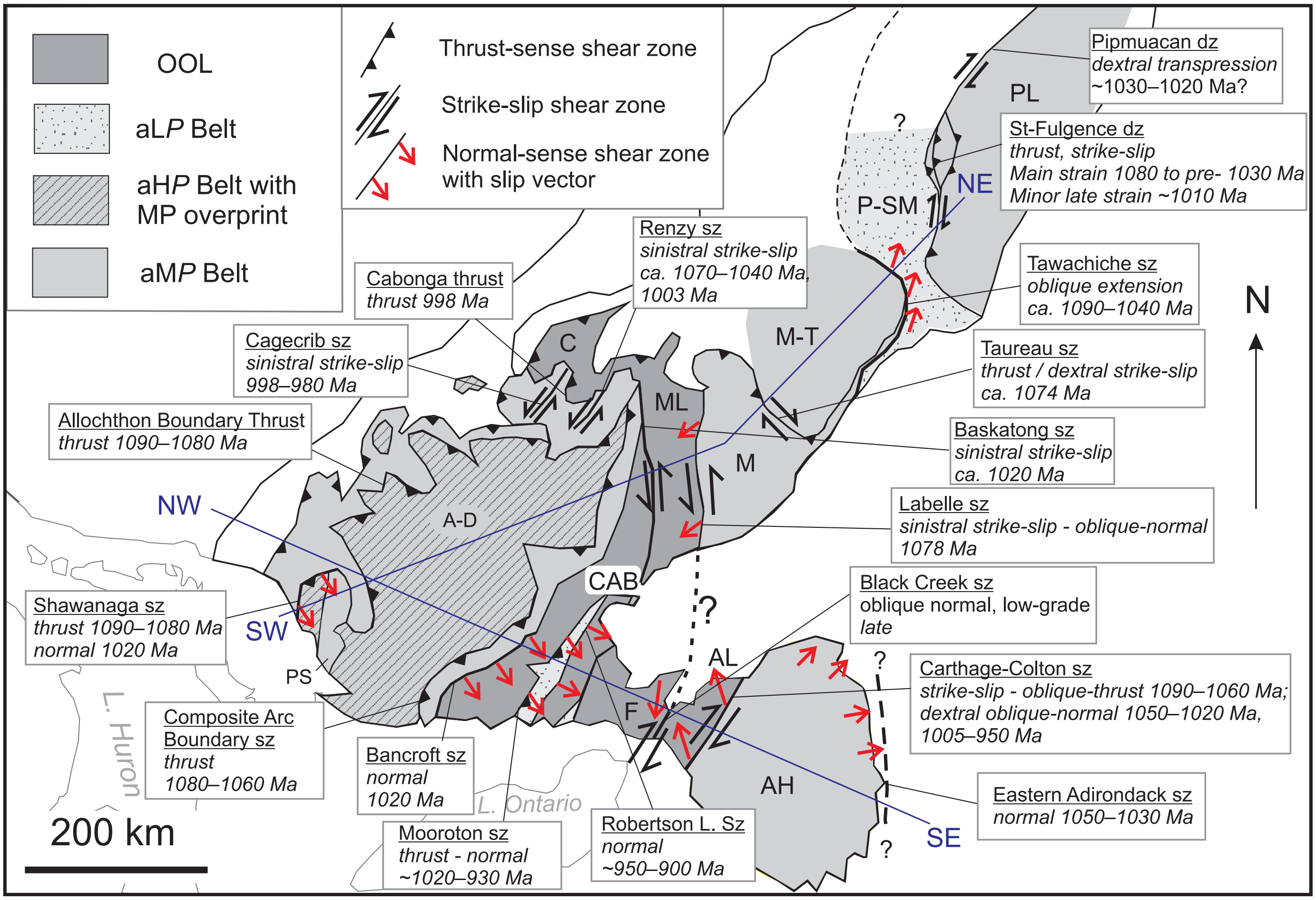
Contrasts in time of deformation, dominant structural style, and structural evolution between mid and upper orogenic crust

Considering first the age of the dominant structures, it is apparent from Table 2 that those in the OOL are principally pre-Ottawan, those in the a*LP* Belt are principally late Ottawan, and those in the a*MP* Belt range from early to mid Ottawan (thrust-sense) to late and post Ottawan (normal-sense). As discussed previously, this distribution can be accommodated by a model in which the active Ottawan deformation progressed from the mid crust to the upper crust, with the uppermost crust escaping penetrative strain.

With respect to structural style, the *LP* Mazinaw terrane and Natashquan domain are both characterized by folds with steep to upright axial surfaces and subhorizontal to gently plunging axes and similarly oriented late Ottawan folds also occur in the Adirondack Lowlands domain, part of the OOL. Open, upright folds such as these are characteristic of an upper-crustal structural style in which variable degrees of horizontal shortening are accompanied by vertical thickening in a bulk pure shear regime. These upright structures provide a striking contrast to the overturned to recumbent axial surfaces of the regional fold and thrust nappes in the a*MP* Belt that imply subhorizontal tectonic transport during crustal thickening and subvertical flattening during subsequent collapse.

This difference in dominant structural style between the a*LP* and a*MP* belts points to the existence of an important rheological discontinuity between the suprastructure and infrastructure in Ottawan time, with the lower degree of strain in the suprastructure, indicating its greater rheological strength (see Culshaw et al. 2006 for a recent discussion of tectonic styles in orogenic suprastructure and infrastructure). Moreover, the Ottawan structural evolutions of M*P* segments, such as the Adirondack Highlands terrane and St-Jean domain, exhibit a progression from overturned–recumbent structures followed by upright structures, which can be interpreted as a temporal record of subhorizontal tectonic transport followed by bulk vertical thickening that may correlate with exhumation from the mid to upper crust. In contrast, evidence in the *LP* Mazinaw terrane for minor Ottawan F3 folds with subhorizontal axial surfaces superimposed on F2 folds with subvertical axial surfaces implies incipient orogenic collapse of the upper crust, a muted reflection of the dominant process in the underlying a*MP* Belt. However, in this case the structural progression is from upright to subhorizontal structures, the opposite of that in the a*MP* Belt.

Fig. 14. Sketch map of the southwest Grenville Province showing locations, directions and time of displacement of major structures active during orogenic collapse (incorporating information from Mezger et al. 1991; Busch et al. 1996*a*; Corrigan and van Breemen 1997; Cureton et al. 1997; Ketchum et al. 1998; Martignole and Friedman 1998; Carr et al. 2000; Johnson et al. 2004; Selleck et al. 2005; Baird et al. 2008; Hébert et al. 2009; McLelland et al. 2009, 2010). A-D, Algonquin – Lac Dumoine terrane; AH, Adirondack Highlands terrane; CAB, Composite Arc Belt; C, Cabonga domain; F, Frontenac domain; ML, Mont-Laurier domain; M, Morin terrane; MT, Mékinak-Taureau terrane; PL, Parc des Laurentides domain; PS, Parry Sound domain; P-SM, Portneuf–St-Maurice domain. Locations of NW–SE and SW–NE cross-sections in Fig. 17 are shown.



Strain mechanisms and driving force for collapse in orogenic mid crust

The inferred evolution of crustal thickness, strain mechanisms, and principal stress directions in the Ottawan mid crust is summarized in Fig. 13. Doubling of crustal thickness by imbrication of thrust and fold nappes in early Ottawan time (Fig. 13A) was followed by lateral tectonic extrusion and the development of a transpressional strike-slip system (Fig. 13B). These observations suggest that σ_T^{mid} was the dominant force in the mid crust throughout this stage. Subsequent evolution in the mid to late Ottawan records thinning of the mid crust through a combination of simple shear on major, subhorizontal, normal-sense shear zones and vertical pure-shear flattening, the latter occurring both within the shear zones and in the intervening crust making up the stacked thrust and fold nappes (Fig. 13C, 13D). These structures imply σ_G^{mid} was the principal force in the mid crust at this time, with brittle–ductile, post-Ottawan structures of similar orientation indicating its continued dominant role as the crust cooled and stiffened (Fig. 13F).

In this context, it is worth emphasising the obvious point that without appropriately oriented steep strain markers, vertical crustal shortening is cryptic in rocks with subhorizontal layering, such as those that characterize much of the aM*P* Belt. As noted, the amount of strain recorded by the ~990 Ma pegmatites is minor, but there is currently no estimate of the presumably much greater amount of mid-crustal thinning in the interval ~1040–990 Ma.

Figure 13C illustrates the interpretation that normal-sense displacement on major shear zones led to the formation of cross-folds and associated lineations that tracked the local stretching direction during collapse. In the figure, which is based on data from near Georgian Bay (Culshaw et al. 1994), these developed at ~1020 Ma and are approximately southeast-trending, but elsewhere other directions and times have been recorded.

Figure 14 is a compilation of the orientations and timing of kinematic indicators in major shear zones in the southwest Grenville Province from mid Ottawan to Rigolet time. From a time-integrated perspective, the overall pattern of extension directions may be considered quasi-radial. Moreover, of relevance to later discussion is the observation that extension associated with some major shear zones was approximately orogen parallel, e.g., Tawachiche and Labelle shear zones. Figure 14 also indicates that extensional displacement overlapped temporally with Ottawan shortening in the orogenic hinterland and with Rigolet shortening in the Parautochthonous Belt. Considered with the quasi-radial pattern, this implies vertical thinning of the Ottawan mid crust was driven by gravitational loading rather than by a regional tensional regime, in accord with theoretical considerations. Although this interpretation is not new, it is more robust than previous inferences on account of the size of the area considered, the larger database, and the wider range of stretching directions. Finally, it is apparent from Fig. 14 that not all the margins of the OOL are the sites of normal-sense shear zones — for instance, thrust emplacement of the Cabonga Allochthon on the Cabonga thrust at ~998 Ma (Martignole and Friedman 1998) was coeval with normal-sense displacements at the margin of the OOL elsewhere. This issue is taken up in a later section.

The aMP Belt as a series of core complexes

As previously discussed, when the crustal-scale, normal-sense structures that sole in the mid crust (Fig. 3) are integrated with the temperature distribution of the orogenic crust (Fig. 4), the resemblance to a series of mid-crustal core complexes (underlain by the hot aM*P* Belt) adjacent to the remnants of their upper-crustal cover (underlain by the cooler aL*P* Belt and OOL) is apparent. This interpretation is compatible with the generic results of numerical modelling of crustal-scale core complex formation (Rey et al. 2009) and also with the structural evolution recorded in individual areas of the Grenville Province. For instance, M*P* terranes, such as the Adirondack Highlands terrane, Mékinak–Taureau terrane and St-Jean domain, are all dominated by large-scale overturned to recumbent thrust and fold nappes that were exhumed against upper to uppermost orogenic crust with steep to upright structures (Adirondack Lowlands, Portneuf–St-Maurice, and Natashquan domains of the OOL and aL*P* respectively; Figs. 8, 10, 11). Moreover, with respect to the L*P* Mazinaw terrane (Fig. 9), it was argued that it resembles a relatively hot upper-crustal horst surrounded by a cool graben of uppermost crust (the OOL). Thus, by late Ottawan time, it is inferred that widespread adjustment in crustal structure was in progress driven by the excess potential energy of the decaying plateau acting on the weak mid crust (σ_G^{mid}). The overall result of this process was thinning of the mid crust by as much as a third and its rise into its cooler tectonic cover. The structures in Fig. 13 imply that the exhumed mid crust remained hot and ductile, undergoing vertical thinning during exhumation until at least ~990 Ma. This inference is compatible with both the late Ottawan age of metamorphism in the overlying aL*P* Belt, and with the ^{40}Ar/^{39}Ar hornblende apparent ages surrounding the OOL, which it has been argued were a result of heat transfer from the Ottawan mid crust during its exhumation.

Late folds and thrusts and the driving force for late contraction

It was argued earlier that the latest structural evolution of the M*P* terranes took place after their juxtaposition with the L*P* terranes and the OOL, and it is recorded by one or more generations of high-level, upright folds that deform their mutual boundaries, e.g., NW-trending folds in Figs. 8, 11, NNE-trending folds in Fig. 12). These folds are transverse to the orogenic grain, i.e., cross-folds. Assuming they are buckle folds, as their shapes suggest, their orientations imply they developed by shortening at high angles to the orogenic front.

On the other hand, Fig. 13C illustrates northwest-trending cross-folds in the Algonquin – Lac Dumoine terrane that are associated with ductile, high-temperature fabrics and prominent co-axial lineations defined by sillimanite and deformed leucosomes. These are mid-crustal structures that formed at an early stage of exhumation. They were interpreted as a-folds by Culshaw et al. (1994), implying they developed by shearing of layering oblique to the shear plane and that their axes tracked the local extension direction (e.g., Malavieille 1987; Jolivet et al. 2004). A similar interpretation was recently adopted by McLelland et al. (2010) for the east–west-trending (i.e., orogen-oblique) ductile folds in the Adirondack Highlands terrane. On the other hand, Schwerdtner and van Berkel (1991) argued that prominent open, northwest-trending, non-cylindrical, cross-folds of high-grade layered gneisses in the M*P* Muskoka terrane, structurally higher in the stack, were buckle folds. This interpretation may be supported by the study of Hobbs et al. (2007), which indicates buckle folds can form in layered sequences even when the viscosity contrast between adjacent layers is very small.

A related question concerns whether any of the high-level, late folds were a product of gravity sliding of nappes off topographic highs formed over exhumed core complexes (e.g., Ramberg 1981). Such an origin is plausible insofar as their trends are commonly at high angles to the orogenic front, and, hence, not readily attributable to the NW–SE regional compressional stress system that resulted in crustal thickening. However, robust evidence for gravity sliding in the Grenville Province is so far lacking. Ideally, the outcrop patterns of nappes resulting from this process should be characterized by thrust structures at their leading edges and normal structures at their trailing edges, and they may be internally deformed by late buckle folds, as discussed. One potential candidate is the Cabonga allochthon, a nappe forming part of the OOL, that was emplaced at ~998 Ma (Fig. 14; Martignole and Friedman 1998). Its inclusion in the OOL implies it was cool (<500 °C) at the time of its emplacement, and since it is in contact with the mid crust, the possibility arises that it may have originated by sliding off an exhumed mid-crustal core complex. Another potential candidate is the undated Natashquan shear zone in the Côte Nord region, which carried the Natashquan domain over the adjacent granitoid terrane to the east (Gobeil et al. 2003; Fig. 11).

However, a counter example was illustrated by Rivers (2009), who showed that the M*P* Lake Melville terrane in the eastern Grenville Province was exhumed to the upper crust and juxtaposed against the Mealy Mountains terrane, part of the OOL, before folding of its high-*T* gneissic fabric into a regional, east-trending upright anticlinorium. In this case, although the evidence similarly implies that contractional deformation continued after initiation of orogenic collapse, the folding took place at high temperature and axial traces of the folds are approximately parallel to the orogenic front rather than perpendicular to it, suggesting an origin by regional compression is more likely.

To conclude, the cause of the late cross-folding in the mid and upper crust and late thrusting in the upper crust remains poorly constrained and is clearly a topic that warrants further research. However, the fact that the cross-folds developed at a range of crustal levels and orientations, and both predated and postdated juxtaposition with the OOL raises the possibility that they may have developed by more than one mechanism.

Coeval magmatic and hydrothermal activity

The spatial and temporal link between leucogranite emplacement and orogenic collapse has already been drawn with respect to the Lyon Mountain granite in the Adirondack Highlands terrane (Selleck et al. 2005). The origin of the leucogranite magma was discussed by Heumann et al. (2006) and McLelland et al. (2010), who proposed that melting was a result of delamination of overthickened crust followed by asthenospheric upwelling. Another possible explanation is decompression-driven melting of anhydrous, granulite-facies mid crust during exhumation, which would have occurred if the *P–T* path intersected the positive slope of the anhydrous melting curve. In this section, four other topics relevant to magmatism during orogenic collapse are addressed: (1) the thermal contribution of geon-10 AMCG intrusions to the temperature of the Ottawan mid crust, (2) the tectonic significance of the geon-10 K-rich alkaline to shoshonitic intrusions emplaced into the OOL, (3) the significance of pegmatite swarms emplaced in the OOL in geons 10 and 9, and (4) evidence for hydrothermal activity.

Geon-10 AMCG suites

Syn-Ottawan AMCG suites in the Grenville Province, although smaller than the voluminous massifs emplaced in southeast Laurentia during geon 11, were nonetheless of considerable size and their emplacement and crystallization must have contributed a significant amount of heat to the mid crust. It is, therefore, pertinent to assess whether it could have promoted orogenic collapse by weakening the mid crust. The issue was first addressed by Corrigan and van Breemen (1997) with respect to the Mauricie region, where geon-10 AMCG intrusions were emplaced not only into the footwall and hanging wall of the Tawachiche shear zone, but also into the shear zone itself (Fig. 10). This distribution, together with field evidence for local melting, led these authors to argue that the local temperature of the crust was significantly augmented by advected magmatic heat and the heat of crystallization, resulting in critical weakening and lubrication of the shear zone. In particular, they speculated that the role of the voluminous, ~1080–1060 Ma Parc des Laurentides complex in the hanging wall of the shear zone may have been especially significant. This unit ranges in composition from norite through mangerite and monzonite to hornblende granite, the predominantly anhydrous character of these lithologies suggesting liquidus temperatures in excess of 850 °C. As such, it is reasonable to infer their emplacement and crystallization would have advected a significant amount of heat into the crust. On the other hand, another area with abundant geon-10 AMCG magmatism, the 30 km wide corridor between the St-Fulgence and Pipmuacan shear zones in the central Grenville Province that is largely occupied by anorthosite and mangerite of the ~1080–1045 Ma Pipmuacan AMCG suite (Hébert et al. 2009) has not so far yielded evidence for normal-sense structures associated with orogenic collapse. Moreover as already discussed, Culshaw et al. (1994) and others have reported abundant evidence for collapse in the Georgian Bay area that lacks geon-10 AMCG suites. In summary, although an increased mantle heat flow beneath the Grenvillian Interior Magmatic Belt is probable, there is no compelling evidence to suggest that heat advected by geon-10 AMCG suites was an essential contributing factor to the high temperature and low viscosity of the Ottawan mid crust on the scale of the orogen, though it may have been significant locally.

Geon-10 K-rich alkaline to shoshonitic intrusions

The Ottawan Orogenic Lid is the site of emplacement of about 25 small, potassic to ultrapotassic, alkaline to shoshonitic plutons belonging to the Kensington–Skootamatta suite (Corriveau et al. 1990; Easton 1992; Corriveau and Gorton 1993), of which six have been dated and yielded crystallization ages of ~1090–1060 Ma. Individual plutons are composite, ranging from pyroxenite through diorite to monzonite and syenite. The compositional range is interpreted to reflect both mantle and crustal sources, as well as magma chamber processes such as fractionation and magma mixing, which in combination led to two geochemical series, one silica-undersaturated and the other critically silica-saturated (Corriveau and Gorton 1993). Also included in the Kensington–Skootamatta suite is a ~1070 Ma minette dyke that contains abundant xenoliths of crustal and mantle origin (Corriveau and Morin 2000). All intrusions of the suite are undeformed, whereas xenoliths in the minette dyke exhibit foliated, gneissic and mylonitic fabrics, the two features collectively indicating the contrasting strain in the uppermost versus mid to lower crust in Ottawan time. In terms of chemistry, both the mantle components of these intrusions and the mantle xenoliths in the minette dyke exhibit an enriched geochemical signature that is interpreted to imply a deep lithospheric origin. Taken together with the arc-like signatures of the ultrapotassic to shoshonitic magmas, this led Corriveau and Morin (2000) to propose the suite originated by delamination-driven melting of a previously enriched metasomatized mantle wedge above the former Elzevirian subduction zone.

With respect to the present discussion, the significance of the Kensington–Skootamatta suite and its location in the OOL lies in its modern analogue in the Tibetan Plateau, where many small active volcanic centres erupt products ranging from shoshonitic to ultrapotassic trachybasalt and trachyte, with subordinate trachyandesite, rhyolite, and basalt. These magmas are similarly interpreted to have formed by variable degrees of mixing and fractionation between crustal- and mantle-derived liquids (Hacker et al. 2000; Gao et al. 2010). The source of the Tibetan potassic volcanism is inferred to be partial melting of mantle enriched during an earlier period of subduction and mixed with sediment-derived melt, although in this case the melting was inferred to have been triggered by either convective thinning of overthickened mantle lithosphere or slab break-off rather than delamination. Regardless of the physical trigger, it appears that this style of magmatism may be a signature of the upper crust of an orogenic plateau, an association that was not recognized in the

Grenville Province until recently (Corrigan and Hanmer 1997; Hynes and Rivers 2010). Moreover, the foliated to gneissic crustal xenoliths in the undeformed minette dyke provide the first critical evidence from a single location for the contrast between the coeval high strain in the hot mid to lower Ottawan crust and the lack of penetrative strain in the cool upper crust forming the Ottawan Orogenic Lid.

Geons 10 and 9 pegmatite swarms

Late tectonic pegmatites, both rare-element-bearing varieties with enrichments in U, Th, Mo, Nb, REE, as well as so-called barren pegmatites, are widespread in the OOL and adjacent aM*P* Belt in the southwest Grenville Province of Ontario and Quebec. Where intruded into marble in the OOL, the pegmatites are commonly associated with skarns. They are interpreted to have originated by fractional crystallization of hydrous, A-type granitoid magmas emplaced in the mid crust between ~1090–1050 Ma. According to Lentz (1991, 1992, 1996), fractionation of these magmas led to the formation of pegmatitic liquids and partitioning of rare elements into a fluid phase, both of which migrated higher in the crust. Most U–Pb crystallization ages of pegmatites in the OOL range from ~1070–1020 Ma (Easton 1986; Lumbers et al. 1990; Easton and Kamo 2008). These results are supported by direct dating of molybdenite by the Re–Os method that has yielded ages of ~1070 Ma for a mineralized skarn and ~1050 Ma for a pegmatite (Lentz and Suzuki 2000; Lentz and Creaser 2005).

Pegmatites are also abundant in the adjacent aM*P* Belt (Algonquin – Lac Dumoine terrane; see distribution in Lentz 1996), where some swarms are as young as ~990–980 Ma (Easton 1986; Bussy et al. 1995). Phase relations of rare-element pegmatites in the Algonquin – Lac Dumoine terrane indicate crystallization at pressures of ~400 MPa and temperatures of ~400–450 °C (Lentz 1996). This implies crystallization occurred in the upper crust after the mid-crustal rocks had undergone significant exhumation and cooling, compatible with an origin during orogenic collapse. Moreover, it is also consistent with the locations of some pegmatites in late tectonic structures (Martignole and Friedman 1998). In the M*P* Adirondack Highlands terrane, pegmatite emplacement took place from ~1090 to 1000 Ma (Lupulescu et al. 2011), with an outlier at ~950 Ma occurring in an area affected by extensive hydrothermal alteration (Mineville; see next section). On the other hand, in the adjacent Adirondack Lowlands domain, part of the OOL, all pegmatites dated by these authors yielded crystallization ages >1150 Ma, implying they were unrelated to Ottawan metamorphism.

In addition to their association with orogenic collapse, the late pegmatites are also relevant to the present discussion because of their elevated U, Th, and K contents, which contribute to a distinctive airborne radiometric signal over the OOL and adjacent aM*P* Belt in the southwest Grenville Province (e.g., Lentz 1996). Guillou-Frottier et al. (1995) and Mareschal et al. (2000) pointed out that much of the Grenville Province, with the exception of the ≤1.3 Ga accreted terranes, stands out as a region of anomalously low heat flow, from which they inferred that Grenvillian mid to lower crust was depleted in radio-elements and, thus, was probably relatively mafic. However this interpretation is not supported by field observations in the aH*P* and aM*P* belts in which mafic rocks are volumetrically minor. Moreover, the presence in the upper crust of abundant pegmatites enriched in U, Th, and K provides an explanation for the low heat flow in the mid crust if, following the petrogenetic model described previously, the aM*P* Belt is the residuum from which the pegmatitic melts and associated fluids were extracted. In this scenario, the low heat flow in much of the Grenville Province is a secondary feature that resulted from migration of the U-, Th-, and K-enriched pegmatitic melts and associated fluids from the mid to the upper crust during orogenic collapse. This reasoning not only provides an explanation for the anomalously low heat flow of the aM*P* Belt and much higher heat flow in the OOL. It also posits that the radio-elements were initially resident in the mid crust during the early Ottawan crustal thickening stage, at which time their decay would have contributed to the elevated temperature, and hence low viscosity, of the orogenic mid crust. In summary, orogenic collapse may have been associated with important element redistribution that fundamentally altered the internal heat-producing potential of the orogenic crust, a conclusion that remained cryptic until recognition of the Ottawan Orogenic Lid as a tectonic entity.

Coeval hydrothermal activity

The evidence for widespread syn-Ottawan hydrothermal activity and associated IO(CG) mineralization was discussed in relation to the distribution of “young” (geon-9) ^{40}Ar/^{39}Ar hornblende apparent ages in the Adirondack Highlands terrane. In addition to the magnetite deposit at Lyon Mountain, others occur at Clintonville, Mineville, Benson, and Skiff Mountain in the Adirondack Highlands terrane and at St-Jérôme in southern Morin terrane (for locations see Fig. 2). Their alteration patterns provide a fingerprint of the channelling of hydrothermal fluids along normal-sense shear zones during orogenic collapse. An additional attribute of some of the Fe deposits in the Adirondack Highlands is that the magnetite ore formed rod-shaped bodies parallel to the east- to northeast-trending stretching lineations associated with the Eastern Adirondack shear zone (Bickford et al. 2008; McLelland et al. 2010), strengthening their link to the collapse process. The presence of these deposits raises the question of whether there are other areas of the Grenville Province that exhibit evidence for hydrothermal fluid activity in mid to late Ottawan time that could be related to orogenic collapse. The brief summary that follows is not intended to be exhaustive, but rather to indicate how widespread such a process might have been.

Another area with evidence for IOCG mineralization is near Manitou Lake in eastern Quebec (Fig. 2), where the Kwyjibo deposits contain elevated concentrations of Cu, rare-earth elements (REE), Y, P, and F, are enriched in Ag, Th, U, Mo, W, Zr, and Au, and exhibit evidence for multiple generations of fluid alteration, e.g., sodic (albite), potassic (microcline), and Fe-rich (magnetite) (Gauthier et al. 2004; Clark et al. 2005; Magrina et al. 2005). The mineralization occurs in a geon-11 granite, is spatially associated with a major south-dipping shear zone, and the deposits were metamorphosed under amphibolite- to granulite-facies conditions — hence, similar in several respects to the Adirondack Highlands examples. Although there is no consensus, a majority of those who have published on the deposit considers the mineralizing fluids to have been coeval with post-Ottawan collapse, as indicated by a titanite age of ~970 Ma, and that collapse was either the principal driving force (Gauthier et al. 2004) or part of an extended late- to post-orogenic event that also involved fluids from nearby post-tectonic plutons (Clark et al. 2005). IO(CG) mineralization is also well known in the New Jersey Highlands inlier (e.g., Edison, Bemco – Sulfur Hill, Dover, Warren, and Scrub Oaks deposits; Gauthier and Chartrand 2005: see Fig. 2 for locations). In this area, hydrothermal activity took place from ~990 to 940 Ma, resulted in sub-economic to economic concentrations of magnetite ± U ± Th ± REE, and the fluids reacted to form skarns where they came into contact with carbonate rocks (Volkert et al. 2005).

Another area with evidence for late fluid activity is near the northern margin of the Composite Arc Belt in the southwestern Grenville Province, specifically along the boundary between the OOL and adjacent aM*P* Belt (see Fig. 2 for location). In describing this area, Lumbers et al. (1990, p. 270) noted "low-grade zones, several-hundred-metres wide that underwent metasomatism by high-temperature alkalic fluids which rose through the crust along dilatant fractures… [t]heir existence … marked by a broad zone (up to 20-km wide) of fenite extending along the northern part of the CMB… and outside the fenite zone by granite pegmatite dykes and small bodies of pegmatitic, alkali feldspar granite". The timing of fenetization (alkali metasomatism) was bracketed between ~1070 and 1040 Ma, thus, overlapping with pegmatite emplacement outside the fenite zone, and it was considered to have developed in an environment of "late extension," all features that are compatible with orogenic collapse. In common with many fenite zones, this one is spatially associated with carbonatite. U–Pb dating of the carbonatite, in this case, has yielded zircon ages of ~1170–1150 Ma (emplacement?) and titanite ages of ~1080 Ma (hydrothermal event?) (Moecher et al. 1997), suggesting the fluids causing fenetization may have followed a long-lived zone of crustal weakness. Other evidence for fluid activity in the OOL was reported by Lentz and Creaser (2005), who described albitization of granitic bodies adjacent to a mineralized pegmatite, and by Bonamici et al. (2010) who presented oxygen isotope evidence for late infiltration of upper-crustal fluids along the Carthage–Colton shear zone. Moreover, fluids were not limited to the OOL because Morrison and Valley (1988) reported cathodoluminescence evidence for thin calcite veins (≤ 0.05 mm) and partial retrogression of anorthosite (to sericite, chlorite, epidote, etc.) in the Adirondack Highlands terrane in the footwall of the OOL. Although undated, the *P*–*T* conditions of retrogression of ~300–500 MPa at 250–450 °C, determined by homogenization of H_2O–CO_2 fluid inclusions, are compatible with formation during exhumation of the mid crust.

These examples indicate that evidence for late fluid activity is widespread in the Grenville Province and was locally associated with major shear zones that exhibit normal-sense kinematics and can reasonably be linked to orogenic collapse. However, the origin of the implied large volumes of hot fluid remains problematic. Moreover, it should be noted that, in several cases, the timing of fluid-driven alteration is disputed, with some authors linking it to the pre-Ottawan host granites (with the hydrothermal mineralization being remobilized in syn- to late-Ottawan time) and others favouring formation of the mineralization by fluids channelled along late- to post-Ottawan normal-sense shear zones.

Discussion

Conceptual model of orogenic collapse in the Grenville Orogen

Figure 15 is a conceptual diagram to explain the geometric evolution of tectonic architecture and structural relations during collapse described earlier in the text. The first panel illustrates a cross-section of an orogen with double crustal thickness, beneath which the asthenospheric thermal boundary layer has risen because of convective thinning of the mantle lithosphere, slab break-off, or some other mechanism, leading to uplift of a plateau of width *l* and dramatic rheological weakening of the mid crust (after Dewey 1988).

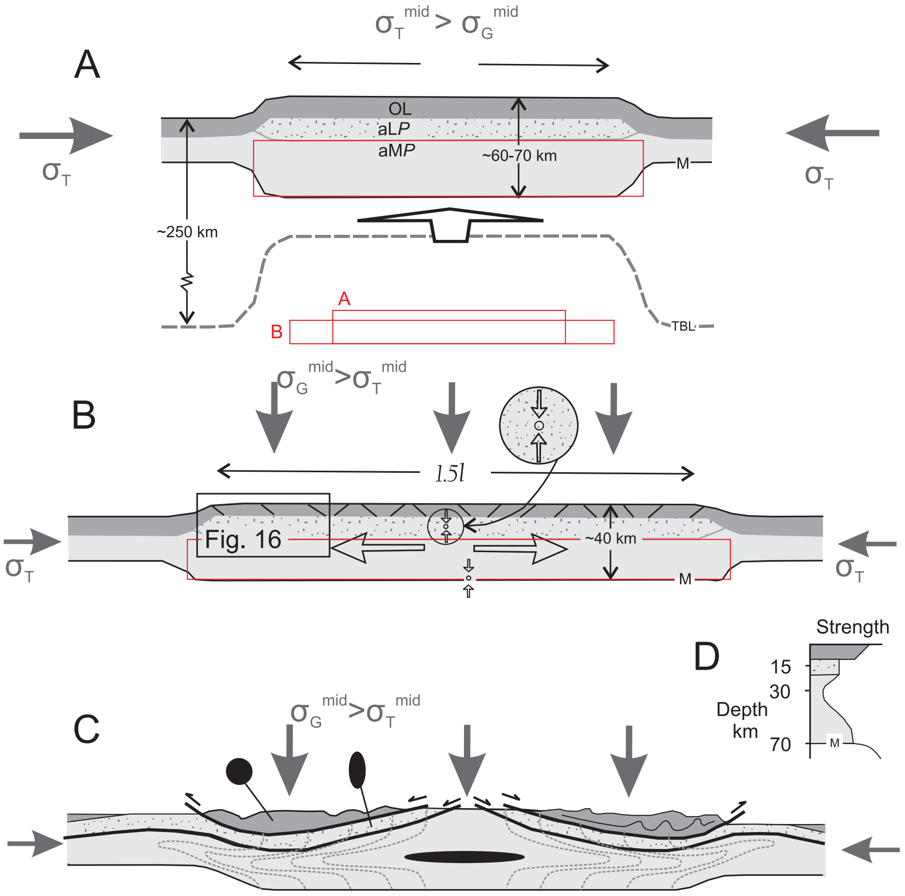
Orogenic collapse is illustrated in Fig. 15B. In this figure, the mid crust is thinned by one third because of subhorizontal flow driven by gravitational loading exceeding its strength (i.e., $\sigma_G^{mid} > \sigma_T^{mid}$), resulting in an overall crustal thickness of ~40 km. Conservation of volume implies the area of the orogen must increase. In the cross-sectional figure, the orogen is shown as 50% wider (width 1.5*l*), a percentage that could be lower if (*i*) volume was not conserved and part of the mid crust was exhumed and eroded at the orogen margin by some form of channel flow and extrusion process (e.g., Beaumont et al. 2006); or (*ii*) the orogen extended horizontally in the third dimension parallel to its margin, as is the case in the Tibetan Plateau (Zhang et al. 2004; Taylor and Yin 2009). While not discounting the possibility of significant mass loss from extrusion and erosion, increase in area is probably inescapable if collapse occurred during regional shortening, as evidence from the Grenville Province suggests. Increase in area of the ductile mid crust must cause a concomitant increase in area of the overlying upper crust, assuming they are not completely decoupled. Given its predominantly brittle rheology, stretching of the uppermost crust is inferred to lead to a horst and graben geometry and eventually to wholesale boudinage and segmentation (e.g., Malavieille 1993; Teyssier and Whitney 2002; Jolivet et al. 2004; Rey et al. 2009). It is this latter process, and the concomitant rise of ductile mid crust into the boudin-neck regions, that signals the initiation of core complex formation.

In the following paragraphs, the case is made for extension of the orogen in the third dimension, i.e., parallel to its orogenic front.

Evidence for a significant component of mid-crustal stretching subparallel to the Grenville Front is shown in Fig. 14, e.g., orientations of displacement vectors for the oblique-normal Carthage–Colton, Tawachiche, and Labelle shear zones noted previously. This is illustrated schematically in 3-D in Fig. 16, which shows mid-crustal stretching leading to boudinage of the brittle upper crust and core complex formation in the section parallel to the orogen margin. On the other hand, the evidence indicates that the L*P* upper crust was undergoing shortening leading to subvertical thickening and the formation of upright structures with axes approximately parallel to the orogenic front at the same time (i.e., $\sigma_T^{upper} > \sigma_G^{upper}$).

Orogen-parallel stretching and thinning of the mid crust can only be reconciled with coeval orogen-parallel thickening

Fig. 15. Schematic diagram showing orogenic cross-sections with proposed mechanisms and regional strain patterns during orogenic collapse. (A) $\sigma_T^{mid} > \sigma_G^{mid}$, rise of asthenospheric thermal boundary layer (TBL) because of convective thinning of mantle lithosphere, slab break-off, or other mechanism causes heating of lower and mid crust and rapid uplift; M, Moho. Wide vertical arrow represents rapid uplift following rise of TBL (modified from Dewey 1988). (B) $\sigma_G^{mid} > \sigma_T^{mid}$, ductile thinning of mid to lower crust by ∼1/3 accomplished by subhorizontal flow; conservation of volume (compare boxes in A and B, red in web version) implies orogen widens by ∼50% if widening occurs in plane of section. A possible component of orogen-parallel strain is indicated by the strain figure (see Fig. 1). Upper crust undergoes brittle faulting. Box shows location of 3-D sketch in Fig. 16. (C) Schematic figure of ductile strain pattern in collapsed orogen. Shapes and orientations of strain ellipses (black) in mid crust (aM*P* Belt), upper crust (aL*P* Belt), and uppermost crust in orogenic lid (OL) are shown. Dashed lines represent flow paths and are approximately parallel to the dominant foliation (modified from Rey et al. 2001). (D) Schematic rheological profile of crust under orogenic plateau showing overall C-shape resulting from relative strengths of the Ottawan Orogenic Lid, aL*P* Belt, and aM*P* Belt.

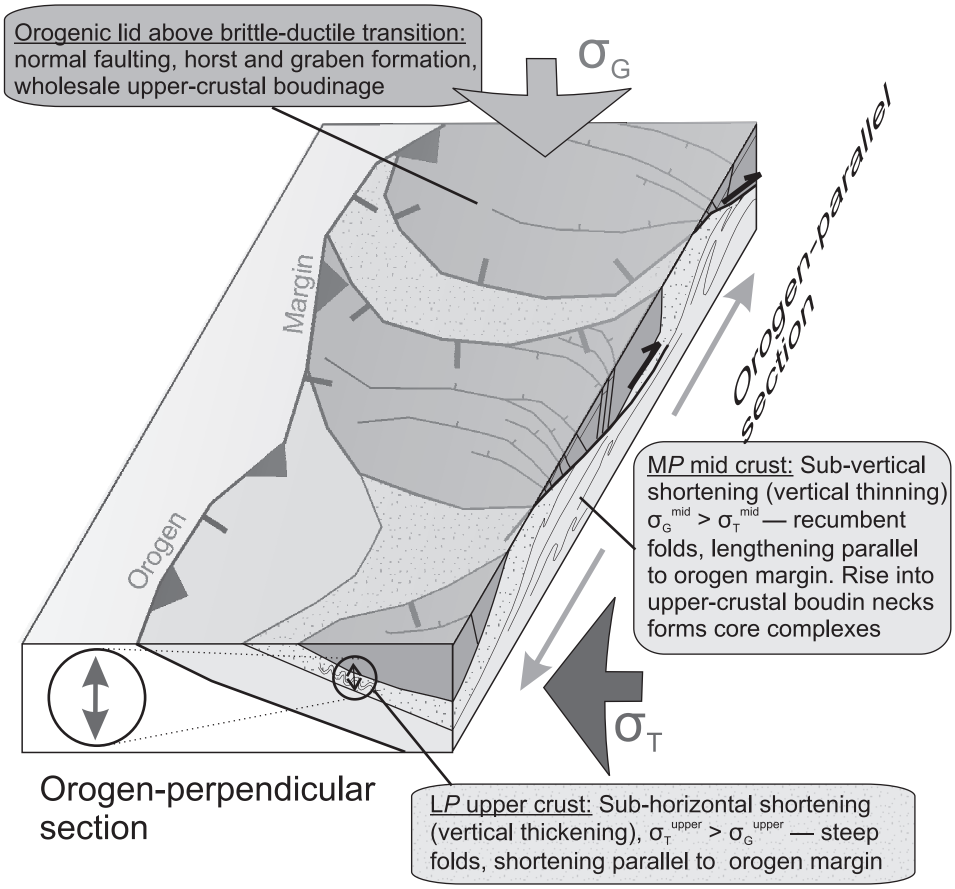


of the boudinaged ductile L*P* upper crust if the two deformation regimes were decoupled. Specifically, as noted, subhorizontal shortening of the upper crust approximately perpendicular to the orogenic front implies $\sigma_T^{upper} > \sigma_G^{upper}$ in the upper crust, whereas flattening and subhorizontal extension of the mid crust parallel to the orogenic front implies $\sigma_G^{mid} > \sigma_T^{mid}$ in the underlying mid crust. Such decoupling has been reported from other orogens (e.g., Axen et al. 1998; Denèle et al. 2009) and may be a common response of doubly thickened crust in a prolonged compressional setting.

Figure 15C schematically illustrates the internal structure of a collapsed orogen and the distribution and orientation of ductile strain (modified from Rey et al. 2001). Note that, in the mid crust, structures are subhorizontal and strains are large; in the upper crust, structures are subvertical and strains are smaller; and in the uppermost crust, structures are principally brittle normal-sense faults and shallow upright folds and penetrative strain is negligable. All of these features are compatible with observations of the collapsed Ottawan hinterland of the Grenville Orogen. This pattern is also consistent with generalized suprastructure–infrastructure models (e.g., Culshaw et al. 2006; Williams et al. 2006). Note that Fig. 15C provides a tectonic context for late emplacement and folding of upper-crustal thrust sheets by gravity sliding off exhumed mid-crustal core complexes, with possible examples in the Grenville Province being the Cabonga Allochthon and Natashquan domain, as discussed previously.

Path lines in Fig. 15C illustrate that mid-crustal strain was driven by gravitational forces acting on the excess potential energy of the plateau ($\sigma_G^{mid} > \sigma_T^{mid}$), whereas upper-crustal strain was predominantly a result of regional shortening ($\sigma_T^{upper} > \sigma_G^{upper}$), as described earlier. This implies a C-shaped crustal strength profile during orogenic collapse (Fig. 15D), with the mid crust being the weakest layer in the middle of

Fig. 16. Detail of Fig. 15 showing schematic 3-D structure of orogenic crust resulting from collapse. Development of structures illustrated in orogen-parallel and orogen-perpendicular sections is coeval. Uppermost brittle crust in orogen-parallel section shows internal horst and graben structure, and wholesale boudinage and segmentation; underlying ductile M*P* mid crust exhibits recumbent folds associated with vertical thinning (flattening) and flow into boudin necks, which develop into core complexes ($\sigma_G^{mid} > \sigma_T^{mid}$; modified from Jolivet et al. 2004). Note excision of L*P* upper crust in part of the section. Orogen-perpendicular section shows coeval shortening and vertical thickening of L*P* upper crust ($\sigma_T^{upper} > \sigma_G^{upper}$). Together, they imply that deformation in the mid and upper crust was decoupled during collapse.



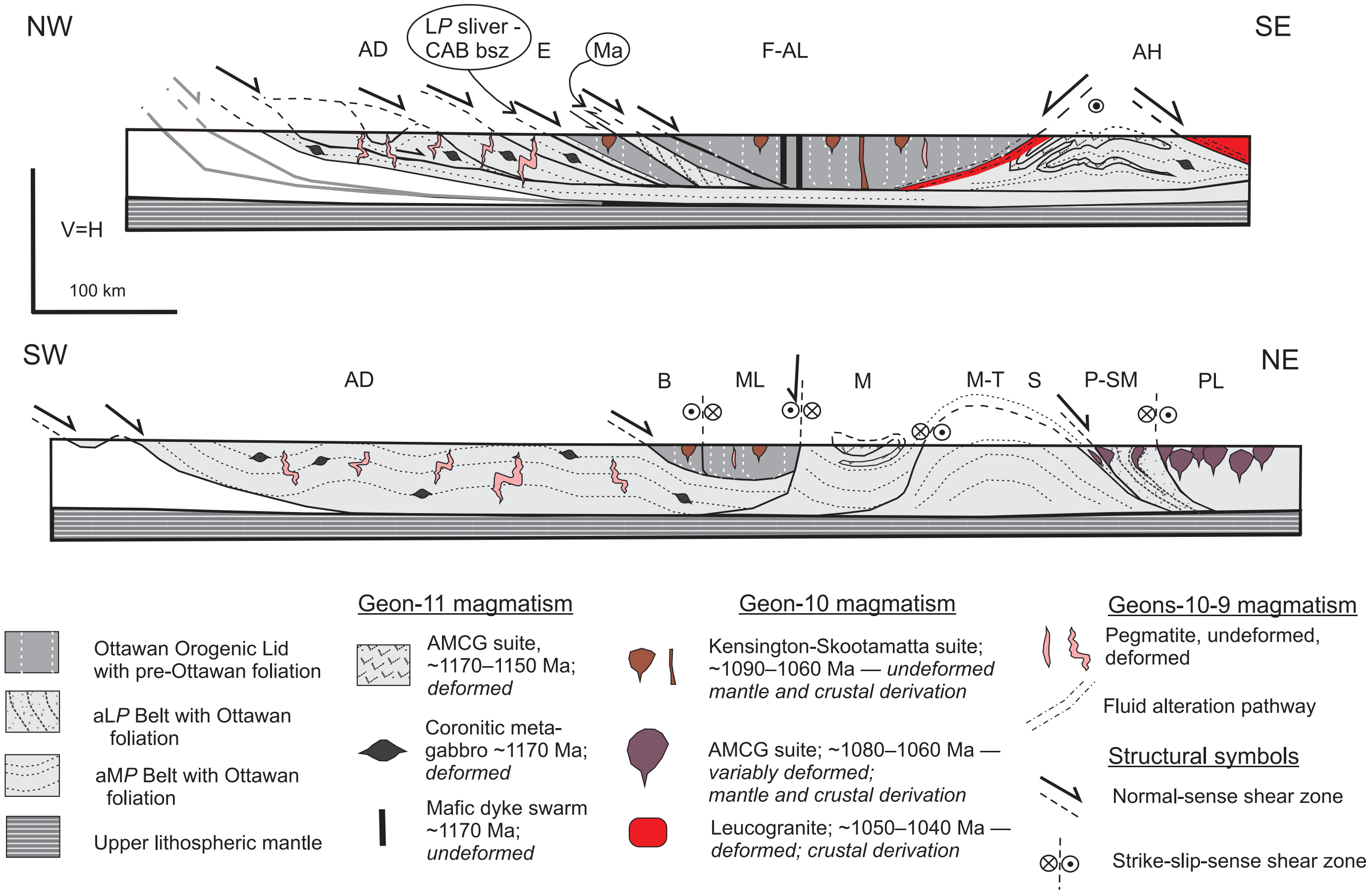
the double crustal sandwich. This rheological profile is compatible with the structural scenario in Fig. 13, in which regional shortening led to crustal thickening, which in turn promoted mid-crustal thinning when the overlying load exceeded the strength of the melt-weakened mid crust.

A final observation of interest is that gneiss domes are not a characteristic feature of the Grenvillian aM*P* Belt, although they occur locally (Gervais et al. 2004). This implies that a scenario in which decompressional melting drove a feedback loop causing diapirism (Teyssier and Whitney 2002; Vanderhaeghe 2009) was not generally applicable. The rarity of gneiss domes is compatible with the inferred exhumation path of the mid crust (path 2a; Fig. 4), which crosses the solidus of the biotite dehydration-melting reaction at relatively elevated pressure (~600–800 MPa), thereby precluding the rise of locally derived melt to shallow depth.

Orthogonal cross-sections of collapsed crust

It was noted in previous sections that the overall architecture of the collapsed crust resembled a series of core complexes. This is explored further in Fig. 17, which shows two cross-sections of the southwestern Grenville Province approximately perpendicular and parallel to the strike of the orogenic front. The NW–SE (perpendicular) section is an enlarged version of the seismic interpretation in Fig. 3 with additional data relating to orientations of important structures, fabrics, and igneous bodies. The physical resemblance of the Adirondack Highlands terrane to a divergent core complex centred on the poly-deformed geon-11 Marcy AMCG massif is striking, as noted by Rivers (2008) and McLelland et al. (2009). Moreover, as pointed out by Selleck et al. (2005) and McLelland et al. (2010) and discussed previously, normal-sense displacement on the Carthage–Colton and Eastern Adirondack shear zones was lubricated by coeval emplacement of leucogranite sheets (~1050 Ma Lyon Mountain granite), a common feature of core complexes (e.g., Wernicke 1985), and the shear zones also served as channel ways for the circulation of hydrothermal fluids (Foose and McLelland 1995). The Algonquin – Lac Dumoine terrane at the northwest end of this cross-section is also interpreted as a core complex, although, in this case, it is asymmetrical and dissected by several important SE-dipping normal-sense shear zones that give it a broader profile. The main, gently dipping to subhorizontal fabrics in both the Adirondack Highlands and Algonquin – Lac Dumoine terranes are of Ottawan age, as schematically indicated by the way they wrap around var-

Fig. 17. Orthogonal, crustal-scale cross-sections illustrating the core-complex architecture of the mid crust in the southwest Grenville Province, and the basin-shaped graben structures occupied by the Ottawan Orogenic Lid. The NW–SE cross-section is constrained by deep-crustal seismic data, whereas the deep structure of the SW–NE section is drawn by hand and constrained at depth only at two locations where it crosses seismic lines. Both cross-sections are schematically ornamented with observed structures and magmatic–hydrothermal features temporally associated with collapse. AD, Algonquin – Lac Dumoine terrane; AH, Adirondack Highlands terrane; B, Bancroft terrane; CAB bsz, Composite Arc Belt boundary shear zone; E, Elzevir terrane; F-AL, Frontenac–Adirondack Lowlands terrane; M, Morin terrane; Ma, Mazinaw terrane; ML, Mont Laurier terrane; M-T, Mékinak–Taureau terrane; PL, Parc des Laurentides terrane; P-SM, Portneuf–St-Maurice domain; S, Shawinigan domain.



iably deformed and partly amphibolitized bodies of ca. 1170 Ma coronitic metagabbro (McLelland and Isachsen 1986; Ketchum and Davidson 2000).

The counterpart to the mid-crustal core complexes is the uppermost crust in the OOL, which is now at the same structural level. The main fabrics in the OOL are steep and pre-Ottawan, their "old" ^{40}Ar/^{39}Ar hornblende ages indicating the temperature in Ottawan time was $\leq$500 °C, and the presence of miarolitic cavities in some geon-10 intrusions testifying to the very high level of preserved crust locally. Note that undeformed units predating the Ottawan orogenic phase (e.g., geon-11 Kingston dykes in the Frontenac domain; Davidson 1998) or coeval with it (e.g., geon-10 Kensington–Skootamatta suite in the Composite Arc and Frontenac–Adirondack belts), are restricted to the OOL. Moreover, the minette dyke in the Kensington–Skootamatta suite, with its abundant crustal and mantle xenoliths, provides a physical link to the character of the underlying crust and mantle that has not been found elsewhere.

With respect to the aL*P* segments in this cross-section, the arrangement of the Mazinaw terrane surrounded by the OOL is reminiscent of the architecture of an inclined asymmetrical horst as discussed previously, its Ottawan isograd pattern indicating it was heated from below and tilted in the third dimension during exhumation (Rivers 2008). This cross-section also shows the thin L*P* segment adjacent to the Ottawan Composite Arc Belt boundary shear zone, which separates the northwestern margin of the OOL from the underlying M*P* Algonquin – Lac Dumoine terrane. The presence of an L*P* segment between the OOL and aM*P* Belt implies that a partial crustal section is preserved at this location.

The SW–NE cross-section is unconstrained by seismic data, except at the locations of two cross-lines. Taking into account the timing of displacements on shear zones from Fig. 14, it shows that the Morin and Mékinak–Taureau terranes are juxtaposed across the transpressional Taureau shear zone that was active at ~1074 Ma and that together they comprise a mid-crustal core complex separating the Mont Laurier domain, part of the OOL, from the L*P* Portneuf–St-Maurice domain. Moreover, the data in Fig. 14 indicate that early Ottawan crustal thickening and transcurrent deformation in the mid crust (i.e., $\sigma_T^{mid} > \sigma_G^{mid}$) overlapped temporally with the oblique-normal displacement on the Tawachiche and Labelle shear zones (at ~1090–1040 and ~1078 Ma, respectively; $\sigma_{\mathrm{G}}^{\mathrm{mid}} > \sigma_{\mathrm{T}}^{\mathrm{mid}}$), which resulted in juxtaposition of the aM*P* Belt and OOL. Note that internal structure in the Algonquin – Lac Dumoine terrane is not well illustrated in this section because it is approximately parallel to strike.

The observation from Fig. 17 that large, core-complex-like structures are evident on two orthogonal cross-sections supports the inferred quasi-radial nature of collapse discussed earlier. On the other hand, the amount of L*P* crust in both cross-sections is much less than anticipated, its apparent absence locally (e.g., between the Mont Laurier and Morin terranes in the SW–NE cross-section) and its extreme thinness elsewhere (e.g., the L*P* sliver along the northern boundary of the OOL in the NW–SE cross-section), raising the possibility that it may have been partially to completely excised during exhumation. Interestingly, Axen et al. (1998) concluded that at least 2 km of crust was excised along a décollement between the mid and upper crust in one of the areas they studied, perhaps suggesting this may not be an uncommon process.

The recumbent folds in geon-9 pegmatites in the aM*P* Belt and the undeformed geon-10 pegmatites in the OOL, both greatly enlarged in Fig. 17 for visibility, are included to illustrate the contrast in late- to post-Ottawan vertical thinning in exhumed ductile mid crust compared with the adjacent OOL. They also serve to alert the reader to the probably much more important vertical thinning in the aM*P* Belt in mid to late Ottawan time, possibly by mechanisms illustrated in Fig. 13, which remains unquantified.

In summary, the presence in two orthogonal cross-sections of normal-sense shear zones separating large footwall domes 100–200 km in diameter, underlain by high-grade M*P* crust, from basinal-shaped hanging-wall terranes, underlain by unmetamorphosed crust of the OOL and slivers of L*P* crust, supports the inference of core-complex architecture. This conclusion is also compatible with the subhorizontal Ottawan fabrics in the M*P* footwall domes and the steep Ottawan fabrics in the structurally overlying L*P* upper crust, their contrasting orientations and metamorphic grade indicating they developed under different, decoupled rheological regimes.

Complex displacement patterns during orogenic collapse

Figures 14, 16, 17 suggest that deformation patterns during orogenic collapse were complex and not readily imaged in 2-D. In addition to differences in directions of finite strain, timing, and kinematics, shear zones occur not only along the boundaries of aM*P*, aL*P*, and orogenic lid segments, but also are entirely enclosed within them. Moreover, as discussed in a previous section, vertical thinning took place within the imbricated thrust sheets, as well as along shear zones. The data relating to the northern segment of the OOL shown in Fig. 14 serve as an example of the degree of complexity that may be typical. In this figure, the Labelle shear zone, which separates the Morin and Mont Laurier terranes, was the site of oblique-normal displacement at ~1078 Ma. On the other hand, the subparallel Baskatong shear zone, which occurs entirely within the Mont Laurier terrane, exhibits evidence for strike-slip at ~1020 Ma. Moreover, the Baskatong shear zone can be traced into the Cabonga thrust, on which thrust-sense displacement took place at ~998 Ma, and the Cabonga Allochthon in its hanging wall overlies two strike-slip shear zones in the underlying mid crust on which displacement occurred at ~998–980 Ma. Attempts to integrate such complex and undoubtedly simplified data sets into formal models are probably doomed to failure. However, such complexities in detail do not detract from the overall pattern of a hanging wall composed of the OOL and aL*P* Belt separated by a system of normal-sense shear zones from a footwall composed of the aM*P* Belt.

Transition from crustal thickening to gravitationally driven crustal thinning

It is appropriate to briefly consider the rheological implications of the transition from crustal thickening – plateau formation to gravitationally driven crustal thinning. It has been argued elsewhere that crustal thickening in the Grenville Orogen probably involved some form of channel flow or ductile nappe transport (e.g., Jamieson et al. 2007, 2010; Rivers 2008, 2009), with the base of the channel being the Allochthon Boundary Thrust. Such models imply that the mid crust was decoupled from the upper crust. On the other hand, as discussed earlier, the preferred model for orogenic collapse involves ductile spreading of the mid crust driving stretching of the overlying upper crust. This implies some degree of coupling of the mid and upper crust at this stage. Hence, a critical step between the thickening and collapse stages is inferred to have been the physical re-attachment of the upper and mid crust. The most likely driver for such a process is a reduction in plate convergence rates, leading to a decline in σ_{T}. Once $\sigma_{\mathrm{T}}^{\mathrm{mid}} < \sigma_{\mathrm{G}}^{\mathrm{mid}}$ in the mid crust, thickening at this crustal level would cease, to be replaced by cooling and gravitational spreading. Subsequent cooling of the mid crust would cause it to stiffen, thereby enhancing rheological linkage with the overlying upper crust, with the result that gravitationally driven flow of the ductile mid crust led to stretching of the overlying carapace, as shown in Figs. 15, 16. Thus, the overall result was collapse of the upper crust into the former mid-crustal channel, as schematically illustrated by Rivers (2008).

Transition from gravitationally driven crustal thinning to regional extension

Most normal-sense shear zones associated with collapse sole in the Ottawan mid crust (Fig. 3), implying that the increase in length of the hanging wall was limited to the mid and upper crust. In contrast, a normal-sense shear zone in the Parautochthonous Belt known as the Dorval detachment displaces the Moho (cross-section C, Fig. 3), implying that displacement on it lengthened all the crust and part of the uppermost lithospheric mantle relative to the underlying footwall. Available evidence loosely brackets displacement on the Dorval detachment between ~1020 and 960 Ma (Martignole and Martelat 2005), indicating it was active during some stage of orogenic collapse. Moreover, as pointed out by Hynes and Rivers (2010), this structure provides the first evidence for regional, whole-crustal extension as opposed to lengthening of the mid and upper crust in an overall compressional orogen. As such, it may be related to a second episode of slab delamination and rollback (e.g., Vanderhaeghe 2009) or alternatively to regional tensional forces external to the Grenville Orogen caused by plate motions elsewhere on the globe. If the latter hypothesis is correct, it implies the Grenville Orogen preserves evidence for an evolution from externally driven crustal thickening, through internally driven orogenic collapse, to externally driven thinning over a period of ~150 Ma.

Significance for seismic interpretations and numerical modelling

The conclusions of this paper have relevance for both seismic interpretations and numerical modelling of collapsed orogens. Firstly, it is appropriate to address the point that the published interpretations of Lithoprobe and COCORP deep seismic data missed or downplayed the collapsed character of the Grenville Orogen. This was not primarily a problem of seismology — many of the critical reflectors were imaged or their existence could be inferred — but rather because of a lack of information concerning the scale of displacement on them. Although the concept of orogenic collapse had already been proposed to explain the distribution of ^{40}Ar/^{39}Ar data (e.g., Cosca et al. 1995), and evidence for double crustal thickness was available (e.g., Indares et al. 1998), links between the two were not made. However, assessment of the amount of displacement on normal-sense shear zones could not be made until the mid, upper, and uppermost levels of Ottawan crust had been distinguished (Rivers 2008). Thus, initial seismic interpretations were made in a context of incomplete information and reflect the piecemeal geological understanding at the time. Perhaps even more important was the lack of conceptual understanding of the scale and pervasiveness of crustal thinning during orogenic collapse. In hindsight, Dewey's (1988) paper appears remarkably prescient.

A second related issue concerns the scale of crustal thinning during orogenic collapse. The estimated ~30% thinning of the mid crust in this study, although imprecise, nonetheless implies that the seismic structure imaged in the mid to lower crust of collapsed orogens should be seen through a prism of substantial vertical flattening. This insight should not only lead to more refined seismic interpretations, for instance relating to the original dips of ramps, but may also provide an opportunity for creative seismologists to retro-deform the crust back to its inferred maximum orogenic thickness and, thus, provide constraints on the internal architecture before collapse. The key point is that the imaged mid-crustal structure likely principally reflects the geometry of collapse rather than that at peak thickness, although, as noted, some shear zones may have been active at both stages.

Concerning numerical modelling of large hot orogens and of the Grenville Orogen in particular, the interpretations in this paper suggest that neither the timing of orogenic collapse nor the significant crustal thinning and concomitant increase in area resulting from collapse has yet been completely captured. In a recent model designed to simulate the Grenville Orogen (Jamieson et al. 2010), collapse of the orogenic plateau was initiated after 50–70 Ma of collision, which is incompatible with the temporal progression of deformation and metamorphism determined from several areas in this study. Perhaps because crustal thickening and collapse were treated as two independent and sequential processes, the models do not appear to reproduce the observed temporal sequence of high-grade metamorphism in the mid crust, followed by medium-grade metamorphism in the upper crust, followed by heating of the margins of the orogenic lid.

Finally, and most importantly, a critical limiting factor relating to both the seismic interpretations and numerical modelling is their restriction to two dimensions. The interpretation of collapse presented here, with decoupled strain regimes at different crustal levels extending the crust in different directions, indicates it is fundamentally a 3-D process, an inference corroborated by empirical evidence from the Tibetan Plateau and other areas. As such, it is not reasonable to expect that 2-D transects and plane strain models will realistically capture all aspects of the process.

Application of orogenic collapse model to Grenvillian inliers

Much of the crust in the discontinuous line of Grenvillian inliers that extends from the western Appalachians to Texas exhibits high-grade Ottawan metamorphism (e.g., Tollo et al. 2004; McLelland et al. 2010; Rivers et al. in press), and available data suggest it comprises part of the a*MP* Belt, implying exhumation from mid-crustal depth. Two exceptions are discussed here.

Firstly, Ottawan (~1090–1030 Ma) metamorphic assemblages in the New Jersey Highlands are uppermost amphibolite to hornblende-granulite facies (Volkert 2004), with peak *P–T* conditions of 400–600 MPa and 670–780 MPa. These estimates are compatible with the a*LP* Belt as defined by Rivers (2008) and herein. Undeformed post-tectonic intrusions include the ~1020 Mount Eve leucogranite, which exhibits a thin contact aureole implying emplacement into cool crust, and crosscutting pegmatite and related silicate–borate skarn with crystallization ages of ~1004 and ~990 Ma. Volkert et al. (2005, p. 1) inferred the locations of the intrusions and hydrothermal fluids were controlled by "…lithosphere-scale fault zones … that were undergoing extension and that [their] emplacement followed orogenic collapse by at least 30 Ma". This interpretation accords with the proposed model, and it is noteworthy that the estimated *P–T* conditions and timing of Ottawan metamorphism are comparable to data from the L*P* Portneuf–St-Maurice domain, and that the duration of late hydrothermal activity (from ~990 to 940 Ma) overlaps with that in the Adirondack Highlands terrane.

The second case is the Llano Uplift in central Texas. This area is notable for its lack of Ottawan regional metamorphism, as most visibly recorded by a widespread suite of large, undeformed to weakly deformed, ~1090 Ma granitoid plutons. The plutons are surrounded by low-pressure metamorphic aureoles (~300 MPa and 525–625 °C), implying emplacement into cool upper crust, and a weak fabric associated with the formation of late cross-folds affects only the older members of the suite (Mosher et al. 2008). Thus, they are of similar age to, and share a "post-tectonic" character with, the Kensington–Skootamatta suite. Moreover, they too were derived from both mantle and crustal sources, although the two suites are geochemically distinct. The undeformed to weakly deformed state and high crustal level of the Llano plutons implies the Llano Uplift was not affected by penetrative ductile strain and high-grade metamorphism in Ottawan time, an interpretation supported by three ^{40}Ar/^{39}Ar hornblende apparent ages between ~1098 and 1076 Ma and two U–Pb titanite ages of ~1090 Ma (Rougvie et al. 1999). Integration of all these data leads to the proposal that the entire Llano Uplift may form a segment of the OOL. Unfortunately, there are no data to constrain the timing of displacement on normal-sense shear zones in the Llano Uplift to test this hypothesis, although according to Mosher et al. (2004, 2008) extension occurred several times, with the latest episode taking place under predominantly brittle conditions. This interpretation that the entire Llano Uplift makes up part of the OOL is not incompatible with independent data indicating that compressional deformation at the Grenville Front in west Texas took place in Rigolet time (~1.0 Ga; Grimes and Copeland 2004). In fact, taken together the pattern is remarkably similar to that in the Grenville Province proper, supporting the inference that the lack of Ottawan metamorphism is a result of collapse leading to the preservation of the uppermost crust, rather than because of a fundamentally different tectonic evolution. If correct, this implies that plateau construction and collapse may have occurred along the ~5000 km length of the Grenville Orogen in North America.

The long-term signature of orogenic collapse on crustal strength

One of the conclusions of Dewey's (1988) paper was that the crust was so pervasively weakened during orogenic collapse that it was liable to be reworked as a site of rifting in later tectonic events — an inference that provides a rheological explanation for repeated orogeny along long-lived plate boundaries. Although he mostly drew on European examples to make his case, this concept clearly has potential application to the Grenville Province, part of which later became the site of early Paleozoic rifting and overprinting during the Appalachian Orogeny, followed ~200 Ma later by renewed rifting and formation of the modern Atlantic margin. Specifically, there is abundant evidence that the Cambrian Appalachian rift basin was floored by stretched Grenville basement, and in New England the Triassic rift basins that signal the break-up of Pangea are floored by high-grade rocks of the Appalachian Orogen, which in turn are underlain by thinned Grenville basement.

Following the principle outlined by Dewey (1988), this cycle of repeated break-up and orogeny in approximately the same place may have been initiated by collapse of the Grenville Orogen. Transcrustal, normal-sense structures, such as the Dorval detachment, are compatible with this concept and suggest the crust and upper mantle in the Grenville Orogen were thoroughly weakened by the collapse process, and moreover, that the weakness endured at least 750 Ma until the beginning of the Mesozoic.

Evaluation of Dewey's (1988) model for orogenic collapse

The data presented in this paper clearly provide general support for Dewey's (1988) empirical model of orogenic collapse, although the temperature of the Grenvillian mid crust was probably higher, and the style of deformation in the mid-crustal core complexes correspondingly more ductile, than in the examples he described. This suggests that the amount of vertical thinning of the mid crust was probably greater. An important result of this greater ductility is that the Grenvillian upper crust is inferred to have collapsed into the underlying mid-crustal channel.

A second difference is that the empirical model of Dewey (1988), as well as the numerical models of Rey et al. (2001), Vanderhaeghe and Teyssier (2001), and Jamieson et al. (2010), all considered collapse as an essentially 2-D process, whereas the evidence presented here, supported by data from the Tibetan Plateau and elsewhere, implies that it is fundamentally 3-D. This is an important difference as it provides a context for the decoupling of coeval subhorizontal structures in the mid crust from subvertical structures in the upper crust, which may be a fundamental aspect of the collapse process.

Nonetheless, despite these differences, the overall geometric similarity between the Grenville Orogen as described herein and the Mesozoic orogens considered by Dewey (1988) is compelling, leading to the conclusion that the first-order architecture of Mesoproterozoic crust in the Grenville Orogen was essentially similar to that in Mesozoic large hot orogens. Thus, there is no indication of a fundamental change in lithospheric rheology since the Mesoproterozoic. It is interesting to debate, however, whether this conclusion can also be extended back into the Paleoproterozoic. For instance, the Trans-Hudson Orogen, although indubitably large, hot, and of long-duration, has not yielded evidence for the formation of an orogenic plateau or for orogenic collapse (e.g., Corrigan et al. 2009). This may suggest that orogenic plateaux are a Mesoproterozoic and later feature of planet Earth that resulted from gradual strengthening of the lithosphere as it cooled and thickened, leading to a visco-elastic rheological response on doubling of crustal thickness. Such a hypothesis is supported by the recent recognition of "ultra-hot orogens" in the Paleoproterozoic (Chardon et al. 2009), which exhibit important contrasts with the Grenville Orogen. If this inference is correct, the Grenville Orogen may have been one of the first mountain belts to develop a large plateau of Himalayan scale in its hinterland, as well as one of the largest, and as such it may constitute a critical marker for secular change of the rheological properties of the lithosphere.

Concerning other modifications to Dewey's (1988) model resulting from interpretations of the "mature" (i.e., completely collapsed) Grenville Orogen discussed herein compared to the "immature" orogens he examined, it is pertinent to note the temporal progression of metamorphism from the mid to the upper crust driven by conductive heat transfer between the crustal levels as they came in to contact. In addition, the extent of preserved uppermost crust is significant, which may be related to, and a cryptic indicator of, the large amount of thinning (≥30%?) of the mid crust.

With regard to associated igneous processes, the involvement of mixed mantle and crustal sources in some orogenic intrusions, and the concomitant wide range of magmatic compositions, is significant, as is their possible role in advecting magmatic heat into the mid crust. Both point to crust–mantle feedback in orogeny. Other significant factors are (1) the role of leucogranite in lubricating normal-sense shear zones and (2) the role of pegmatites in rearranging the distribution of heat-producing elements in the crust. Finally, the interpretation that collapse triggered the circulation of hydrothermal fluids is also significant, not only in terms of the source of the fluids and the economic potential of the resultant deposits, but also because it appears to represent a widespread response of the crust as fluids gained access to its exhumed hot deeper levels.

Conclusions

This paper integrates a wide range of information, including structural, metamorphic, and geochronologic data, $^{40}Ar/^{39}Ar$ apparent ages, and aspects of late- and post-orogenic magmatism and hydrothermal activity, to propose the first comprehensive model for collapse of the Ottawan hinterland of the

Grenville Orogen. Underlying the model is a tripartite rheological subdivision of Ottawan orogenic crust into (1) a hot, rheologically weak, mid crust with subhorizontal structures and peak *P–T* conditions of ~1000 ± 100 MPa and 800 ± 50 °C; (2) a cooler, rheologically stronger, upper crust with subvertical structures and peak *P–T* conditions of ~350–700 MPa and 450–750 °C; and (3) a "cold" (≤500 °C), strong, uppermost orogenic lid with pre-Ottawan structures and metamorphism. It is argued that the pre-eminent features of the crustal architecture in the orogenic hinterland can be interpreted in terms of a series of mid-crustal, M*P* core complexes up to a few hundred kilometres in width, above which large segments of the upper and uppermost orogenic crust consisting of the L*P* Belt and orogenic lid were displaced in a quasi-radial pattern along normal or normal-oblique shear zones. The present architecture, in which the mid crust of the M*P* Belt is preserved at the same level as the uppermost crust of the orogenic lid, suggests collapse involved ~30% vertical thinning and significant orogen-parallel lengthening. The temporal and spatial progression of Ottawan metamorphism over a period of ~100 million years, from granulite facies in the mid crust, through amphibolite facies in the upper crust, to heating to ≤500 °C in the uppermost crust, is interpreted in terms of progressive exhumation of the hot mid crust and its juxtaposition against successively higher crustal levels, which experienced heating by conduction as a result. Thinning of the mid crust was accommodated by a combination of simple shear and pure shear mechanisms and was driven by the excess potential energy of the plateau. It was in part coeval with the development of steep to upright ductile structures that caused thickening in the overlying upper crust, implying that deformation of the mid and upper crust was decoupled at that time. Finally, normal-sense shear zones active during collapse were the sites of emplacement of a wide range of magmatic products and the channelling of hydrothermal fluids, all of which may have reduced the viscosity of the shear zones and facilitated displacement on them, contributing to the profound re-organization of the crustal architecture.

Acknowledgements

It is a pleasure to submit this paper to a volume celebrating the many and varied contributions of Ward Neale to the Canadian geoscience community. Ward always enjoyed a discussion and I trust he would have found it thought-provoking. The paper owes an obvious debt to John Dewey's perceptive paper on orogenic collapse published over 20 years ago, which provided the conceptual framework and many of the answers before those of us working in the Grenville Province had even posed the relevant questions. I thank Fried Schwerdtner for many discussions and a very useful informal review that helped clarify my thinking. It is a pleasure to acknowledge journal reviewers Jim McLelland, for detailed and informative comments and for pointing out relevant literature I had missed, and John Dewey for encouragement and grasp of the big picture. I thank Jim Hibbard, Jeff Pollock, and Hank Williams for organizing this special issue. Sadly Hank didn't live to see it go to press. Also thanks to Associate Editor Brendan Murphy who handled the manuscript. My research in the Grenville Province has been supported by the Natural Sciences and Engineering Research Council for many years.

References

Anderson, S.L. 1988. Interpretation of K–Ar mineral ages from the Grenville orogenic belt. American Journal of Science, **288**(7): 701–734. doi:10.2475/ajs.288.7.701.

Annovitz, L.M., and Essene, E.J. 1990. Thermobarometry and pressure–temperature paths in the Grenville Province of Ontario. Journal of Petrology, **31**: 197–241.

Axen, G.J., Selverstone, J., Byrne, T., and Fletcher, J.M. 1998. If the strong crust leads, will the weak crust follow? GSA Today, **8**(12): 1–8.

Baird, G.B., and MacDonald, W.D. 2004. Deformation of the Diana syenite and Carthage–Colton mylonite zone: Implications for timing of Adirondack Lowlands deformation. *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, Boulder, Colo., Memoir 197. pp. 285–297.

Baird, G.B., Shrady, C., and Jones, J. 2008. Preliminary constraints on the timing of deformation in the northwest Adirondack Lowlands and along the Black Lake shear zone. Geological Association of Canada – Mineralogical Association of Canada (GAC–MAC), Vol. 33, No. 12, Abstract. http://www.gac.ca/activities/abstracts/2973_abst...pdf

Beaumont, C., Jamieson, R.A., Nguyen, M.H., and Medvedev, S. 2004. Crustal channel flows: 1. Numerical models with applications to the tectonics of the Himalayan–Tibetan orogen. Journal of Geophysical Research, **109**(B6): B06406. doi:10.1029/2003JB002809.

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. pp. 91–145.

Berger, G.W., and York, D. 1981. ^{40}Ar/^{39}Ar dating of the Thanet gabbro, Ontario: looking through the Grenvillian metamorphic veil and implications for paleomagnetism. Canadian Journal of Earth Sciences, **18**(2): 266–273. doi:10.1139/e81-020.

Bernier, L.R., and MacLean, W.H. 1993. Lithogeochemistry of a metamorphosed VMS alteration zone at Montauban, Grenville Province, Quebec. Exploration and Mining Geology, **2**: 367–386.

Bickford, M.E., McLelland, J.M., Selleck, B.W., Hill, B.M., and Heumann, M.J. 2008. Timing of anatexis in the eastern Adirondack Highlands: Implications for tectonic evolution during ca. 1050 Ma Ottawan orogenesis. Geological Society of America Bulletin, **120**(7–8): 950–961. doi:10.1130/B26309.1.

Bird, P. 1979. Continental delamination and the Colorado Plateau. Geophysical Research, **83**: 7561–7571.

Bonamici, C., Valley, J., Kozdon, R., and Ushikubo, T. 2010. Linking the intra-grain oxygen isotope record to the *P–T* deformation history of Grenville-age shear zones, Adirondack Mountains, New York, USA. GeoCanada 2010, Geological Association of Canada, Abstract.

Bonnet, A.-L., Corriveau, L., and La Flèche, M.R. 2005. Chemical imprint of highly metamorphosed volcanic-hosted hydrothermal alterations in the La Romaine supracrustal belt, eastern Grenville Province, Quebec. Canadian Journal of Earth Sciences, **42**(10): 1783–1814. doi:10.1139/e05-098.

Brown, C.E. 1983. Mineralization, mining, and mineral resources in the Beaver Creek area, St. Lawrence County, NY. US. Geological Survey, Paper 1279.

Burg, J.-P., and Gerya, T.V. 2005. The role of viscous heating in Barrovian metamorphism of collisional orogens: thermomechanical models and application to the Lepontine Dome in the Central Alps. Journal of Metamorphic Geology, **23**(2): 75–95. doi:10.1111/j.1525-1314.2005.00563.x.

Busch, J.P., and van der Pluijm, B.A. 1996. Late orogenic, plastic to brittle extension along the Robertson Lake shear zone: implications for the style of deep-crustal extension in the Grenville orogen, Canada. Precambrian Research, **77**(1–2): 41–57. doi:10.1016/0301-9268(95)00044-5.

Busch, J.P., van der Pluijm, B.A., Hall, C.M., and Essene, E.J. 1996*a*. Listric normal faulting during postorogenic extension revealed by ^{40}Ar/^{39}Ar thermochronology near the Robertson Lake shear zone, Grenville orogen, Canada. Tectonics, **15**(2): 387–402. doi:10.1029/95TC03501.

Busch, J.P., Essene, E.J., and van der Pluijm, B.A. 1996*b*. Evolution of deep-crustal normal faults: Constraints from thermobarometry in the Grenville orogen, Ontario, Canada. Tectonophysics, **265**(1–2): 83–100. doi:10.1016/S0040-1951(96)00147-3.

Busch, J.P., Mezger, K., and van der Pluijm, B.A. 1997. Suturing and extensional reactivation in the Grenville orogen. Geology, **25**(6): 507–510. doi:10.1130/0091-7613(1997)025<0507:SAERIT>2.3.CO;2.

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 zircon geochronology. Canadian Journal of Earth Sciences, **32**(5): 660–671. doi:10.1139/e95-055.

Carl, J.D. 1988. Popple Hill Gneiss as dacite volcanic: A geochemical study of mesosome and leucosome, northwest Adirondacks, New York. Geological Society of America Bulletin, **100**(6): 970–992. doi:10.1130/0016-7606(1988)100<0841:PHGADV>2.3.CO;2.

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.

Chardon, D., Gapais, D., and Cagnard, F. 2009. Flow of ultra-hot orogens: A view from the Precambrian, clues for the Phanerozoic. Tectonophysics, **477**: 105–118.

Cherniak, D.J. 1993. Lead diffusion in titanite and preliminary results on the effects of radiation damage on Pb transport. Chemical Geology, **110**(1–3): 177–194. doi:10.1016/0009-2541(93)90253-F.

Cherniak, D.J. 2000. Pb diffusion in rutile. Contributions to Mineralogy and Petrology, **139**(2): 198–207. doi:10.1007/PL00007671.

Clark, M.K., and Royden, L.H. 2000. Topographic ooze: Building the eastern margin of Tibet by lower crustal flow. Geology, **28**(8): 703–706. doi:10.1130/0091-7613(2000)28<703:TOBTEM>2.0.CO;2.

Clark, T., Gobeil, A., and David, J. 2005. Iron oxide – copper – gold-type and related deposits in the Manitou Lake area, eastern Grenville Province, Quebec: variations in setting, composition, and style. Canadian Journal of Earth Sciences, **42**(10): 1829–1847. doi:10.1139/E05-048.

Corfu, F., and Easton, R.M. 1995. U–Pb geochronology of the Mazinaw terrane, an imbricate segment of the Central Metasedimentary Belt, Grenville Province, Ontario. Canadian Journal of Earth Sciences, **32**(7): 959–976. doi:10.1139/e95-081.

Corrigan, D., and Hanmer, S. 1997. Anorthosites and related granitoids in the Grenville orogen: A product of convective thinning of the Lithosphere? Geology, **25**: 61–64.

Corrigan, D., and van Breemen, O. 1997. U–Pb age constraints for the lithotectonic evolution of the Grenville Province along the Mauricie transect, Québec. Canadian Journal of Earth Sciences, **34**(3): 299–316. doi:10.1139/e17-027.

Corrigan, D., Pehrsson, S., Wodicka, N., and DeKemp, E. 2009. The Palaeoproterozoic Trans-Hudson Orogen; a prototype of modern accretionary processes. *In* Ancient orogens and modern analogues. *Edited by* J.B. Murphy, J.D. Keppie, and A. Hynes. Geological Society (of London), Special Publication 327. pp. 457–479. doi:10.1144/SP327.19.

Corriveau, L. 1990. Proterozoic subduction and terrane amalgamation in the southwestern Grenville Province, Canada: Evidence from ultrapotassic to shoshonitic plutonism. Geology, **18**(7): 614–617. doi:10.1130/0091-7613(1990)018<0614:PSATAI>2.3.CO;2.

Corriveau, L., and Gorton, M.P. 1993. Co-existing K-rich alkaline and shoshonitic magmatism of arc affinities in the Proterozoic: a reassessment of syenitic stocks in the southwestern Grenville Province. Contributions to Mineralogy and Petrology, **113**(2): 262–279. doi:10.1007/BF00283233.

Corriveau, L., and Morin, D. 2000. Modelling 3D architecture of western Grenville from surface geology, xenoliths, styles of magma emplacement, and Lithoprobe reflectors. Canadian Journal of Earth Sciences, **37**: 235–251.

Corriveau, L., and Mumin, H., editors. 2010. Exploring for iron oxide copper–gold deposits: Canada and global analogues. Geological Association of Canada, Short Course Notes 20. 185 p.

Corriveau, L., Heaman, L.M., Marcantonio, F., and van Breemen, O. 1990. 1.1 Ga K-rich alkaline plutonism in the SW Grenville Province, U–Pb constraints for the timing of subduction-related magmatism. Contributions to Mineralogy and Petrology, **105**(4): 473–485. doi:10.1007/BF00286834.

Cosca, M.A., Sutter, J.F., and Essene, E.J. 1991. Cooling and inferred uplift/erosion history of the Grenville orogen, Ontario: constraints from ^{40}Ar/^{39}Ar thermochronology. Tectonics, **10**(5): 959–977. doi:10.1029/91TC00859.

Cosca, M.A., Essene, E.J., Kunk, M.J., and Sutter, J.F. 1992. Differential unroofing within the Central Metasedimentary Belt of the Grenville Orogen: Constraints from ^{40}Ar/^{39}Ar geochronology. Contributions to Mineralogy and Petrology, **110**(2–3): 211–225. doi:10.1007/BF00310739.

Cosca, M.A., Essene, E.J., Mezger, K., and van der Pluijm, B.A. 1995. Constraints on the duration of tectonic processes: Protracted extension and deep-crustal rotation in the Grenville orogen. Geology, **23**(4): 361–364. doi:10.1130/0091-7613(1995)023<0361:COTDOT>2.3.CO;2.

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**(1): 160–175. doi:10.1139/e94-013.

Culshaw, N.G., Jamieson, R.A., Ketchum, J.W.F., Wodicka, N., Corrigan, D., and Reynolds, P.H. 1997. Transect across the northwestern Grenville orogen, Georgian Bay, Ontario: Polystage convergence and extension in the lower orogenic crust. Tectonics, **16**(6): 966–982. doi:10.1029/97TC02285.

Culshaw, N.G., Beaumont, C., and Jamieson, R.A. 2006. The orogenic superstructure–infrastructure concept: Revisited, quantified, and revived. Geology, **34**(9): 733–736. doi:10.1130/G22793.1.

Culshaw, N., Gerbi, C., and Marsh, J. 2010. Softening 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**(5): TC5013. doi:10.1029/2009TC002537.

Cureton, J.S., van der Pluijm, B.A., and Essene, E.J. 1997. Nature of Elzevir–Mazinaw domain boundary, Grenville Orogen, Ontario. Canadian Journal of Earth Sciences, **34**(7): 976–991. doi:10.1139/e17-081.

Dahl, P.S., Pomfrey, M.E., and Foland, K.A. 2004. Slow cooling and apparent tilting of the Adirondack Lowlands, Grenville Province, New York, based on ^{40}Ar/^{39}Ar ages. *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, Boulder, Colo., Memoir 197. pp. 299–323.

Darling, R.S., Florence, F.P., Lester, G.W., and Whitney, P.R. 2004.

Petrogenesis of prismatine-bearing metapelitic gneiss along the Moose River, west-central Adirondacks, 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, Boulder, Colo., Memoir 197. pp. 325–336.

Davidson, A. 1998, An overview of Grenville Province geology, *In* Geology of the Precambrian Superior and Grenville provinces and Precambrian fossils of North America. *Edited by* S. Lucas and M. R. St-Onge. Geological Society of America, Geology of North America, Vol. C-1. pp. 205–270.

Davidson, A. 2001. Evidence for and significance of high-level plutonism in the southeast Composite Arc Belt, Grenville Province, Ontario. *In* Geological Association of Canada – Mineralogical Association of Canada (GAC–MAC) annual meeting, St. John’s, NL, Abstract Vol. 26, p. A34.

Davidson, D.M., Jr. 1983. Strain analysis of deformed granitic rocks (Helikian), Muskoka district, Ontario. Journal of Structural Geology, **5**(2): 181–195. doi:10.1016/0191-8141(83)90043-3.

Davis, D.W., and Bartlett, J.R. 1988. Geochronology of the Belmont Lake metavolcanic complex and implications for crustal development in the Central Metasedimentary Belt, Grenville Province, Ontario. Canadian Journal of Earth Sciences, **25**(11): 1751–1759. doi:10.1139/e88-166.

DeCelles, P.G., Robinson, D.M., and Zandt, G. 2002. Implications of shortening in the Himalayan fold-thrust belt for uplift of the Tibetan Plateau. Tectonics, **21**(6): 1062. doi:10.1029/2001TC001322.

Denèle, Y., Olivier, P., Gleizes, G., and Barbey, P. 2009. Decoupling between the middle and upper crust during transpression-related lateral flow: Variscan evolution of the Aston gneiss dome (Pyrenees, France). Tectonophysics, **477**(3–4): 244–261. doi:10.1016/j.tecto.2009.04.033.

Dewey, J.F. 1988. Extensional collapse of orogens. Tectonics, **7**(6): 1123–1139. doi:10.1029/TC007i006p01123.

Dunning, G., and Indares, A. 2010. New insights on the 1.7–1.0 Ga crustal evolution of the central Grenville Province from the Manicouagan – Baie Comeau transect. Precambrian Research, **180** (3–4): 204–226. doi:10.1016/j.precamres.2010.04.005.

Easton, R.M. 1986, Geochronology of the Grenville Province. *In* The Grenville Province. *Edited by* J.M. Moore, A. Davidson, and A.J. Baer. Geological Association of Canada, Special Paper 31. pp. 127–173.

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 Vol. 4. pp. 715–904.

Easton, R.M., and Kamo, S.L. 2008. New U–Pb zircon ages reveal a long-lived magmatic history for the Harvey–Cardiff domain of the Composite Arc Belt of the Grenville Province in Ontario. Geological Society of America, Abstracts with Program, **40**(6): 288.

England, P.C., Houseman, G.A., Osmaston, M.F., and Ghosh, S. 1988. The mechanics of the Tibetan Plateau. Philosophical Transactions of the Royal Society of London, Series A, **326** (1589): 301–320. doi:10.1098/rsta.1988.0089.

Evans, D.A.D. 2009. The palaeomagnetically viable, long-lived and all-inclusive Rodinia supercontinent reconstruction. *In* Ancient orogens and modern analogues. *Edited by* J.B. Murphy, J.D. Keppie, and A. Hynes. Geological Society (of London), Special Publication 327. pp. 371–404. doi:10.1144/SP327.16.

Fisher, C.M., Loewy, S.L., Miller, C.F., Berquist, P., Van Schmus, W. R., Hatcher, R.D., Wooden, J.L., and Fullagar, P.D. 2010. Whole-rock Pb and Sm–Nd isotopic constraints on the growth of southeastern Laurentia during Grenvillian orogenesis. Geological Society of America Bulletin, **122**(9–10): 1646–1659. doi:10.1130/B30116.1.

Foose, M.P., and McLelland, J. 1995. Proterozoic low-Ti iron-oxide deposits in New York and New Jersey: Relation to Fe-oxide (Cu–U–Au–REE) deposits and tectonic implications. Geology, **23**: 665–668. doi:10.1130/0091-7613(1995)023<0665:PLTIOD>2.3.CO;2.

Gao, Y., Yang, Z., Hou, Z., Wei, R., Meng, X., and Tiang, S. 2010. Eocene potassic and ultrapotassic magmatism in south Tibet: new constraints on mantle source characteristics and geodynamic processes. Lithos, **117**(1–4): 20–32. doi:10.1016/j.lithos.2010.02.003.

Gauthier, M., and Chartrand, F. 2005. Metallogeny of the Grenville Province revisited. Canadian Journal of Earth Sciences, **42**(10): 1719–1734. doi:10.1139/E05-051.

Gauthier, M., Chartrand, F., Cayer, A., and David, J. 2004. The Kwyjibo Cu–REE–U–Au–Mo–F property, Quebec: a Mesoproterozoic polymetallic iron oxide deposit in the northeastern Grenville Province. Economic Geology and the Bulletin of the Society of Economic Geologists, **99**(6): 1177–1196. doi:10.2113/99.6.1177.

Gervais, F., Nadeau, L., and Malo, M. 2004. Migmatitic structures and solid-state diapirism in orthogneiss domes, eastern Grenville Province. *In* Gneiss domes in orogeny. *Edited by* D.L. Whitney, C. Teyssier, and C.S. Siddoway. Geological Society of America, Special Paper 380. pp. 359–378.

Gerya, T.V., Yuen, D.A., and Maresch, W.V. 2004. Thermomechanical modelling of slab detachment. Earth and Planetary Science Letters, **226**(1–2): 101–116. doi:10.1016/j.epsl.2004.07.022.

Gobeil, A., Brisebois, D., Clark, T., Verpaelst, P., Madore, L., Wodicka, N., and Chevé, S. 2003. Géologie de la moyenne Côte-Nord. *In*: Géologie et Ressources Minérales de la Partie Est de la Province de Grenville. *Edited by* D. Brisebois and T. Clark. Ministère de Richesses naturelles, Québec, Qué., DV2002-2003. pp. 9-57.

Gower, C.F., Hall, J., Kilfoil, G.J., Quinlan, G.M., and Wardle, R.J. 1997. Roots of the Labradorian orogen in the Grenville Province in southeast Labrador: Evidence from marine, deep-seismic reflection data. Tectonics, **16**(5): 795–809. doi:10.1029/97TC01284.

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.

Green, A.G., Milkereit, B., Davidson, A., Spencer, C., Hutchinson, D.R., Cannon, W.F., et al. 1988. Crustal structure of the Grenville Front and adjacent terranes. Geology, **16**(9): 788–792. doi:10.1130/0091-7613(1988)016<0788:CSOTGF>2.3.CO;2.

Grimes, S.W., and Copeland, P. 2004. Thermochronology of the Grenville Orogeny in west Texas. Precambrian Research, **131**: 23–54.

Guillou-Frottier, L., Mareschal, J.C., Jaupart, C., Gariépy, C., Lapointe, R., and Bienfait, G. 1995. Heat flow variations in the Grenville Province, Canada. Earth and Planetary Science Letters, **136**(3–4): 447–460. doi:10.1016/0012-821X(95)00187-H.

Hacker, B.R., Gnos, E., Ratschbacher, L., Grove, M., McWilliams, M., Sobolev, S.V., Wan, J., and Zhenhan, W. 2000. Hot and dry deep crustal xenoliths from Tibet. Science, **287**(5462): 2463–2466. doi:10.1126/science.287.5462.2463. PMID:10741961.

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. The Journal of Geology, **101**(5): 575–589. doi:10.1086/648250.

Hanmer, S., and McEachern, S. 1992. Kinematic and rheological evolution of a crustal-scale ductile thrust zone, central Metasedimentary Belt, Grenville orogen, Ontario. Canadian Journal of Earth Sciences, **29**(8): 1779–1790. doi:10.1139/e92-140.

Hébert, C., van Breemen, O., and Cadieux, A.-M. 2009. Région du réservoir Pipmuacan (SNRC 22E): Synthèse géologique et carte à 1 : 250 000. Ministère des Ressources naturelles et de la faune, Gouvernement du Québec, RG 2009-01.

Herd, R.K., Ackerman, D., Windley, B.F., and Rondot, J. 1986. Sapphirine–garnet rocks, St. Maurice area, Québec: petrology and implications for tectonics and metamorphism. *In* The Grenville Province. *Edited by* J.M. Moore, A. Davidson, and A.J. Baer. Geological Association of Canada, Special Paper 31. pp. 241–253.

Heumann, M.J., Bickford, M.E., Hill, B.M., McLelland, J.M., Selleck, B.W., and Jercinovic, M.J. 2006. Timing of anatexis in metapelites from the Adirondack Lowlands and southern Highlands: A manifestation of the Shawinigan orogeny and subsequent anorthosite–mangerite–charnockite–granite magmatism. Geological Society of America Bulletin, **118**(11–12): 1283–1298. doi:10.1130/B25927.1.

Hitzman, M.W., Oreskes, N., and Einaudi, M.T. 1992. Geological characteristics and tectonic setting of Proterozoic iron oxide (Cu–U–Au–REE) deposits. Precambrian Research, **58**(1–4): 241–287. doi:10.1016/0301-9268(92)90121-4.

Hobbs, B., Regenauer-Lieb, K., and Ord, A. 2007. Thermodynamics of folding in the middle to lower crust. Geology, **35**(2): 175–178. doi:10.1130/G23188A.

Hynes, A., and Rivers, T. 2010. Protracted continental collision — evidence from the Grenville Orogen. Canadian Journal of Earth Sciences, **47**: 591–620.

Hynes, A., Indares, A., Rivers, T., and Gobeil, A. 2000. Lithoprobe line 55: integration of out-of-plane seismic results with surface structure, metamorphism, and geochronology, and the tectonic evolution of the eastern Grenville Province. Canadian Journal of Earth Sciences, **37**: 341–358.

Indares, A., and Martignole, J. 1990. Metamorphic constraints on the evolution of the gneisses from the parautochthonous and allochthonous polycyclic belts, Grenville Province, western Quebec. Canadian Journal of Earth Sciences, **27**(3): 357–370. doi:10.1139/e90-033.

Indares, A., Dunning, G., and Cox, R. 2000. Tectonothermal evolution of deep crust in a Mesoproterozoic continental collision setting: the Manicouagan example. Canadian Journal of Earth Sciences, **37**: 325–340.

Indares, A., Dunning, G., Cox, R., Gale, D., and Connelly, J. 1998. High-pressure, high-temperature rocks from the base of thick continental crust: geology and age constraints from the Manicouagan Imbricate Zone, eastern Grenville Province. Tectonics, **17** (3): 426–440. doi:10.1029/98TC00373.

Jamieson, R.A., Culshaw, N.G., Wodicka, N., Corrigan, D., and Ketchum, J.W.F. 1992. Timing and tectonic setting of Grenvillian metamorphism — constraints from a transect along Georgian Bay, Ontario. Journal of Metamorphic Geology, **10**(3): 321–332. doi:10.1111/j.1525-1314.1992.tb00087.x.

Jamieson, R.A., Culshaw, N.G., and Corrigan, D. 1995. North-west propagation of the Grenville orogen: Grenvillian structure and metamorphism near Key Harbour, Georgian Bay, Ontario, Canada. Journal of Metamorphic Geology, **13**(2): 185–207. doi:10.1111/j.1525-1314.1995.tb00213.x.

Jamieson, R.A., Beaumont, C., Nguyen, M.H., and Culshaw, N.G. 2007. Synconvergent ductile flow in variable-strength continental crust: numerical models with application to the western Grenville orogen. Tectonics, **26**(5): TC5005. doi:10.1029/2006TC002036.

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**(4): 517–539. doi:10.1139/E09-070.

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, Boulder, Colo., Memoir 197. pp. 357–378.

Jolivet, L., Famin, V., Mehl, C., Parra, T., Aubourg, C., Hébert, R., and Philippot, P. 2004. Strain localization during crustal-scale boudinage to form extensional metamorphic domes in the Aegean Sea. *In* Gneiss domes in orogeny. *Edited by* D.L. Whitney, C. Teyssier, and C.S. Siddoway. Geological Society of America, Special Paper 380. pp. 185–210.

Kellett, R.L., Barnes, A.E., and Rive, M. 1994. The deep structure of the Grenville Front: a new perspective from western Quebec. Canadian Journal of Earth Sciences, **31**(2): 282–292. doi:10.1139/e94-027.

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**(1–2): 25–45. doi:10.1016/S0301-9268(97)00079-X.

Klemens, W.P., and Schwerdtner, W.M. 1997. Emplacement and deformation of granite pegmatite dykes in a mid-crustal regime of late orogenic extension, southwest Grenville Province, Ontario, Canada. Geological Magazine, **134**(3): 287–295. doi:10.1017/S0016756897007048.

Klemperer, S.L. 2006. Crustal flow in Tibet: geophysical evidence for the physical state of Tibetan lithosphere, and inferred patterns of active flow. *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. pp. 39–70.

Kretz, R. 1983. Symbols for rock-forming minerals. The American Mineralogist, **68**: 277–279.

Laubscher, H.P. 1983. Detachment, shear and compression in the central Alps. Geological Society of America, Memoir 158. pp. 191–211.

Lentz, D. 1991. Radioelement distribution in U, Th, Mo, and rare-earth-element pegmatites, skarns and veins in a portion of the Grenville Province, Ontario and Quebec. Canadian Journal of Earth Sciences, **28**(1): 1–12. doi:10.1139/e91-001.

Lentz, D. 1992. Petrogenesis and geochemical composition of biotites in rare-element granitic pegmatites in the southwestern Grenville Province, Canada. Mineralogy and Petrology, **46**(3): 239–256. doi:10.1007/BF01164649.

Lentz, D. 1996. U, Mo and REE mineralization in late-tectonic granitic pegmatites, southwestern Grenville Province, Canada. Ore Geology Reviews, **11**(4): 197–227. doi:10.1016/0169-1368(95)00034-8.

Lentz, D.R., and Creaser, R.A. 2005. Re–Os model age constraints on the genesis of the Moss molybdenite pegmatite-aplite deposit, southwestern Grenville Province, Quyon, Quebec, Canada. Exploration and Mining Geology, **14**(1–4): 95–103. doi:10.2113/gsemg.14.1-4.95.

Lentz, D.R., and Suzuki, K. 2000. A low F pegmatite-related Mo skarn from the southwestern Grenville Province, Ontario, Canada:

phase equilibria and petrogenetic implications. Economic Geology and the Bulletin of the Society of Economic Geologists, **95**(6): 1319–1337. doi:10.2113/95.6.1319.

Li, Z.X., Bogdanova, S.V., Collins, A.S., Davidson, A., De Waele, B., Ernst, R.E., et al. 2008. Assembly, configuration, and breakup history of Rodinia: a synthesis. Precambrian Research, **160**(1–2): 179–210. doi:10.1016/j.precamres.2007.04.021.

Loewy, S.L., Connelly, J.N., Dalziel, I.W.D., and Gower, C. 2003. Eastern Laurentia in Rodinia: Constraints from whole-rock Pb and U/Pb geochronology. Tectonophysics, **375**(1–4): 169–197. doi:10.1016/S0040-1951(03)00338-X.

Lopez-Martinez, M., and York, D. 1983. Further thermochronometric unravelling of the age and paleomagnetic record of the southwest Grenville Province. Canadian Journal of Earth Sciences, **20**(6): 953–960. doi:10.1139/e83-084.

Ludden, J., and Hynes, A. 2000. The Lithoprobe Abitibi–Grenville transect: two billion years of crust formation and recycling in the Precambrian Shield of Canada. Canadian Journal of Earth Sciences, **37**: 459–476.

Lumbers, S.B., Heaman, L.M., Vertolli, V.M., and Wu, T.-W. 1990. Nature and timing of Middle Proterozoic magmatism in the Central Metasedimentary Belt, Ontario. *In*: Mid-Proterozoic Laurentia–Baltica. *Edited by* C.F. Gower, T. Rivers, and A.B. Ryan. Geological Association of Canada, Special Paper 38. pp. 243–276.

Lupulescu, M.V., Chiarenzelli, J.R., Pullen, A.T., and Price, J.T. 2011. Using pegmatite geochronology to constrain temporal events in the Adirondack Mountains. Geosphere, **7**(1): 23–39. doi:10.1130/GES00596.1.

Madore, L., Verpaelst, P., Brisebois, D., Hocq, M., and Dion, D.J. 1999. Géologie de la région du lac Allard (SNRC 12 L/11). Ministère de Richesses naturelles, Québec, Qué., RG 96-05.

Magrina, B., Jébrak, J., and Cuney, M. 2005. Le magmatisme de la région de Kwyjibo, Province du Grenville (Canada): intérêt pour les minéralisations de type fer-oxydes associées. Canadian Journal of Earth Sciences, **42**(10): 1849–1864. doi:10.1139/E05-103.

Malavieille, J. 1987. Extensional shearing deformation and kilometer-scale “a”-type folds in a Cordilleran metamorphic core complex (Raft River Mountains, northwestern Utah). Tectonics, **6**(4): 423–448. doi:10.1029/TC006i004p00423.

Malavieille, J. 1993. Late orogenic extension in mountain belts: insights from the Basin and Range and the late Paleozoic Variscan belt. Tectonics, **12**(5): 1115–1130. doi:10.1029/93TC01129.

Mareschal, J.C., Jaupart, C., Gariépy, C., Cheng, L.Z., Guillou-Frottier, L., Bienfait, G., and Lapointe, R. 2000. Heat flow and deep thermal structure near the southeastern edge of the Canadian Shield. Canadian Journal of Earth Sciences, **37**: 399–414.

Martignole, J., and Friedman, R. 1998. Geochronological constraints on the last stages of terrane assembly in the central part of the Grenville Province. Precambrian Research, **92**(2): 145–164. doi:10.1016/S0301-9268(98)00072-2.

Martignole, J., and Martelat, J.-E. 2005. Proterozoic mafic dykes as monitors of HP granulite facies metamorphism in the Grenville Front Tectonic Zone (western Quebec). Precambrian Research, **138**(3–4): 183–207. doi:10.1016/j.precamres.2005.05.002.

Martignole, J., and Reynolds, P. 1997. ^{40}Ar/^{39}Ar thermochronology along a western Québec transect of the Grenville Province, Canada. Journal of Metamorphic Geology, **15**(2): 283–296. doi:10.1111/j.1525-1314.1997.00018.x.

Martignole, J., Machado, N., and Indares, A. 1994. The Wakeham Terrane: A Mesoproterozoic terrestrial rift in the eastern part of the Grenville Province. Precambrian Research, **68**(3–4): 291–306. doi:10.1016/0301-9268(94)90034-5.

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**: 291–306.

McDougall, I., and Harrison, T.M. 1999. Geochronology and thermochronology by the ^{40}Ar/^{39}Ar method. 2nd ed. Oxford University Press, Oxford, UK. 212 p.

McEachern, S.J., and van Breemen, O. 1993. Age of deformation within the Central Metasedimentary Belt boundary thrust zone, southwest Grenville orogen: constraints on the collision of the Mid-Proterozoic Elzevir terrane. Canadian Journal of Earth Sciences, **30**(6): 1155–1165. doi:10.1139/e93-098.

McLelland, J.M., and Isachsen, Y.W. 1986. Synthesis of geology of the Adirondack Mountains, New York, and their tectonic setting within the southwestern Grenville Province. *In*: Mid-Proterozoic Laurentia–Baltica. *Edited by* C.F. Gower, T. Rivers, and A.B. Ryan. Geological Association of Canada, Special Paper 38. pp.61–94.

McLelland, J.M., Chiarenzelli, J., Whitney, P., and Isachsen, W.J. 1988. U–Pb zircon geochronology of the Adirondack Mountains and implications for their geologic evolution. Geology, **16**(10): 920–924. doi:10.1130/0091-7613(1988)016<0920:UPZGOT>2.3.CO;2.

McLelland, J., Daley, J.S., and McLelland, J.M. 1996. The Grenville orogenic cycle (ca. 1350–1000 Ma): an Adirondack perspective. Tectonophysics, **265**(1–2): 1–28. doi:10.1016/S0040-1951(96)00144-8.

McLelland, J., Hamilton, M., Selleck, B., McLelland, J., Walker, D., and Orrell, S. 2001. Zircon U–Pb geochronology of the Ottawan Orogeny, Adirondack Highlands, New York: regional and tectonic implications. Precambrian Research, **109**(1–2): 39–72. doi:10.1016/S0301-9268(01)00141-3.

McLelland, J., Morrison, J., Selleck, B., Cunningham, B., Olson, C., and Schmidt, K. 2002. Hydrothermal alteration of late- to post-tectonic Lyon Mountain granitic gneiss, Adirondack Mountains, New York; origin of quartz–sillimanite segregations, quartz–albite lithologies, and associated Kiruna-type low-Ti Fe-oxide deposits. Journal of Metamorphic Geology, **20**(1): 175–190. doi:10.1046/j.0263-4929.2001.00345.x.

McLelland, J.M., Selleck, B.W., Bickford, M.E., Williams, M.L., and Jercinovic, M.J. 2009. The Adirondacks as the Grenville–Appalachian Link. GAC–MAC–AGU meeting, Toronto, Ont. Abstract.

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. Geological Society of America, Memoir 206.

Mezger, K., Rawnsley, C.M., Bohlen, S.R., and Hanson, G.N. 1991. U–Pb garnet, sphene, monazite, and rutile ages: Implications for the duration of high-grade metamorphism and cooling histories, Adirondack Mtns, New York. The Journal of Geology, **99**(3): 415–428. http://www.jstor.org/stable/30062625. doi:10.1086/629503.

Mezger, K., Essene, E.J., van der Pluijm, B.A., and Halliday, A.N. 1993. U–Pb geochronology of the Grenville Orogen of Ontario and New York: constraints on ancient crustal tectonics. Contributions to Mineralogy and Petrology, **114**(1): 13–26. doi:10.1007/BF00307862.

Mix, H., Mulch, A., Kent-Corson, M.L., and Chamberlain, C.P. 2011. Cenozoic migration of topography in the North American Cordillera. Geology, **39**(1): 87–90. doi:10.1130/G31450.1.

Moecher, D.P., Anderson, E.D., Cook, C.A., and Mezger, K. 1997. The petrogenesis of metamorphosed carbonatites in the Grenville Province, Ontario. Canadian Journal of Earth Sciences, **34**(9): 1185–1201. doi:10.1139/e17-095.

Molnar, P., England, P., and Martinod, J. 1993. Mantle dynamics, uplift of the Tibetan plateau, and the Indian monsoon. Reviews of Geophysics, **31**(4): 357–396. doi:10.1029/93RG02030.

Moore, J.M., and Thompson, P.H. 1980. The Flinton Group: a late Precambrian metasedimentary succession in the Grenville Province. Canadian Journal of Earth Sciences, **17**: 1685–1707.

Morrison, J., and Valley, J.W. 1988. Post-granulite facies fluid infiltration in the Adirondack Mountains. Geology, **16**(6): 513–516. doi:10.1130/0091-7613(1988)016<0513:PGFFII>2.3.CO;2.

Mosher, S., Hoh, A.M., Zumbro, J.A., and Reese, J.F. 2004. Tectonic evolution of the eastern Llano Uplift, central Texas: A record of Grenville orogenesis along southern Laurentia. *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, Boulder, Colo., Memoir 197. pp. 783–798.

Mosher, S., Levine, J.S.F., and Carlson, W.D. 2008. Mesoproterozoic plate tectonics: a collisional model for the Grenville-aged orogenic belt in the Llano Uplift, central Texas. Geology, **36**(1): 55–58. doi:10.1130/G24049A.1.

Nadeau, L., Brouillette, P., and Hébert, C. 2006. Architecture et évolution tectonique de l'orogène du Grenville, région Portneuf-Mauricie, Québec. Réunion annuelle conjointe, Association géologique du Canada — Association minéralogique du Canada, Montréal, Qué., 2006, Livret-guide d'excursion B6.

Onstott, T.C., and Peacock, M.W. 1987. Argon retentivity of hornblendes: A field experiment in a slowly cooled metamorphic terrane. Geochimica et Cosmochimica Acta, **51**(11): 2891–2903. doi:10.1016/0016-7037(87)90365-6.

Postel, A.W. 1952. Geology of the Clinton County magnetite district. US. Geological Survey, Professional Paper 237. 87 p.

Pratt, T., Culotta, R., Hauser, E., Nelson, D., Brown, L., Kaufman, S., Oliver, J., and Hinze, W. 1989. Major Proterozoic basement features of the eastern mid-continent of North America revealed by recent COCORP profiling. Geology, **17**(6): 505–509. doi:10.1130/0091-7613(1989)017<0505:MPBFOT>2.3.CO;2.

Ramberg, H. 1981. The role of gravity in orogenic belts. *Edited by* K. C. McClay and N.J. Price. Geological Society (of London), Special Publication 9. pp. 125–140.

Rey, P., Vanderhaeghe, O., and Teyssier, C. 2001. Gravitational collapse of the continental crust: definition, regimes and modes. Tectonophysics, **342**(3–4): 435–449. doi:10.1016/S0040-1951(01)00174-3.

Rey, P.F., Teyssier, C., and Whitney, D.L. 2009. The role of partial melting and extensional strain rates in the development of metamorphic core complexes. Tectonophysics, **477**: 135–144.

Reynolds, P.H., Culshaw, N.G., Jamieson, R.A., Grant, S.L., and McKenzie, K.J. 1995. ^{40}Ar/^{39}Ar traverse — Grenville Front Tectonic Zone to Britt domain, Grenville Province, Ontario, Canada. Journal of Metamorphic Geology, **13**(2): 209–221. doi:10.1111/j.1525-1314.1995.tb00214.x.

Rivers, T. 1976. Structures and textures of metamorphic rocks, Ompah area, Grenville Province, Ontario. Ph.D. thesis, University of Ottawa, Ottawa, Ont.

Rivers, T. 1997. Lithotectonic elements of the Grenville Province: review and tectonic implications. Precambrian Research, **86**(3–4): 117–154. doi:10.1016/S0301-9268(97)00038-7.

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.precamres.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. pp. 405–444. doi:10.1144/SP327.17.

Rivers, T., and Culshaw, N. Hynes, A., Indares, A., Jamieson, R., and Martignole, J. In press. The Grenville Orogen. *In* Tectonic styles in Canada, revisited: the Lithoprobe perspective. *Edited by* F.A. Cook, J. Percival, and R. Clowes. Geological Association of Canada (GAC), Special Paper 49.

Rivers, T., Martignole, J., Gower, C.F., and Davidson, A. 1989. New tectonic divisions of the Grenville Province, southeast Canadian shield. Tectonics, **8**(1): 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**(5): 867–893. doi:10.1139/e02-025.

Rosenberg, C.L., and Handy, M.R. 2005. Experimental deformation of partially melted granite revisited: implications for the continental crust. Journal of Metamorphic Geology, **23**(1): 19–28. doi:10.1111/j.1525-1314.2005.00555.x.

Rougvie, J.R., Carlson, W.D., Copeland, P., and Connelly, J.N. 1999. Late thermal evolution of Proterozoic rocks in the northeastern Llano Uplift, central Texas. Precambrian Research, **94**(1–2): 49–72. doi:10.1016/S0301-9268(98)00103-X.

Schwerdtner, W.M. 1987. Interplay between folding and ductile shearing in the Proterozoic crust of the Muskoka – Parry Sound region, central Ontario. Canadian Journal of Earth Sciences, **24**(8): 1507–1525. . doi:10.1139/e87-148.

Schwerdtner, W., and Klemens, W. 2008. Structure of Ahmic domain and its vicinity, southwestern Central Gneiss Belt, Grenville Province of Ontario (Canada). Precambrian Research, **167**(1–2): 16–34. doi:10.1016/j.precamres.2008.07.002.

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**(1–2): 39–59. doi:10.1016/0301-9268(91)90055-F.

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): 1021–1032. doi:10.1139/e77-094.

Schwerdtner, W.M., Klemens, W.P., Waddington, D.H., and Vertolli, V.M. 1998. Late-Grenvillian horizontal extension and vertical thinning of Proterozoic gneisses, central Ontario. Geological Society of America Annual Meeting, Toronto, Canada. Field Trip Guide No. 14.

Schwerdtner, W.M., Downey, M.W., and Alexander, S.A. 2004. L-S shape fabrics in the Mazinaw domain and the issue of northwest-directed thrusting in the Composite Arc Belt, southeastern Ontario. *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, Boulder, Colo., Memoir 197. pp. 183–207.

Schwerdtner, W.M., Riller, U.P., and Borowik, A. 2005. Structural testing of tectonic hypothesis 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**(10): 1927–1947. doi:10.1139/e05-047.

Schwerdtner, W.M., Sheng, J.L., and Yang, J.F. 2010. 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**(6): 875–899. doi:10.1139/E10-007.

Selleck, B.W., McLelland, J.M., and Bickford, M.E. 2005. Granite emplacement during tectonic exhumation: The Adirondack example. Geology, **33**(10): 781–784. . doi:10.1130/G21631.1.

Shi, D., Zhao, W., Brown, L., Nelson, D., Zhao, X., Kind, R., et al. 2004. Detection of southward intracontinental subduction of Tibetan lithosphere along the Bangong-Nujiang suture by *P*-to-*S* converted waves. Geology, **32**(3): 209–212. doi:10.1130/G19814.1.

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 U–Pb (SHRIMP) data, Muskoka and Shawanaga domains, Grenville Province, Ontario. Canadian Journal of Earth Sciences, **41**(11): 1339–1365. doi:10.1139/e04-068.

Spear, F.S., and Markussen, J.C. 1997. Mineral zoning, *P–T–X–M* phase relations, and metamorphic evolution of some Adirondack granulites. Journal of Petrology, **38**(6): 757–783. doi:10.1093/petrology/38.6.757.

Streepey, M.M., van der Pluijm, B.A., Essene, E.J., Hall, C.M., and Magloughlin, J.F. 2000. Late Proterozoic (ca. 930 Ma) extension in eastern Laurentia. Geological Society of America Bulletin, **112** (10): 1522–1530. doi:10.1130/0016-7606(2000)112<1522:LPCMEI>2.0.CO;2.

Streepey, M.M., Johnson, E.L., Mezger, K., and van der Pluijm, B.A. 2001. Early history of the Carthage–Colton Shear Zone, Grenville Province, Northwest Adirondacks, New York (U.S.A.). The Journal of Geology, **109**: 479–492. doi:10.1086/320792.

Streepey, M.M., Lithgow-Bertelloni, C., van der Pluijm, B.A., Essene, E.J., and Magloughlin, J.F. 2004. Exhumation of a collisional orogen: A perspective from the North American Grenville Province. *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, Boulder, Colo., Memoir 197: 391–410.

Taylor, M., and Yin, A. 2009. Active structures of the Himalayan–Tibetan orogen and their relationships to earthquake distribution, contemporary strain field, and Cenozoic volcanism. Geosphere, **5** (3): 199–214. doi:10.1130/GES00217.1.

Teyssier, C., and Whitney, D.L. 2002. Gneiss domes and orogeny. Geology, **30**(12): 1139–1142. doi:10.1130/0091-7613(2002)030<1139:GDAO>2.0.CO;2.

Timmermann, H., Jamieson, R.A., Culshaw, N.G., and Parrish, R.R. 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**(7): 1023–1029. doi:10.1139/e17-084.

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**(2): 239–258. doi:10.1139/e01-076.

Tohver, E., Bettencourt, J.S., Tosdal, R., Mezger, K., Leite, W.B., and Payolla, B.L. 2004. Terrane transfer during the Grenville orogeny: Tracing the Amazonian ancestry of southern Appalachian basement through Pb and Nd isotopes. Earth and Planetary Science Letters, **228**(1-2): 161–176. doi:10.1016/j.epsl.2004.09.029.

Tollo, R.P., Corriveau, L., McLelland, J.M., and Bartholomew, M.J. (*Editors*). 2004. Proterozoic tectonic evolution of the Grenville Orogen in North America. Geological Society of America, Memoir 197.

Tuccillo, M.E., Essene, E.J., and van der Pluijm, B.A. 1990. Growth and retrograde zoning in garnets from high-grade pelites: Implications for pressure–temperature paths. Geology, **18**(9): 839–842. doi:10.1130/0091-7613(1990)018<0839:GARZIG>2.3.CO;2.

Tuccillo, M.E., Mezger, K., Essene, E.J., and van der Pluijm, B.A. 1992. Thermobarometry, geochronology and the interpretation of *P–T–t* data in the Britt domain, Ontario Grenville Orogen, Canada. Journal of Petrology, **33**: 1225–1259.

Valley, P.M., Hanchar, J.M., and Whitehouse, M.J. 2009. Direct dating of Fe oxide–(Cu–Au) mineralization by U/Pb zircon geochronology. Geology, **37**(3): 223–226. doi:10.1130/G25439A.1.

van Breemen, O., and Corriveau, L. 2005. U–Pb age constraints on arenaceous and volcanic rocks of the Wakeham Group, eastern Grenville Province. Canadian Journal of Earth Sciences, **42**(10): 1677–1697. doi:10.1139/e05-079.

van der Pluijm, B.A., Mezger, K., Cosca, M.A., and Essene, E.J. 1994. Determining the significance of high-grade shear zones by using temperature–time paths, with examples from the Grenville orogen. Geology, **22**(8): 743–746. doi:10.1130/0091-7613(1994)022<0743:DTSOHG>2.3.CO;2.

Vanderhaeghe, O. 2009. Migmatites, granites and orogeny: Flow modes of partially molten rocks and magmas associated with melt/solid segregation in orogenic belts. Tectonophysics, **477**: 119–134.

Vanderhaeghe, O., and Teyssier, C. 2001. Partial melting and flow of orogens. Tectonophysics, **342**(3–4): 451–472. doi:10.1016/S0040-1951(01)00175-5.

Volkert, R.A. 2004. Mesoproterozoic rocks of the New Jersey Highlands, north-central Appalachians: Petrogenesis and tectonic history. *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, Boulder, Colo., Memoir 197. pp. 697–728.

Volkert, R.A., Zartman, R.E., and Moore, P.B. 2005. U–Pb zircon geochronology of Mesoproterozoic postorogenic rocks and implications for post-Ottawan magmatism and metallogenesis, New Jersey Highlands and contiguous areas, USA. Precambrian Research, **139**(1–2): 1–19. doi:10.1016/j.precamres.2005.06.003.

Wernicke, B. 1985. Uniform-sense normal simple shear of the continental lithosphere. Canadian Journal of Earth Sciences, **22**(1): 108–125. doi:10.1139/e85-009.

White, D.J., Forsyth, D.A., 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 western Quebec. Canadian Journal of Earth Sciences, **37**: 183–192.

Wiener, R.W. 1983. Adirondack Highlands – Northwest Lowlands ‘boundary’: A multiply folded intrusive contact with associated mylonitization. Geological Society of America Bulletin, **94**(9): 1081–1108. doi:10.1130/0016-7606(1983)94<1081:AHLBAM>2.0.CO;2.

Wiener, R.W., McLelland, J., Isachsen, Y.W., and Hall, L.M. 1984. Stratigraphy and structural geology of the Adirondack Mountains, New York: Review and synthesis. *In* The Grenville Event in the Appalachians and related topics. *Edited by* M.J. Bartholomew. Geological Society of America, Special Paper 194. pp. 1–55.

Williams, P.F., Jiang, D., and Lin, S. 2006. Interpretation of deformation fabrics of infrastructure zone rocks in the context of channel flow and other tectonic models. *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. pp. 221–235.

Wodicka, N., David, J., Parent, M., Gobeil, A., and Verpaelst, P. 2003. Géochronologie U–Pb et Pb–Pb de la région de Sept-Îles–Natashquan, Province de Grenville, moyenne Côte-Nord. *In* Géologie et Ressources Minérales de la Partie Est de la Province de Grenville. *Edited by* D. Brisebois and T. Clark. Ministère de Richesses naturelles, Québec, Qué, DV2002–2003. pp. 59–117.

Zhang, P., Shen, Z., Wang, M., Gan, W., Burgmann, R., Molnar, P., et al. 2004. Continuous deformation of the Tibetan Plateau from global positioning system data. Geology, **32**(9): 809–812. doi:10.1130/G20554.1.