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[1]

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Abstract: Lithoprobe line 55, in the Grenville Province of eastern Quebec, provides unusually good control on the three-dimensional (3-D) geometry and structural relationships among the major lithological units there. Archean basement underlies the exposed Proterozoic rocks, along the entire seismic line, and there is a lateral ramp in this basement immediately behind a lobate stack of thrust slices of high-pressure metamorphic rocks comprising the Manicouagan Imbricate Zone (MIZ). Integration of the 3-D geometry with *P–T* and geochronological data allows derivation of a tectonic model for the region. The MIZ was buried to depths >60 km at 1050 Ma. Preservation of its high-pressure assemblages, and the absence of metamorphism at 990 Ma, which is characteristic of lower pressure metamorphic rocks that tectonically overlie them, indicates the MIZ rocks were rapidly unroofed, early in the tectonic history. There were two discrete pulses of crustal thickening during the Grenvillian Orogeny in this region. The first, involving imbrication of Labradorian and Pinwarian rocks that comprised part of southeast Laurentia, culminated in the Ottawan pulse at ca. 1050 Ma, and produced the high-pressure metamorphism of the MIZ. Its effects were rapidly reversed, with extrusion of the MIZ rocks to shallow crustal levels at ca. 1020 Ma. The crust was again thickened, with the Moho subsiding to depths >60 km, in the Rigolet pulse at ca. 990 Ma. The site of extrusion of the MIZ was probably controlled by the subsurface lateral ramp. High geothermal gradients indicate that extrusion may have been aided by lithospheric delamination in the crustal-thickening zone.

Résumé : La ligne 55 du projet Lithoprobe, dans la Province de Grenville du Québec oriental, procure un contrôle exceptionnellement utile pour préciser la géométrie tridimensionnelle et établir les relations structurales parmi les plus importantes unités lithologiques de cette région. Le socle archéen est sous-jacent aux roches protérozoïques qui sont exposées tout le long de la ligne sismique; ce socle comprend une rampe latérale qui apparaît immédiatement derrière un empilement d'écailles divisé en lobes de roches métamorphiques de haute-pression, incluant la zone imbriquée de Manicouagan (ZIM). La combinaison de la géométrie tridimensionnelle avec les données de *P–T* et la géochronologie permet l'élaboration d'un modèle tectonique adapté à cette région. Il y a 1050 Ma, la ZIM était enfouie à des profondeurs >60 km. La préservation des assemblages de haute-pression, et l'absence du métamorphisme daté de 990 Ma qui caractérise les roches métamorphiques de plus basse-pression les recouvrant, révèlent que les roches de la ZIM furent rapidement dénudées, et dès le début de l'histoire tectonique. Deux pulsions d'épaississement crustal distinctes ont surgi durant l'orogénie de Grenville dans cette région. La première est une imbrication des roches du Labradorien et du Pinwarien incorporant une portion du sud-est de Laurentia, et dont la culmination date d'environ 1050 Ma lors de la pulsion d'Ottawan responsable du métamorphisme de haute-pression dans la ZIM. Ses effets ont rapidement rétrogradés, avec l'extrusion vers 1020 Ma des roches de la ZIM à des niveaux crustaux peu profonds. Un nouvel épaississement de la croûte a suivi, il était accompagné de la subsidence du Moho à des profondeurs >60 km vers 990 Ma lors de la pulsion de Rigolet. Il est probable que la rampe latérale de subsurface ait joué un rôle important dans la détermination du site d'extrusion de la ZIM. Les gradients géothermiques élevés suggèrent que l'extrusion a pu être facilitée par une délamination lithosphérique dans la zone d'épaississement de la croûte.

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Introduction

The Grenville Province consists of Archean and Paleoproterozoic rocks that formed part of the southeastern margin of the Canadian Shield and several terranes that were accreted during the Mesoproterozoic (e.g., Rivers 1997). Rocks of the province were reworked in the Grenvillian Orogeny, due to a continental collision between 1.2 and 1.0 Ga. The geology of the western and eastern parts of the Grenville Province is relatively well understood (e.g., Easton 1992; Gower et al. 1992, 1997; Rivers et al. 1993; Davidson 1995; McLelland et al. 1996; Martignole and Calvert 1996; Carr et al. 2000; Martignole et al. 2000), but much of the central part of the Grenville Province has not been studied in detail. Structural, metamorphic and geochronological studies associated with the acquisition of Lithoprobe line 55 (Eaton et al. 1995), and recent mapping by the Ministère des richesses naturelles of Quebec, have resulted in significant improvements in our understanding of the geological evolution of the central Grenville Province. This paper summarizes that understanding.

Seismic-reflection data from line 55, in conjunction with structural data from field mapping, illustrate the three-dimensional architecture of the orogen (Hynes and Eaton 1999). Metamorphic (Hynes and St-Jean 1997; Indares et al. 2000) and geochronological data (Scott and Hynes 1994; Indares et al. 1998) constrain the depths of burial of the different tectonic units and the timing of the burial. In this paper, we review the relationships between the structure, the metamorphism, and the geochronology, and use them to interpret the tectonic evolution of the central Grenville Province.

Geological setting

The Grenville Province is bounded to the northwest by the Grenville Front, which is the northwestern limit of perceptible effects of the Grenvillian Orogeny. Archean and Paleoproterozoic rocks northwest of the front can be traced into the Grenville Province in several places (Rivers et al. 1989). These rocks, and terranes that were accreted to them during the Mesoproterozoic, are disposed in a crustal-scale, broadly southeast-dipping imbricate stack (e.g., Davidson 1986; Rivers et al. 1989; White et al. 1994). Prior to the Grenvillian Orogeny, the southeast margin of Laurentia was composed of an Archean (Superior Province) and Paleoproterozoic basement, tectonically overlain by imbricated-arc and back-arc deposits of Paleoproterozoic (ca. 1670–1610 Ma, Labradorian; Gower et al. 1992) and Mesoproterozoic (ca. 1500–1350 Ma, Pinwarian; Tucker and Gower 1994) age. The collisional Grenvillian Orogeny took place from about 1190 to 990 Ma, and at least three pulses of crustal thickening and associated metamorphism have been recognized (Rivers 1997), of which two are known in the Manicouagan area. These are the ca. 1080–1020 Ma Ottawan pulse and the ca. 1010–990 Ma Rigolet pulse.

Rivers et al. (1989) referred to the northwestern belt of the Grenville Province, which contains many units that can be lithologically linked to the foreland, as the Parautochthonous Belt. In this belt, the Grenvillian metamorphic overprint typically exhibits a Barrovian character. The structurally overlying Allochthonous Belt, in contrast, contains tectonic slices with several units not found in the foreland or the Parautochthonous Belt and, although not exotic to Laurentia, was inferred to have been substantially transported from the southeast. Grenvillian metamorphism in the Allochthonous Belt is variable in intensity, but is generally of high-temperature, medium-pressure character.

Since division of the Grenville Province into belts by Rivers et al. (1989), high-pressure rocks have been recognized at several locations. In the Lithoprobe line 55 transect, high-pressure rocks occur in the Manicouagan Imbricate Zone (MIZ), which is structurally sandwiched between the Parautochthonous and Allochthonous belts. The high-pressure signature of the MIZ implies that it has undergone a different metamorphic evolution from the adjacent belts, and, therefore, that it comprises an additional element in the tectonic stratigraphy. It has been referred to as the High Pressure belt by Indares (1997), and in eastern Quebec and adjacent parts of western Labrador it is distinguished by the widespread occurrence of high-pressure granulite and subordinate eclogite-facies rocks (Indares and Rivers 1995; Indares et al. 2000), testifying to the presence at the surface of formerly deeply buried rocks of continental affinity. The structurally overlying tectonic slices in the Allochthonous Belt have a high-temperature, medium-pressure metamorphic signature (Scott and Hynes 1994; Hynes and St-Jean 1997).

Lithoprobe line 55 was shot along Quebec Highway 389, which skirts the eastern shoreline of Manicouagan Reservoir for a distance of 150 km (Fig. 1). At its northeastern end it lies in parautochthonous Paleoproterozoic rocks of the Gagnon terrane. It passes southwestward into Labradorian and Pinwarian rocks of the MIZ, and then into the tectonically overlying Hart Jaune and Berthé terranes of the Allochthonous Belt. Line 55 spans only a relatively short section of exposed high-pressure rocks, in the MIZ, because of limitations imposed by the road network. High-pressure rocks are, however, widely exposed to the northwest of the line, and the structures in this region project into and are prominent in the seismic data. As a result, line 55 provides the potential for detailed interpretation of the subsurface configuration of the High Pressure belt and its relationship to the medium-pressure tectonic cover in the overlying Allochthonous Belt.

The prominent circular form of Manicouagan Reservoir (Fig. 1) resulted from a Triassic meteorite impact (Murtaugh 1976), and the central island in the Manicouagon Reservoir is approximately two-thirds covered by Triassic volcanic and volcaniclastic rocks. Many of the underlying Precambrian rocks on the island exhibit evidence of the impact, in the form of intense cataclasis, brecciation, and locally widespread pseudotachylite veining. On the outer margin of the Manicouagon Reservoir, however, evidence of the impact is sparse, and there is no indication that the structure and mineral assemblages of the Precambrian rocks were significantly affected. The outer margin of the Manicouagon Reservoir provides incomparable access and exposure when the water level is low and many of our data are derived from there (e.g., Hynes and St-Jean 1997; Indares 1997; Hynes and Eaton 1999; Indares et al. 2000). Detailed mapping away from the Manicouagon Reservoir shorelines in the region of

Fig. 1 is restricted to the vicinity of Gagnon (Clarke 1977) and the Haut Plateau de Manicouagan, southeast of Highway 389, between Lac Petit Manicouagan and Relais Gabriel (Kish 1968; Gobeil 1996*a*, 1996*b*). In the following paragraphs, the regional geology in the vicinity of line 55 is summarized, from the structurally lowest to the highest units.

Archean basement

Archean basement in the region consists of granoblastic quartzofeldspathic gneisses with hornblende- and pyroxene-rich mafic layers, and strongly foliated and fissile gneisses with varying degrees of compositional layering and migmatization (Clarke 1977). The foliated gneisses are similar in appearance to some of the Proterozoic metasedimentary rocks of the overlying Gagnon terrane, and are not always distinguishable from them.

Gagnon terrane

In eastern Quebec and western Labrador, Archean rocks are overlain unconformably by the Knob Lake Group (Rivers 1980), a Paleoproterozoic sequence of quartzofeldspathic and pelitic metasedimentary rocks, marble, quartzite, and iron formation. The Knob Lake Group and its interleaved, reworked basement collectively compose the Gagnon terrane (Rivers et al. 1989). In western Labrador, Gagnon terrane is deformed into a northwest-vergent fold–thrust belt with a Barrovian metamorphic field gradient increasing to the southeast (Rivers 1983*a*, 1983*b*; van Gool 1992). Farther southwest in Quebec, basement reworking is more extensive, and the fold–thrust character of northern Gagnon terrane is progressively overprinted by map-scale upright northwest-trending cross folds (Clarke 1977; Rivers 1983*a*; Fig. 1, this paper). Peak metamorphism in northern Gagnon terrane has been dated by the U–Pb method at 1000–990 Ma (Rigolet pulse; Connelly and Heaman 1993), and there is no evidence for substantive tectonism any earlier than this. $^{40}Ar/^{39}Ar$ dating in Gagnon terrane has yielded cooling ages between 968 and 905 Ma (Dallmeyer and Rivers 1983; Connelly et al. 1995). Northern Gagnon terrane underwent greenschist- to amphibolite-facies metamorphism during the Rigolet pulse. In western Labrador, peak metamorphic conditions were approximately 1100 MPa, 700°C (van Gool 1992). Farther south in eastern Quebec (east of the area in Fig. 1), peak pressure–temperature *(P–T)* conditions attained a minimum of 1600 MPa and 750°C (Indares 1995), but this metamorphism is undated, and may have occurred earlier than the Rigolet pulse (see below).

On the western shore of Manicouagan Reservoir, Gagnon terrane forms a ductile thick- and thin-skinned thrust belt (A. Indares and T. Rivers, unpublished). Archean basement slices are common, but on a scale too fine to be indicated on Fig. 1. Gagnon terrane has been traced as far south as the north- to northeast-trending Relay shear zone, well exposed in the southwest of the Manicouagon Reservoir (Fig. 1). On the western side of this shear zone, *P–T* data derived from garnet cores in metapelites indicate equilibration pressures of at least 1300 MPa (McKee 1997; Fig. 2). No quantification of metamorphic pressures on the eastern side has been possible.

Manicouagan Imbricate Zone

Southwestern Gagnon terrane is tectonically overlain by the MIZ (see Indares et al. 2000), which is exposed in a northwest-trending lobate arc to the north of Manicouagan Reservoir. The contact between Gagnon terrane and the overlying MIZ is a tectonic mélange, at least 1 km thick, in which meta-igneous rocks of the MIZ are mechanically mixed with ductilely deformed metasedimentary rocks of Gagnon terrane (L. MacDonald, A. Indares, and T. Rivers, unpublished data 1996–1998). The MIZ comprises at least four imbricated units. The lowest three, composing Lelukuau terrane, consist of thrust slices of a dismembered Labradorian (ca. 1630 Ma) anorthosite–mangerite–charnockite–granite (AMCG) complex, comprising olivine gabbro, anorthosite and granitoid rocks, locally intruded by 1300 Ma leucogranite (Indares et al. 1998). These imbricate slices are characterized by metamorphic ages between ca. 1050 and 1030 Ma (Ottawan pulse; Indares et al. 1998) and by ubiquitous high-pressure mineral assemblages. Calculated *P–T* conditions increase toward the top of the stack, from 1400 MPa and 700°C at the base to 1800 MPa and 850–900°C in the highest imbricate slice (Indares et al. 2000; Fig. 2). The highest imbricate slices are cut by synmetamorphic mafic intrusions with ages of 1040 Ma (Indares et al. 1998).

The upper unit of MIZ, exposed in the northeastern part of the Manicouagon Reservoir and known as Tshenukutish terrane, comprises a variety of metaigneous and supracrustal rocks in gently south-dipping tectonic slices bounded by extensional shear zones. Tshenukutish terrane consists of Labradorian (~1650 Ma) AMCG rocks and Pinwarian (1467 Ma) diorite with rafts of supracrustal rocks (metapelite, marble, and quartzite) intruded by 1170 Ma ophitic metagabbro (Indares et al. 1998). In addition, there are two tectonic slivers of anorthosite, the 1214 Ma Baie du Nord anorthosite and the 1170 Ma Brien anorthosite (Indares et al. 1998; Scott and Hynes 1994; too small to be illustrated on Fig. 1). Evidence of high-pressure metamorphism, dated at ~1046 Ma at the bottom of the stack and ~1020 Ma at the top, is variably preserved in Tshenukutish terrane, with maximum *P–T* conditions in the range 1400–1800 MPa and 750–850°C (Indares et al. 2000). The southeastern boundary of Tshenukutish terrane is intruded by the post-Ottawan Hart Jaune granite, which has been dated at 1017–1007 Ma (Indares et al. 1998).

Tshenukutish terrane is interpreted to extend across the central island of Manicouagan Reservoir, incorporating the Memory Bay anorthosite (Fig. 1), which possesses mineralogical evidence of high-pressure metamorphism (A. Indares and T. Rivers, unpublished data). Tshenukutish terrane continues south of the Manicouagon Reservoir, where it has been subdivided into the Southwest and Island domains (Eaton et al. 1995). These domains are thought to represent distinct fault slices, based on a prominent topographic lineament between them. There is no direct evidence for a fault, which is consequently named the Cryptic shear zone (CSZ, Fig. 1). The Southwest and Island domains consist largely of granitoid intrusions, with subordinate metasedimentary screens. Local mafic rocks preserve evidence of high-pressure metamorphism, but were widely reequilibrated under amphibolite-facies conditions (900 MPa and 700°C;

Fig. 1. Geological sketch map of Manicouagan region adjacent to line 55. Location identified in index map, lower-left corner. Lithoprobe line 55 follows Highway 389 (dashed line); white circles labelled “1,” “10,” “20,” “30,” “40,” and “51”: positions of shotpoints on line. Heavy solid lines**,** faults. BAn, Berthé anorthosite; MAn, Memory Bay anorthosite; RAn, Raudot anorthosite; SAn, Seignelay anorthosite. CSZ, Cryptic shear zone; GaSZ, Gabriel shear zone; HJSZ, Hart Jaune shear zone; HWSZ, Highway shear zone; RSZ, Relay shear zone; TNSZ, Triple-Notch shear zone. ID, Island domain; SWD, Southwest domain. Anorthositic rocks are coloured according to the terrane to which they belong. Geology after Kish 1968; Clarke 1977; Avramtchev 1983, Gobeil 1996*a*, 1996*b*, as well as unpublished mapping by the authors.

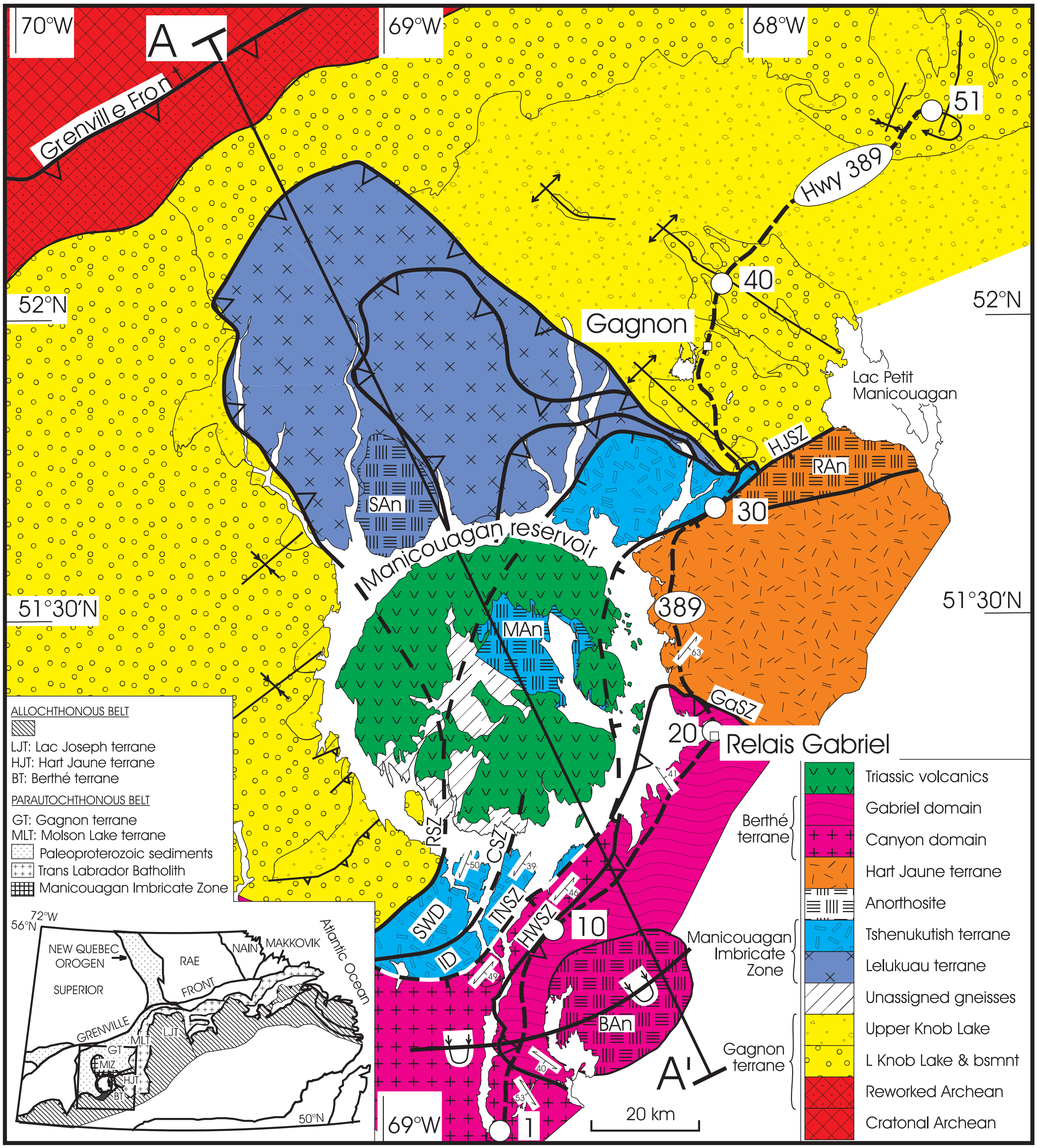


Fig. 2). There are as yet no age data from these two domains.

Northeast of Manicouagan Reservoir, the MIZ is juxtaposed to Hart Jaune terrane (see below) along the late, high-angle Hart Jaune shear zone (Fig. 1), a ductile structure with late normal displacement that strikes toward the southwest into the Manicouagon Reservoir. This shear zone is inferred to reappear on the southern shore of the Manicouagon Reservoir, where it is known as the Triple Notch shear zone (Fig. 1). Henceforth, this boundary is referred to as the Hart Jaune–Triple Notch shear zone (HJ–TNSZ). The HJ–TNSZ juxtaposes the Island domain (part of Tshenukutish terrane) to the Canyon domain (part of Berthé terrane; see below).

Hart Jaune terrane

Hart Jaune terrane consists predominantly of granulite-facies metagabbroic rocks and layered two-pyroxene metabasites, with subordinate calc-silicates and metapelites. Along the shores of the Manicouagon Reservoir, where the dominant rock types are layered two-pyroxene metabasites, foliations dip consistently toward the southeast. Rocks of Hart Jaune terrane experienced an early stage of largely garnet-free, granulite-facies metamorphism and a later granulite-facies overprint in which garnet developed in some rocks (Hynes and St-Jean 1997). Metamorphic pressures for the second event were below 700 MPa (Hynes and St-Jean 1997; Fig. 2). Geochronological studies of these gneisses have provided support for their polymetamorphic character, with the earlier metamorphism dated at ca. 1470 Ma (Pinwarian) and the later metamorphic overprint (presumably the second granulite-facies event) dated at 989 Ma (Rigolet pulse; Scott and Hynes 1994). The Pinwarian age of the earlier metamorphism in Hart Jaune terrane, together with the presence of mafic dykes chemically similar to those in Tshenukutish terrane north of the Manicouagon Reservoir, provide evidence for lithological linkages between Hart Jaune terrane and Tshenukutish terrane, despite their very different metamorphic signatures.

Berthé terrane

Berthé terrane comprises the Canyon and Gabriel domains. The Canyon domain, southeast of the HJ–TNSZ in the southern part of the Manicouagon Reservoir (Fig. 1), consists of a granitoid complex that encloses large rafts of metapelite. These rocks show no evidence for more than one phase of metamorphism. Where well layered, they locally display evidence of S-plunging sheath folds indicating very high strains. The metapelites display amphibolite-facies mineral assemblages, with peak metamorphic conditions of ca. 900 MPa and 700°C, and subsequent reequilibration at ca. 700 MPa and 700°C (Hynes and St-Jean 1997; Fig. 2, this paper). Ages of the granitoid rocks and the metamorphism have not been determined.

Canyon domain is tectonically overlain by Gabriel domain along the folded southeast-dipping Highway shear zone (HWSZ; Fig. 1). Gabriel domain consists of well-layered, migmatitic, quartzofeldspathic gneisses of probable metasedimentary origin, with concordant amphibolite sheets, a thick sheet of enderbite (orthopyroxene-bearing alkali feldspar granite), and other granitoid rocks, and a large body of strongly recrystallized and deformed anorthosite, the Berthé anorthosite. Gabriel domain appears to have had a simple metamorphic history and to have been metamorphosed under P–T conditions similar to those of Canyon domain (Fig. 2). Geochronological studies in Gabriel domain have yielded evidence for a single metamorphic event at 1011–990 Ma (Rigolet pulse; Scott and Hynes 1994), consistent with the apparently simple metamorphic history.

Gneissic layering in Canyon domain generally dips toward the southeast in the north and toward the southwest in the south of the domain, defining a large-scale, inclined, south-plunging fold that envelops the Berthé anorthosite in

Fig. 2. Estimated physical conditions of metamorphism near Lithoprobe line 55. Shaded arrows show approximate P–T trajectories for metapelites from the Hart Jaune, Berthé, Tshenukutish (Island domain) and Gagnon terranes, after Hynes and St-Jean (1997), from metamorphic peak (start of arrow), as recorded by garnet cores, to closure of garnet-consuming, net-transfer reactions as recorded by garnet rims. Shaded boxes show estimated conditions of metamorphism in highest ("Lelukuau 1") and lowest ("Lelukuau 3") Lelukuau, and Tshenukutish terranes, after Indares et al. (2000). Metamorphic conditions estimated with 1996 version (TWQ 202) of TWQ software (Berman 1991) using garnet mixing-model of Berman et al. (1995), plagioclase mixing-model of Fuhrman and Lindsley (1988) and mixing model for biotite of L.Y. Aranovich and R.G. Berman (unpublished). Also shown are positions for the kyanite–sillimanite–andalusite transitions (Holdaway 1971), and generalized continental geotherms for surface heat-flows of 60, 90, and 120 mW m^{-2} (Pollack and Chapman 1977). Note that metamorphism in the Berthé and Hart Jaune terranes took place in the Ottawan pulse (~1060–1020 Ma), whereas that in Gagnon, Berthé and Hart Jaune terranes took place in the Rigolet pulse (~1010–990 Ma).

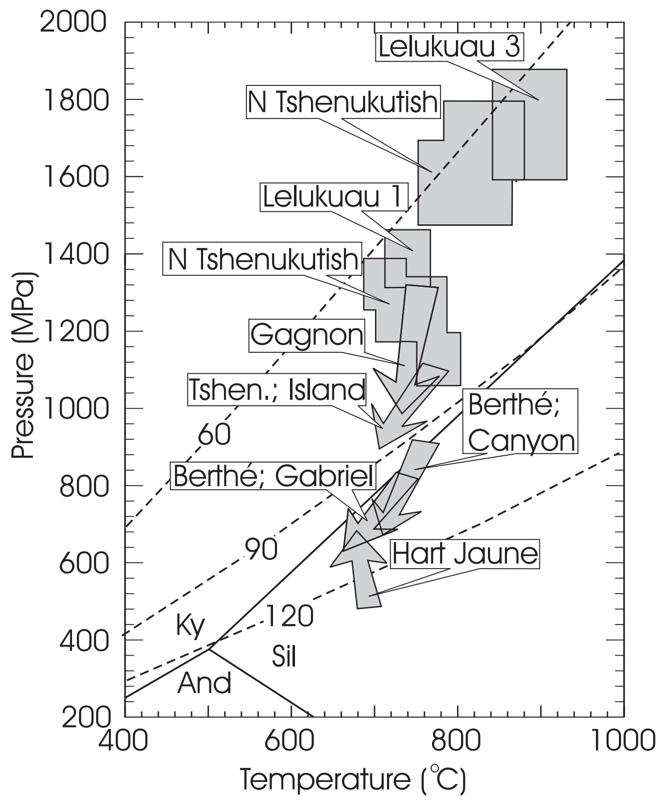
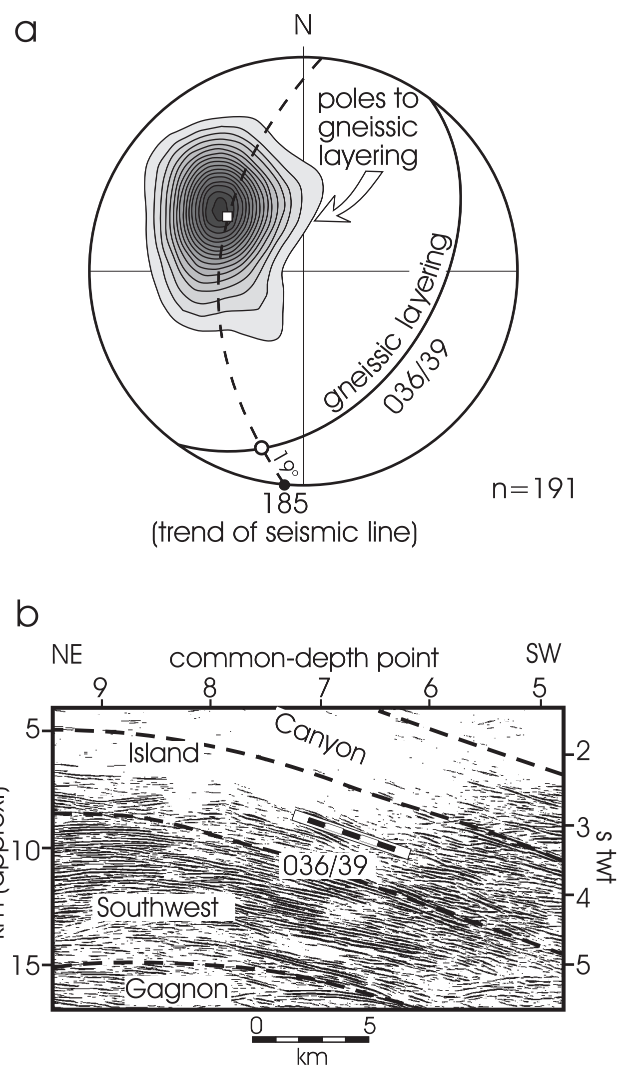


Fig. 3. (*a*) Poles to gneissic layering in the Island domain near common-depth points 6 and 7. White square, maximum eigenvector; solid great circle, gneissic layering corresponding to eigenvector; dashed great circle, seismic section (plane containing maximum eigenvector and trend of seismic line); filled circle, trend of the seismic line; open circle, expected attitude of seismic reflector due to gneissic layering. Lower-hemisphere equal-area projection contoured using Spheristat (Pearce and Stesky 1990). Lowest contour level (lightest shading) shows expected value for randomly distributed data; successively higher (shaded darker) levels step up by 2σ, where σ is statistical dispersion of data (Robin and Jowett 1986). See Fig. 4*a* for region from which structural data were collected. (*b*) Part of unmigrated section for Lithoprobe line 55 between common-depth point 5 and 9. Dashed lines, interpreted positions of major faults; black and white bar, expected attitude of reflector due to gneissic layering in the Island domain, from Fig. 3*a*. Scale selected to give approximately no vertical exaggeration.



the centre of Gabriel domain. The south-plunging sheath folds in metasedimentary rocks of Canyon domain are subparallel to this map-scale structure. The Berthé terrane is thus a map-scale fold nappe, with a core of anorthosite enveloped by the Canyon and Gabriel domains (Fig. 1). It is limited to the west by the HJ–TNSZ and to the north by the Gabriel shear zone (GaSZ, Fig. 1), which separates it from Hart Jaune terrane.

The GaSZ is a folded structure that swings from steeply south-southwest-dipping near Highway 389 to southeast-dipping on islands in the Manicouagon Reservoir farther west. Despite the presence of prominent, south-side-down (i.e., normal sense) kinematic criteria in well-exposed segments of the GaSZ on islands in the Manicouagon Reservoir, marked increases in calculated pressure associated with retrogression of the later granulite-facies mineral assemblages in Hart Jaune terrane are evident as the shear zone is approached, leaving little doubt that rocks of the Gabriel domain were originally thrust onto Hart Jaune terrane (Hynes and St-Jean 1997). The apparent anomaly of amphibolite-facies rocks thrust over granulite-facies rocks is due to the relatively low-pressure character of the granulite-facies metamorphism; peak metamorphic pressures in the granulite-facies Hart Jaune terrane were lower than those in the amphibolite-facies Gabriel domain (Fig. 2; Hynes and St-Jean 1997).

Berthé and Hart Jaune terranes, therefore, form a complex, north-vergent thrust and fold stack with the Berthé anorthosite in its core. The stack was assembled during the Rigolet pulse of the Grenvillian Orogeny, and peak pressures in the stack were apparently no greater that 900 MPa (i.e., medium-pressure metamorphism). There is no evidence for high-pressure metamorphism in either Hart Jaune terrane or the Canyon domain. Only rocks of the MIZ and the upper tectonic units of Gagnon terrane, beneath the HJ–TNSZ, preserve evidence of tectonic burial to depths of greater than 30 km. Line 55 therefore provides an opportunity to determine the disposition of the originally mid-crustal boundary between the high-pressure and medium-pressure terranes.

Structure in three dimensions derived from seismic data

In the region around line 55, there is remarkably good correlation between the attitudes predicted for seismic reflectors, if these reflectors are parallel to the gneissic layering observed at the surface, and the attitudes of the observed reflectors. For example, gneissic layering in the Island domain northwest of common-depth points 7, 8 and 9 on Lithoprobe line 55 has very uniform attitudes, with a mean strike of 036° and a dip of 39°SE (Fig. 3*a*). A reflector derived from this layering should be inclined at an angle equal to the pitch of the layering in the plane containing the pole (normal) to the layering and the trend of the seismic line (see Hynes and Eaton 1999 for a more detailed discussion). Near common-depth point 7, where the seismic line trends at 185°, this pitch is 19° toward the south (Fig. 3*a*). The attitudes of reflectors observed at common-depth point 7 at the appropriate depth almost exactly match this predicted attitude (Fig. 3*b*). In addition, shear zones and faults observed at the surface around line 55 are typically parallel to the gneissic layering in the domains they separate (Hynes and Eaton 1999), so that the attitudes of reflectors are also generally parallel to domain boundaries.

Lithoprobe line 55 follows a sinuous path dictated by the route of Highway 389. It rarely runs perpendicular to the strike of regional gneissic layering, and in some cases (e.g., immediately southeast of the circular part of the Manicouagon Reservoir; Fig. 1), it even runs parallel to strike for considerable distances. In consequence, the majority of the reflections (and reflectivity boundaries) observed are derived from positions outside the vertical section following the highway. Since the structure in the region dips predominantly to the southeast, the true physical positions of most of the observed reflectors are displaced northwest of the line, by distances contingent on the attitudes of the reflectors and the seismic velocities of the rocks between the reflectors and the shotpoints. These true positions can be estimated by assuming a seismic velocity (e.g., 6 km·s^{-1}) for the intervening rocks, and constant dips between features observed at the surface and the same features at the depths at which their equivalent reflectors or reflectivity boundaries are observed on the seismic records.

The actual dip of a feature producing a reflector affects both its apparent dip in the seismic record (Fig. 3*a*) and the shortest distance between the reflector and the seismic line, which becomes the apparent depth of the feature observed in the seismic record. For a given feature imaged at a given common-depth point, increasing the assumed dip, therefore, increases both the predicted apparent dip and the predicted apparent depth of the feature on the seismic record. Since both the apparent dip and the apparent depth of a reflector are constrained by the seismic record, the true dip of a feature that gives rise to a reflector may be uniquely determined if the strike and one surface position of the reflector are known, and its dip is assumed to be constant. In Fig. 4, for example, we illustrate the effect of different assumed dips for the Triple Notch shear zone (TNSZ) and the Cryptic shear zone on their expected positions and attitudes in the seismic record at common-depth point 4. Gneissic layering to the northwest of common-depth point 4 dips uniformly to the southeast, and is parallel to the trends of the shear zones at the surface (Fig. 4*a*). Variation of the assumed dip of the TNSZ, from 35° through 37° to 40°, effectively brackets the interpreted position of the shear zone at common-depth point 4, and indicates that a dip of 37° provides a good fit to the position of the shear zone and the attitudes of seismic reflectors in its vicinity (Fig. 4*b*). The Cryptic shear zone is likewise constrained to have a dip between 35° and 40°. We have used this procedure to estimate the attitudes of subsurface features responsible for the seismic reflectors throughout Lithoprobe line 55 (Hynes and Eaton 1999). The procedure provides structure contours on the surfaces of shear zones, for the regions between their surface exposures and the positions at which they are sampled seismically (e.g., Fig. 4*a*). The departures of the subsurface positions of reflectors sampled by the seismic survey from the vertical section is well illustrated on Fig. 4*a*, where the point at which the TNSZ is sampled at common-depth point 4, for an assumed subsurface dip of 37°, is displaced 10 km from the seismic line. This amply demonstrates the dangers of treating seismic records for structurally complex regions as sections.

With the seismic data for line 55 and the surface geology of Fig. 1, we can satisfactorily model the subsurface positions of all the major boundaries observed at the surface to the west and northwest of the line, to seismic depths of approximately 8 s two-way time (~24 km) between common-depth points 1 and 30 (Fig. 5). North of common-depth point 30, the shallow regional and highly variable attitudes of rocks of the Knob Lake Group limit the utility of these procedures. A detailed justification for the positions of boundaries depicted on Fig. 5 is given elsewhere (Hynes and Eaton 1999). Here, we confine ourselves to summarizing the most important features.

The seismic record for line 55 is dominated by a major "antiform" centred near common-depth point 10, and a "synform" centred near common-depth point 19 (Fig. 5). These features are largely artifacts due to northwestward curvature of the road near common-depth points 9 and 10, and southeastward curvature of the road near common-depth

Fig. 4. (*a*) Surface geology between common-depth point 4 and common-depth point 10. Ornament as for Fig. 1. Circled numbers, shotpoints; structural symbols, mean attitudes of gneissic layering; evenly-spaced solid black lines, structure contours on Triple Notch shear zone in subsurface; filled circle labelled "CDP4," position in subsurface at which Triple Notch shear zone is imaged by seismic survey (Fig. 4*b*) (*b*) Portion of line 55 (unmigrated) near common-depth point 4. Barber's poles, modelled positions and attitudes of reflectors associated with Triple Notch shear zone, for dips of 35°, 37° and 40°, and strike of 060, labelled "TNSZ35", "TNSZ37" and "TNSZ40", and of Cryptic shear zone (CSZ), for dips of 35° and 40°, also for a strike of 060, labelled "CSZ35" and "CSZ40"; heavy dashed lines, interpreted trajectories of faults.

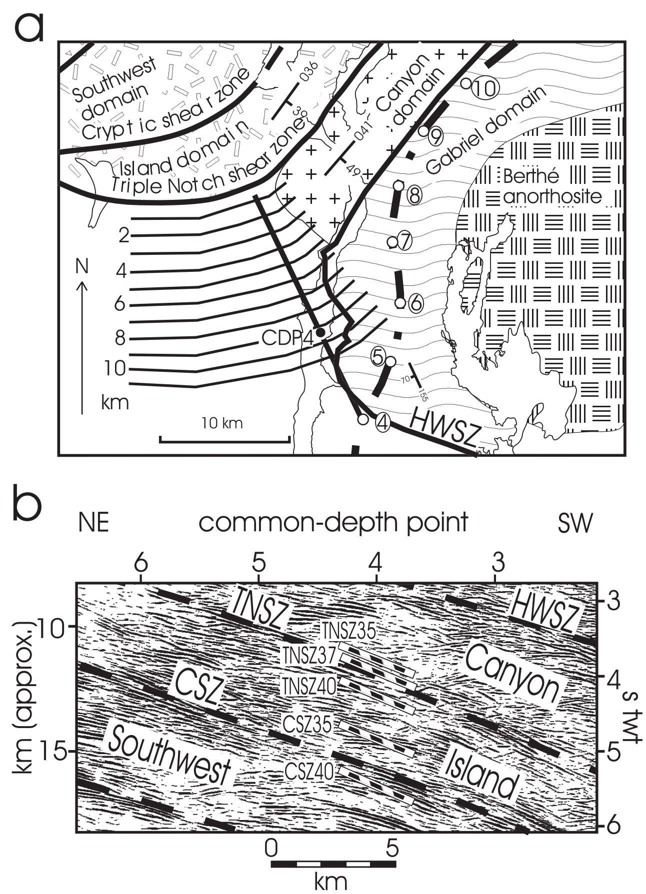


Fig. 5. The unmigrated seismic record for line 55 with the interpreted positions of the major faults marked by heavy dashed lines. The heavy solid line marks the interpreted position of the décollement at the base of Gagnon terrane.

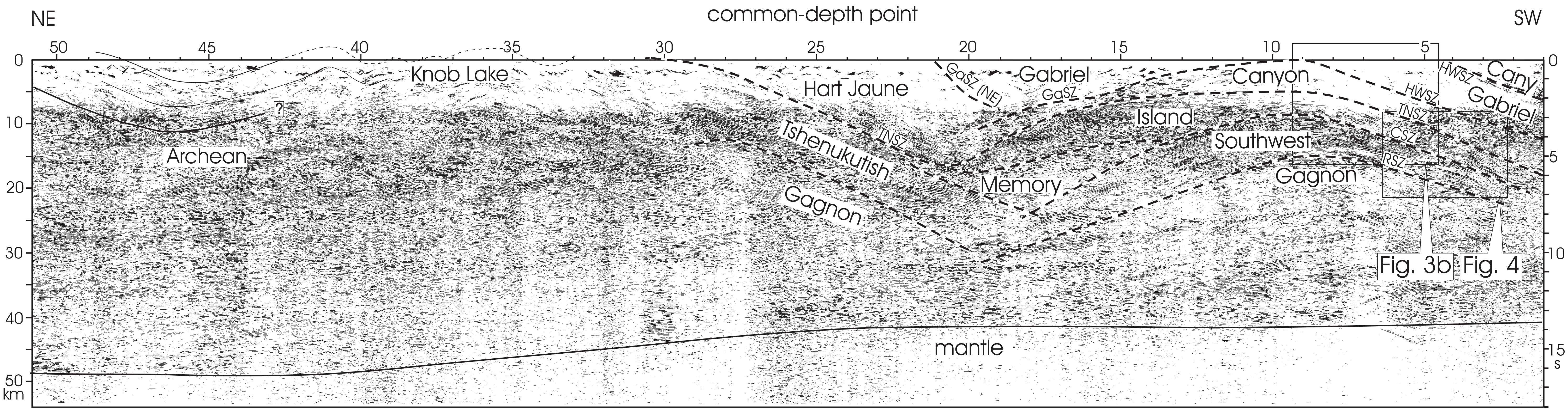
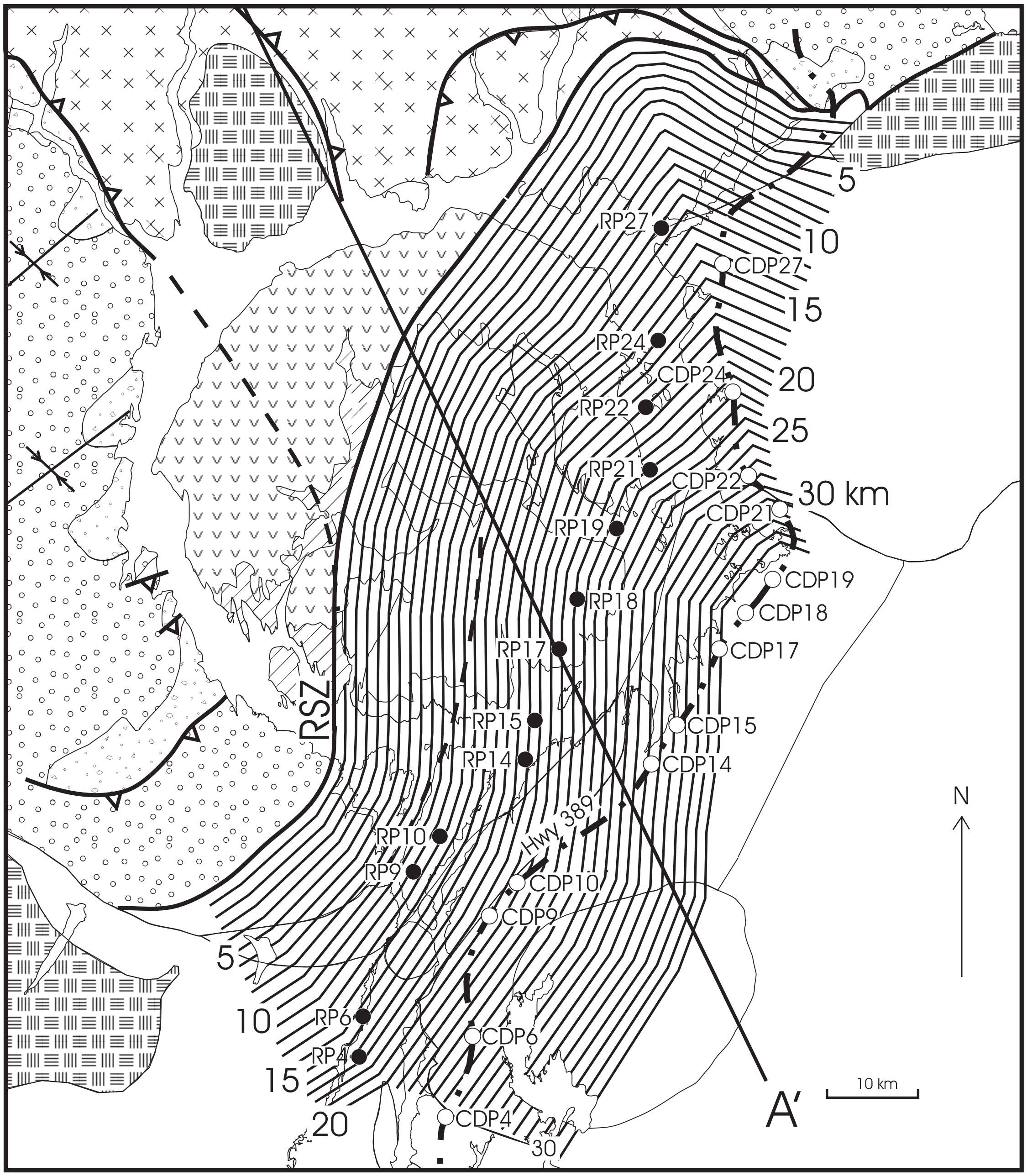


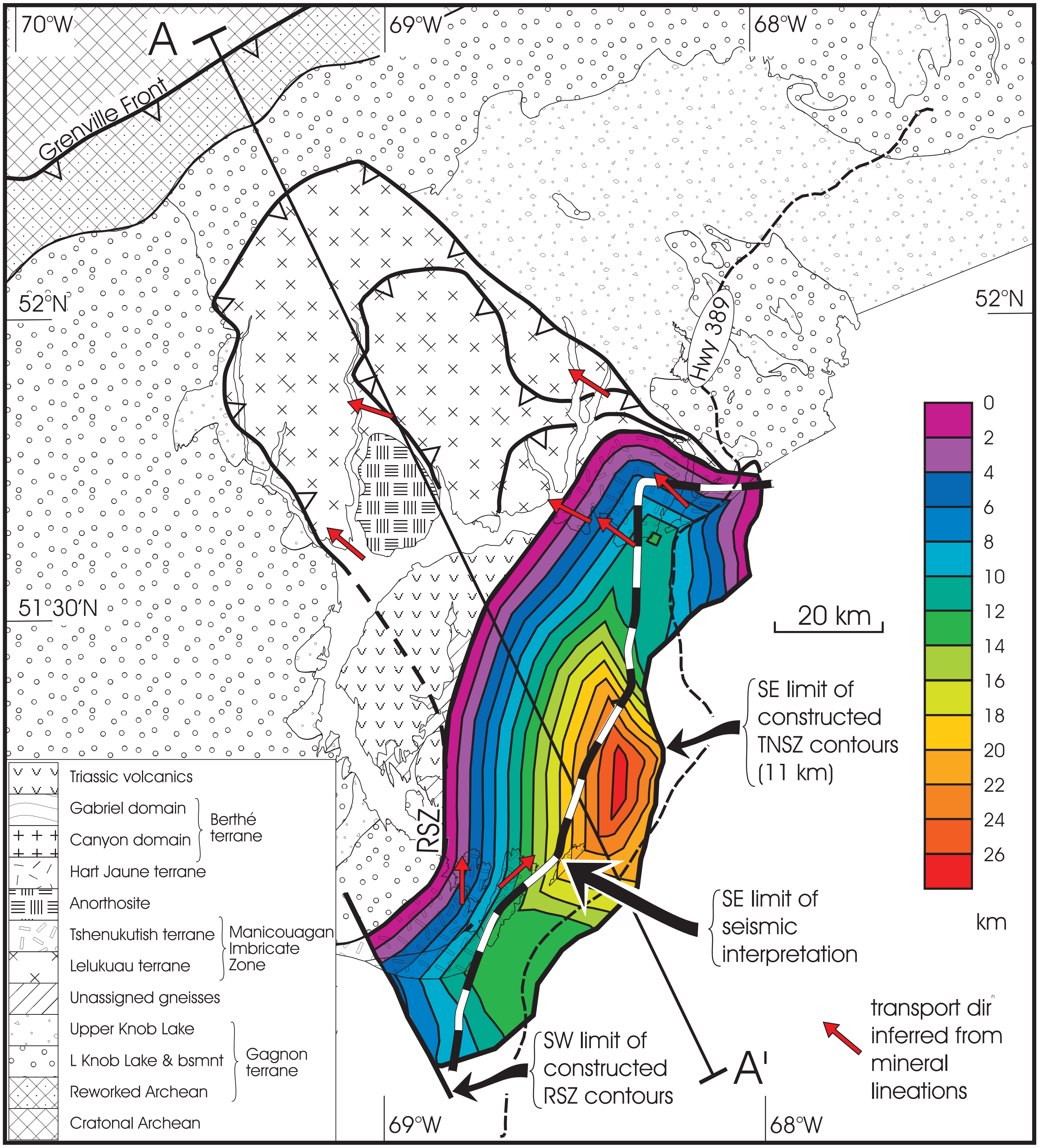
Fig. 6. Structure–contour map for the Relay shear zone. Solid black lines, contours at 1 km intervals. Filled circles labelled "RP 10", etc. show positions in subsurface from which reflectors are sampled from shotpoints "common-depth point 10", etc. Ornament for bedrock geology as for Fig. 1.



points 19 and 20. The curves in the road result in seismic sampling of a southeast-dipping surface at relatively small distances from the road at common-depth point 9 and greater distances at common-depth point 19. A structure-contour map for the Relay shear zone (Fig. 6) derived from the seismic data in the fashion outlined above, on which the positions at which the shear zone is sampled by the seismic survey are marked, shows how much closer these positions are to the road for common-depth point 10 than for common-depth point 19. Our modelled structure contours for the Relay shear zone require only limited curvature of the surface.

The HJ–TNSZ dips moderately southeast in the north and south, and more steeply to the east between common-depth point 14 and common-depth point 22 (see Hynes and Eaton 1999 for details). The total preserved thickness of Tshenukutish terrane between its surface exposure and Highway 389 may be assessed from the differences in depth of the HJ–TNSZ, which overlies it, and the Relay shear zone, which underlies it (Fig. 7). In the subsurface, the Tshenukutish-terrane rocks occur in a north-northeast-trending lozenge-shaped block, with maximum vertical thicknesses exceeding 26 km. The form of the easternmost part of the lozenge (east of the dashed line on Fig. 7) is not well constrained, because it is derived from extrapolation of the structure contours for the Relay shear zone beyond the depths at which it is imaged on line 55. The lozenge-like shape of the block is suggested, however, even by the

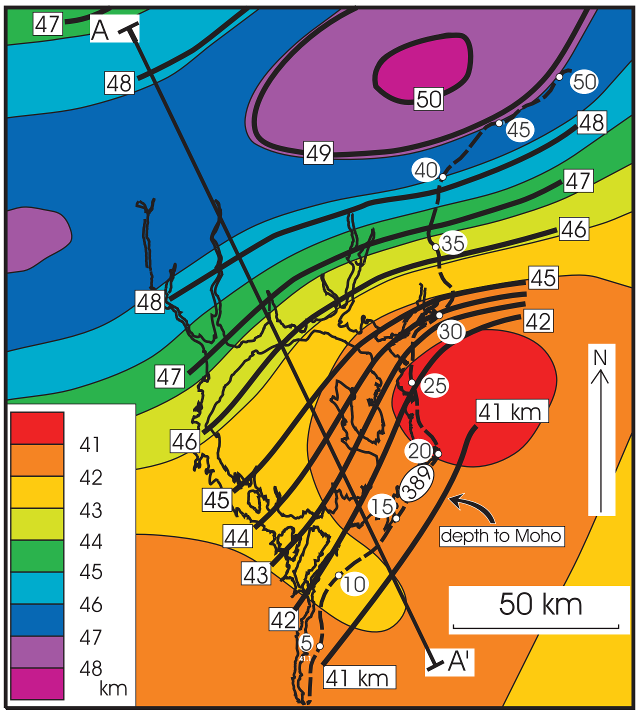
Fig. 7. Isopach map for Tshenukutish terrane. Heavy dashed line, southeastern limit of structure contours whose position is constrained by observed reflectors. To the southeast of this line, structure contours on the Relay shear zone were constructed by linear extrapolation; the isopachs are therefore less well constrained than to the northwest. Red arrows show inferred transport directions for rocks within Tshenukutish terrane, based on the predominant trends of mineral lineations (Indares et al. 2000; and unpublished field data).



isopachs for the region northwest of the positions of the Relay shear zone imaged on line 55, where no extrapolation was necessary to construct the isopachs.

Surface geology indicates that rocks of the Lelukuau terrane are pinched out between those of Tshenukutish terrane and Gagnon terrane northwest of common-depth point 31 and, despite the very broad exposure of such rocks to the northwest of the Manicouagon Reservoir, they also pinch out to the southwest before the Relay shear zone intersects the shoreline in the southwest of the Manicouagon

Fig. 8. Calculated Moho geometry near Manicouagan Reservoir. Colours, estimated depth to the Moho calculated from Decade of North American Geology 6 km gridded Bouguer gravity data using methods of Zheng and Arkani-Hamed (1998); thick solid lines, interpreted contours on Moho at 1 km intervals. Their values on Highway 389 are constrained by the measured Moho depths on line 55; their regional trends are parallel to those calculated from the gravity inversion, except east of the Manicouagon Reservoir, where the gravity field is strongly influenced by the high-density rocks of Hart Jaune terrane; and the Moho is constrained by the seismic data to have little relief.



Reservoir (Fig. 1). The lobe of Lelukuau terrane is, therefore, limited to a region adjacent to Tshenukutish terrane. We believe the broad lobe of Lelukuau terrane is a thin lip-like feature in front of and only partly beneath Tshenukutish terrane (see also Eaton et al. 1995). Such an interpretation is compatible with the regionally shallow dips of structures within Lelukuau terrane and with observed structural relationships in the northeast of the Manicouagon Reservoir.

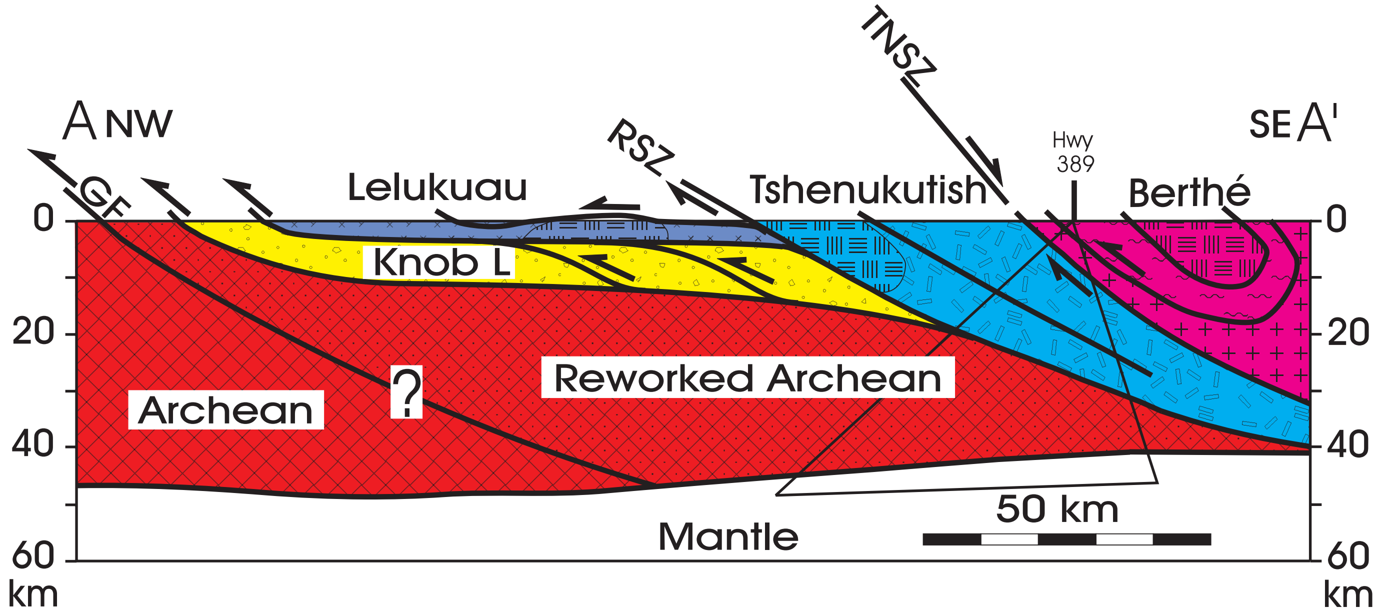
North of common-depth point 32, line 55 lies entirely within rocks of Gagnon terrane, where it crosses the axes of the major, northwest-trending cross-folds at high angle (Fig. 1). The mapped regional antiforms and synforms, defined by the stratigraphy of the Knob Lake Group, are clearly visible on line 55 and are evident in arrivals as late as 3 s two-way time (Fig. 5). Archean basement cannot be clearly distinguished from Knob Lake metasedimentary rocks on the seismic record, probably because the metasedimentary rocks are interleaved with basement rocks in many places and because the metamorphic grade in the metasedimentary rocks is high, so that reflection coefficients with gneissic-basement contacts may not be significant. At the northeastern end of line 55, an approximate position for a décollement at the base of the thrust stack may be drawn beneath a zone of coherent reflectors on the migrated seismic section (Hynes and Eaton 1999), but the disposition of this surface farther southeast is unclear, and it is not identified on Fig. 5.

The base of the crust

There is a marked decline in reflectivity that can be traced continuously from 13.5 s two-way time at common-depth point 1 s to more than 16 s two-way time at common-depth point 50 (Fig. 5). Although this decline is not itself marked by a prominent reflector, we interpret it to represent the base of the crust. We have no evidence for substantial gradients in total crustal thickness perpendicular to the line, and although there is clearly a gradient parallel to the line, it is sufficiently low to have little effect on the development of out-of-vertical-plane reflections. We have, therefore, assumed that the depth to this decline in reflectivity is a measure of crustal thickness immediately beneath the shotpoints. Assuming a uniform velocity of 6 km·s^{-1} for the crust, its thickness is relatively constant at ca. 41 km between common-depth point 1 and common-depth point 22, then increases to a thickness of ca. 48 km between common-depth point 22 and common-depth point 41, after which it is again constant (Fig. 5). Line 55, therefore, crosses a ramp in the Moho.

The seismic data provide no information concerning the three-dimensional form of the Moho, but the gravity field of the Grenville Province in eastern Quebec and western Labrador is characterized by a prominent gravity low (the "Grenville Front gravity low") that trends east-northeast, parallel to the structural grain, and that has been interpreted as due to crustal thickening beneath the Grenville Front (Berry and Fuchs 1973; Thomas and Tanner 1975; Rivers et al. 1989; Pilkington 1990; Hynes 1994). The data from line 55 are consistent with this interpretation. The apparent dip of the Moho on line 55 (Fig. 5) is shallow, but this is largely due to the obliquity of the line. Crustal thickness increases more than 2 km between common-depth points 28 and 32, where the road is trending northeast, at a very shallow angle to the regional trends of the gravity field. If structure on the Moho maintains its east-northeast trend through this region, the slope of the Moho must be particularly high here. In Fig. 8, we have sketched a contour map for the Moho in the region around the Manicouagon Reservoir, derived from regional inversion of the long and intermediate wavelength features of the Bouguer gravity field for the entire Grenville Province, using the methods of and Zheng and Arkani-Hamed (1998). For this inversion, the anomalies in the gravity field were assumed to arise exclusively from Moho relief, and the density contrast between the lower crust and the underlying mantle (250 kg·m^{-3}) was chosen to provide the closest fit to the relief on the Moho evident in the seismic data. Contours on the Moho derived in this way are influenced by local variations in crustal density. In the region of Fig. 8, in particular, they are influenced by the high densities of rocks in Hart Jaune terrane, which give rise to local apparent shallowing of the Moho. The regional model for the gravity provides support for the 4 km deepening of the Moho between common-depth point 32 and common-depth point 51, but does not reproduce the steep ramp between

Fig. 9. Vertical section along line AA′ (Fig. 1, Fig. 8). Depths of boundaries within the triangular field beneath the surface trace of Highway 389 are constrained by the structure–contour maps derived from interpretation of line 55. Outside that region, the subsurface structure is based on extrapolation parallel to the regional tectonic trend of the Grenvillian Orogen. Colour and ornament is as for Fig. 1.



common-depth point 28 and common-depth point 32, probably because of the small wavelength of the latter feature. The ramp derived from the gravity inversion exhibits a clear north-northeast trend west of Manicouagan Reservoir (Fig. 8), in contrast to the more typical east-northeast trend of the feature along the Grenville Front. Direct support for this kink is provided by the seismic data, which indicate negligible change in crustal thickness between common-depth point 5 and common-depth point 20 (Fig. 5).

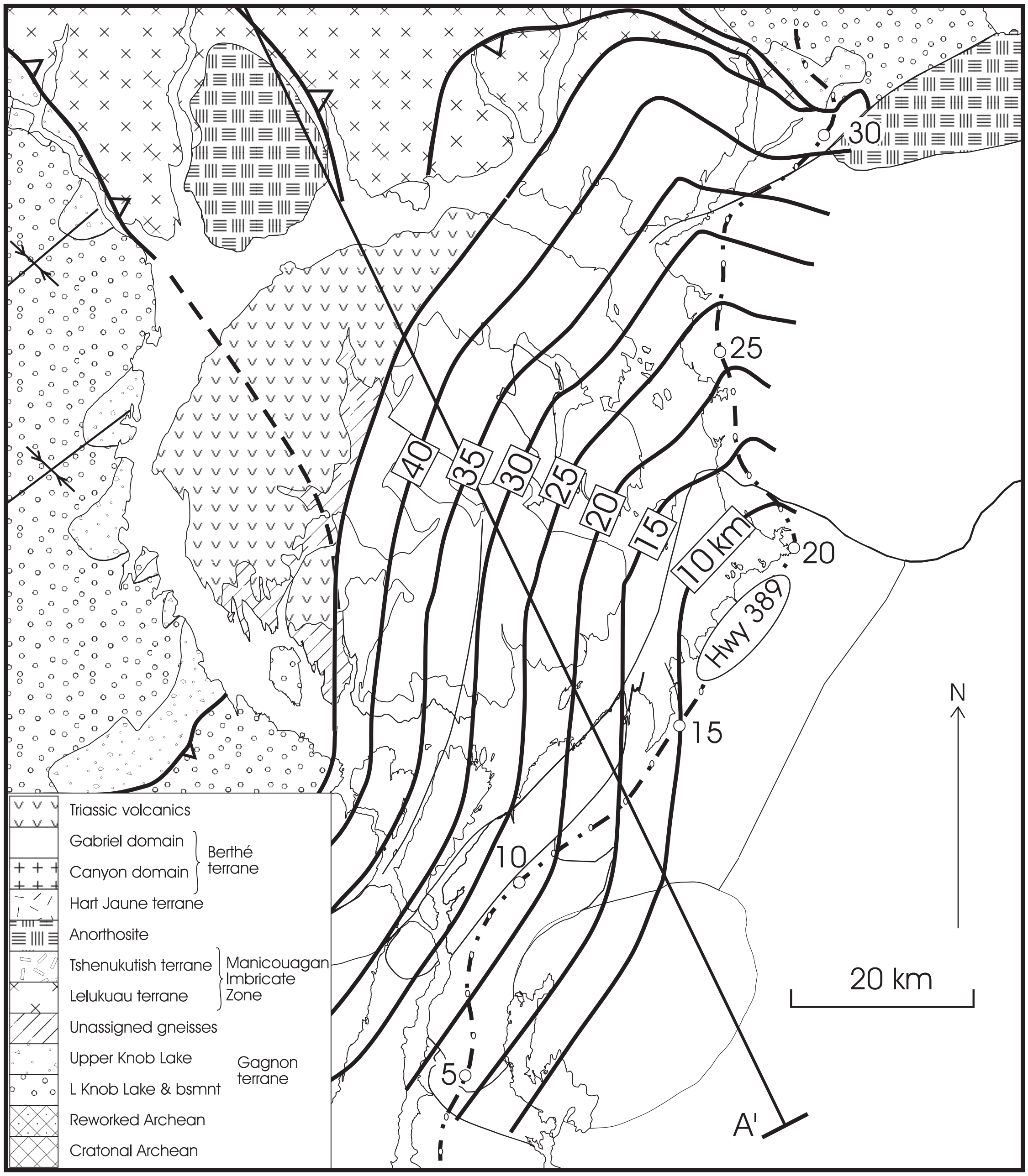
The flattening of the Moho between common-depth point 41 and 50 that is evident on line 55 suggests that the maximum thickness of the crust had been achieved by the northern end of line 55. In fact, however, line 55 runs subparallel to the regional trends of the gravity field in this region, and the deeper parts of the Grenville-Front gravity low lie northwest of the termination of the line (Fig. 8). It is probable, therefore, that crustal thickness beneath the Grenville Front in the Manicouagan region is larger than the 48–49 km estimated at common-depth point 50. If all the relief in the gravity field is due to structural relief at Moho depth, the further crustal thickening required is on the order of 1 km, giving crustal thicknesses of 49–50 km just south of the Grenville Front.

Implications for relationship of deep structure to emplacement of high-pressure rocks

Line 55 itself (see Fig. 5) provides a poor illustration of the subsurface structure of the northern Grenville Province in eastern Quebec, because overall it is highly oblique to the regional structure, due to the circuitous route followed by the road, and because much of the signal on the line is from outside the vertical plane containing the road. The structure–contour maps we have derived from the line, however, permit the construction of vertical sections. Because of the obliquity of the line to the regional structural trends, a section that well illustrates the structure of the region necessarily departs significantly from the trend of the line, and there is the additional problem that some of the trends in the Manicouagan region are locally more northerly than parallel to the regional east-northeasterly trends of the orogen. Nonetheless, a section trending north-northwest, from the Berthé anorthosite through the broadest embayment of the MIZ, illustrates many of the important features of the Grenville of eastern Quebec that are apparent from the seismic analysis (Fig. 9). On this section, only the region in the triangular zone beneath the surface position of Highway 389 is rigorously constrained by the structure–contour maps. Outside this zone the characteristics of the section are derived from extrapolation parallel to the regional tectonic strike.

Gagnon terrane is shown on Fig. 9 with a nearly constant structural thickness of approximately 10 km over a considerable distance across the regional strike, in conformity with the evidence from the northeastern part of line 55. The crust beneath the northern Parautochthonous Belt is shown thickening to 49 km beneath the Grenville Front gravity low. The outstanding feature of the section, however, is the prominent ramp beneath Tshenukutish terrane up which the MIZ must have been transported. The combination of this ramp and the southward shallowing of the Moho suggests that the Archean basement terminates in the subsurface somewhere near the southeastern end of the section (see also Eaton et al. 1995). Isopachs for the crust underlying Tshenukutish terrane (Fig. 10) show that this dramatic tapering of the Archean crust is in a more easterly than southeasterly direction in the vicinity of Manicouagan Reservoir. This architecture would have produced a marked lateral ramp during northwest-directed tectonic transport, as is inferred for the

Fig. 10. Aggregate thickness of crust beneath Tshenukutish terrane. Thick solid lines, contours, at 5 km intervals, for the total thickness of crust beneath rocks of Tshenukutish terrane. The figure was derived from Fig. 6 and Fig. 8. It provides an estimate of the maximum thickness of pre-Grenvillian crust. Ornament as for Fig. 1.



Grenvillian Orogeny in eastern Quebec and western Labrador (Rivers et al. 1993; Connelly et al. 1995). We do not have control on the nature or regional disposition of this ramp outside the area covered by Fig. 10, but there is independent evidence that Archean rocks do not extend much farther east than Gagnon terrane (Rivers 1997). This may therefore be one of a series of lateral ramps as the southern boundary of the Archean (Superior Province) crust swings toward the Grenville Front in western Labrador. Both the ramp and the bend may have been critical to the emplacement and preservation of the high-pressure rocks of the MIZ. It is also clear that Gagnon terrane was substantially thickened during emplacement of the MIZ as a thrust wedge on top of it, and that the southern parts of Gagnon terrane were transported into their present positions from considerable depths (see also Rivers et al. 1993).

Jamieson and Beaumont (1988) modelled the efficacy of crustal ramps in promoting the rapid exhumation of deeply buried rocks. Although their models were designed to illustrate overthrusting at a simple rifted margin, the high-pressure rocks of the MIZ and the subjacent Knob Lake Group exhibit essentially the same spatial relationship to a crustal ramp as predicted in their model. Their model was concerned primarily with the exhumation of material from the deeper part of an accretionary wedge, but it does not preclude contemporaneous accretion to the wedge during uplift (cf. Chemenda et al. 1995). In this case, it is inferred that rocks of the Knob Lake Group, together with the upper part of their Archean basement, were progressively accreted to the base of the thrust wedge as it was unroofed, giving rise to the northwest vergence of Gagnon terrane (Rivers et al. 1993).

The broad embayment of the MIZ northwest of Manicouagan Reservoir is directly adjacent to the thick lens of Tshenukutish rocks in the subsurface (Fig. 7) and it is in this embayment that the formerly most-deeply buried rocks are preserved (Indares et al. 2000). As illustrated in Fig. 10, the thick lens of Tshenukutish terrane requires a substantial lateral ramp in the underlying reworked Archean (and possibly also Knob Lake Group cover). The lateral ramp may have provided a channel through which upward flow of the rheid lower crust was favoured, and perhaps a local pressure shadow that enhanced the flow.

Direct support for a link between the lateral ramp and broad embayment of the MIZ is provided by the attitudes of lineations in the embayment. These lineations are indicative of flow trajectories to the west-northwest within the embayment (Fig. 7), consistent with flow of material into the embayment from the position of the lateral ramp.

Grenvillian tectonic evolution in eastern Quebec

The metamorphic contrast in the Manicouagan region across the HJ–TNSZ is a feature of regional significance. Even after substantial unroofing, the rocks to the northwest of the shear zone reequilibrated at pressures at least 200 MPa higher than those to the southeast (Fig. 2). In many places, the contrast is considerably greater (600–1000 MPa). Kinematic evidence for early thrust displacements on the HJ–TNSZ is cryptic, but late displacements can be unequivocally shown to have been extensional in character. This late normal faulting is an important feature of the current architecture of the orogen, and rebound of the footwall to the northwest of the HJ–TNSZ could explain part or all of the southward shallowing of the Moho evident on Fig. 9. It is clear, however, that the late normal displacement on the HJ–TNSZ was not responsible for much of the unroofing of the high-pressure rocks of the MIZ, which must have occurred immediately following their attainment of peak metamorphic conditions during the Ottawan pulse of the Grenvillian Orogeny.

Estimated peak metamorphic temperatures were >800°C in both the footwall and hanging wall of the HJ–TNSZ (Fig. 2), exceeding Pollack and Chapman's (1977) estimate for a continental geotherm for a surface heat flow of 90 mW m^{-2}. These high temperatures, and the presence of metamorphic ages ranging from 1050 to 990 Ma, are incompatible with a simple model of crustal thickening and thermal relaxation. In addition, the high temperatures in the MIZ cannot be explained by a two-stage thickening model, involving initial burial during the Ottawan pulse followed by a second period of tectonic burial during the Rigolet pulse, as the geochronological data indicate that the peak temperatures in the MIZ occurred during the earlier Ottawan pulse. Indares et al. (1998) suggested that the unusually high temperatures in the MIZ were a result of asthenospheric delamination beneath a thickened crustal welt, citing the presence of synmetamorphic mafic intrusions in Lelukuau terrane in support of this contention. Whatever their cause, the high temperatures must have significantly enhanced the ductility of the deeply buried rocks of the MIZ, thereby reducing their resistance to buoyancy-driven upward flow (cf. Hsü 1991).

In Fig. 11 we present a sequence of figures, developed through retrodeformation of the section in Fig. 9 while maintaining area balance, to illustrate our conception of the tectonic evolution of this part of the Grenvillian Orogen. Before the onset of the Ottawan pulse of the Grenvillian Orogeny (ca. 1100 Ma; Fig. 11*a*), Archean basement in the northwest tapered toward the southeast and was overlain by a thin cover of the undeformed Middle Paleoproterozoic Knob Lake Group. To the southeast, Archean and Middle Paleoproterozoic rocks were overlain, probably tectonically, by a terrane composed of Labradorian (Late Paleoproterozoic) and some Pinwarian (Early Mesoproterozoic) rocks, of which the MIZ formed a part. In the position of our section, most of this terrane has subsequently been removed so that its former character is uncertain, but by analogy to the Labradorian terranes farther east (e.g., Gower et al. 1992), it probably consisted of intrusive rocks with accreted supracrustal arc – back-arc sequences in the upper crust. The Labradorian rocks are interpreted to have been overlain to the southeast, also probably tectonically, by a Pinwarian plutonic terrane with accreted supracrustal arc – back-arc sequences. In the line of our section, this terrane has been completely removed in subsequent events (see below), but Hart Jaune terrane is probably a remnant, preserved to the northeast of the cross sections of Figs. 9 and 11. The initial configuration of Berthé terrane is uncertain; it is depicted here as an outboard region of inferred Pinwarian age, but there are as yet no geochronological data confirming this, although there is some evidence for Pinwarian-age rocks farther to the southeast (Davidson 1995; Rivers 1997; Dickin 2000).

During the Ottawan pulse of the Grenvillian Orogeny (ca. 1050 Ma), part of the Laurentian margin was tectonically buried to depths of 60 km or more and then rapidly unroofed. The previously undeformed Middle Paleoproterozoic crust and its Archean basement, together with the accreted Pinwarian and Labradorian crust, experienced high-pressure metamorphism. Hart Jaune and Berthé terranes, however, also composed of Pinwarian crust, do not exhibit evidence of this high-pressure event. We interpret the tectonic burial and high-pressure metamorphism to be a result of progressive northwest-directed imbrication of Tshenukutish and Lelukuau terranes and the southeastern edge of Gagnon terrane under a tectonic cover derived from the southeast. At least double crustal thicknesses were achieved (pressure ~ 1800–2000 MPa in Lelukuau terrane, Fig. 2), but there is no evidence to suggest that the crust was subducted into the mantle (Indares et al. 1998). Berthé and Hart Jaune terranes were not penetratively deformed or metamorphosed and were probably located high in the structural pile at this time. They may have comprised part of the tectonic load on the high-pressure terranes. The unusually high metamorphic temperatures at the base of the crust during the high-pressure event (800–850°C in MIZ) were interpreted by Indares et al.

Fig. 11. Panel diagram showing the tectonic evolution of the northern Grenville Province of eastern Quebec. Colour and ornament as for Fig. 1, except for units of inferred former crust that was subsequently eroded. This is shown as green and grey stipple on either side of Berthé terrane. It was probably of Pinwarian age. Other subsequently eroded material in the earlier panels is stippled, but has the colour of the terrane to which it belongs. Faults that will be active in the next frame are marked by dashed lines; faults that are or have been active are shown as thick solid lines. Faults active in a given frame are labelled with arrows. Note that Fig. 11*e* is derived from the present structural section (Fig. 9) with only the addition of inferred faults in the reworked Archean. The panels depicting earlier stages in the tectonic evolution were derived from this section, preserving area balance for different units between sections. Rocks of the Manicouagan Imbricate Zone were in all likelihood transported into the plane of section from outside the plane of Fig. 11*c* (up a lateral ramp), rather than having originated within it as implied by their positions on Figs. 11*a* and 11*b*.

(1998) to be a result of replacement (convective removal?) of the thickened root of the subcontinental lithosphere by asthenosphere, allowing mantle heat and magmas access to the base of the crust.

Burial of the MIZ was short-lived, since before onset of the Rigolet pulse of the Grenvillian Orogeny (ca. 1010–990 Ma) rocks of Lelukuau terrane had been transported to shallow structural levels far to the northwest. Uplift and cooling of the thermally weakened high-pressure rocks into the mid-upper crust by about 1020 Ma is inferred to have occurred by extrusion of Tshenukutish and Lelukuau terranes, and deeply buried parts of Gagnon terrane, between the footwall shear zone (within Gagnon terrane) and a hanging-wall normal fault at the base of the allochthonous terranes (Fig. 11*c*). Extrusion is compatible with the numerical models of Beaumont et al. (1996) for the uplift and unroofing of deeply buried crust. The role of the structural ramp comprising the relatively competent Archean crust and its ductile Paleoproterozoic cover in channelling uplift toward the northwest has been discussed by Indares et al. (1998). Movement was concentrated along several major ductile shear zones up to a kilometre wide, some of which are located between contacts of the terranes and are marked by metamorphic contrasts. In addition, there are several major shear zones within Gagnon terrane and the underlying Archean basement gneisses. Inferred coeval normal faulting in the hangingwall would have brought the little-deformed Berthé and Hart Jaune terranes into contact with high-pressure rocks of Tshenukutish terrane at this time. Evidence for this phase of normal faulting is cryptic in both Berthé and Hart Jaune terranes, but is inferred on geometric grounds, and it may correspond to normal displacements evident in Tshenukutish terrane (Indares et al. 1998). There are few constraints on crustal thicknesses in the region after extrusion of the MIZ. In Fig. 11*c*, we depict a uniform crustal thickness of approximately 50 km, consistent with the attainment of gravitational equilibrium following widespread thermal weakening in the lower crust.

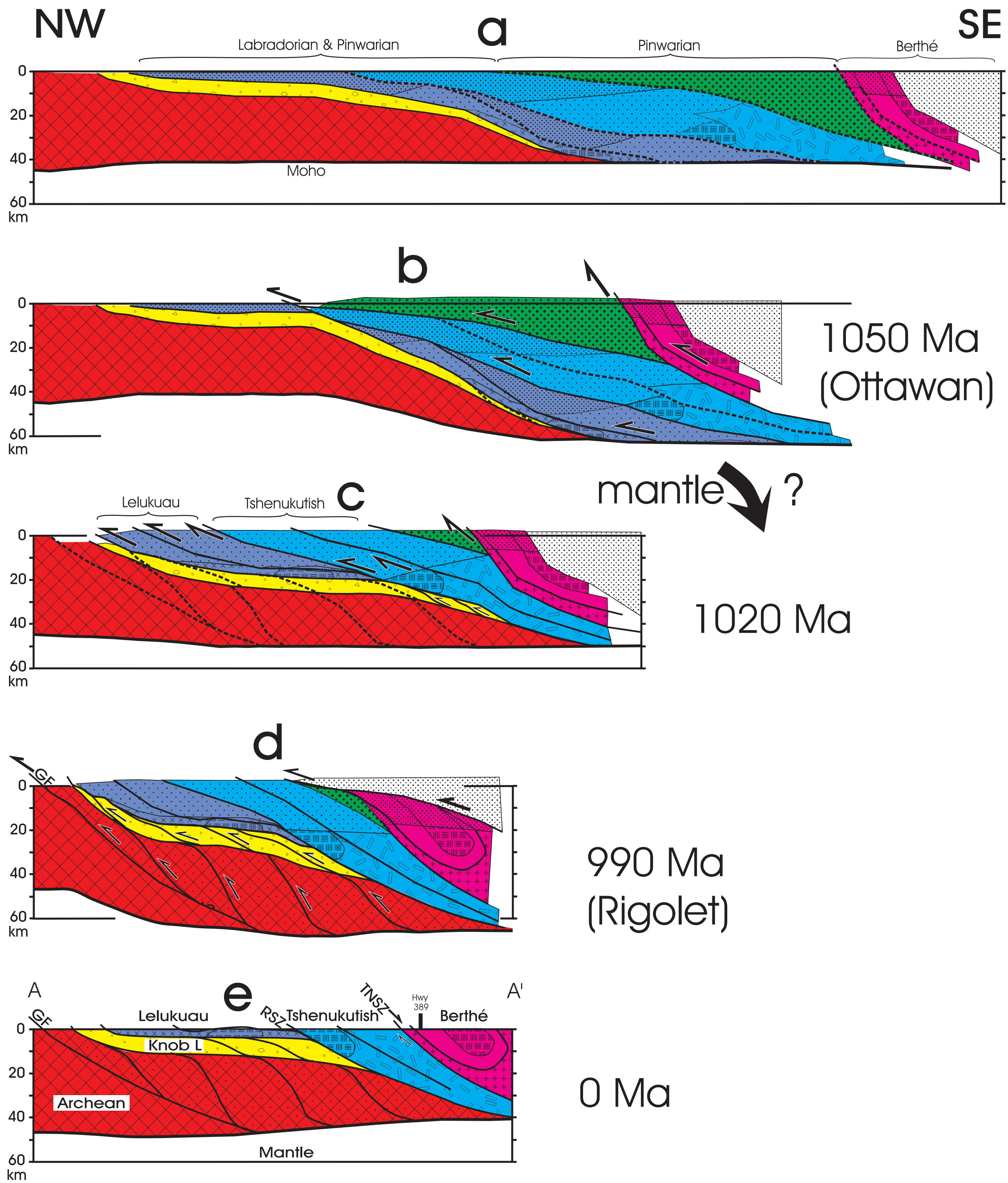
Extrusion was followed by a second phase of crustal thickening in the Rigolet pulse, leading to metamorphism at around 1010–990 Ma, in two separate parts of the orogen (Fig. 11*d*). In the southeast, the high-pressure terranes of the MIZ were overthrust by an outboard terrane of uncertain, but probably Pinwarian, age and a highly ductile, mid-crustal nappe, the Berthé nappe, also probably of Pinwarian age, beneath it. The Berthé nappe was characterized by medium-pressure metamorphism (pressure ~ 600–900 MPa, Fig. 2). There is no evidence for penetrative deformation of the high-pressure rocks of the Lelukuau and Tshenukutish terranes that were in the footwall of the nappe at this time. They appear to have acted as a relatively cool and rigid buttress, against which the overlying more ductile rocks of Hart Jaune and Berthé terranes were transported and deformed. The northwest vergence and southeast dips of Berthé terrane are compatible with renewed northwest-directed tectonic transport, with the MIZ and its footwall again forming a structural ramp, up which the ductile rocks of Berthé terrane were transported. Sheath-fold formation in Berthé terrane implies extremely high strains and may be a result of strain partitioning against the subjacent buttress. Coeval northwest-directed thrusting also took place in the northern part of Gagnon terrane, structurally beneath the MIZ, leading to development of the Grenville Front foreland fold–thrust belt. This involved further accretion of Paleoproterozoic rocks and their Archean basement of Gagnon terrane into the base of a northwest-propagating thrust wedge (van Gool 1992; Rivers et al. 1993). Following this thickening episode, the crust is inferred to have had a relatively constant thickness of more than 65 km over a broad region.

The relatively high geothermal gradient that is implied by the medium-pressure metamorphism (90–120 mW m^{-2}, Fig. 2), and the lateral spreading of the orogen toward the northwest, are both compatible with a continuing elevated mantle heat flow. Although no mafic magmatism of this age has yet been identified, none of the mafic intrusions in the allochthonous terranes of southeastern Quebec, including the Berthé anorthosite, has yet been dated.

The final tectonic episode that gave rise to the present configuration of terranes in the Manicouagan area involved late extensional faulting (Indares et al. 1998). This was probably not significant on a regional scale, but resulted in truncation of the high-pressure terranes of the MIZ by the normal-sense HJ–TNSZ, which further dropped Hart Jaune, Berthé and associated medium-pressure terranes to the south by about 6 km.

Conclusions

Lithoprobe line 55 has provided unusually good correlations of the attitudes of gneissic foliations at the surface to the attitudes of seismic reflectors, leaving little doubt that many of the reflectors in the upper half of the crust are parallel to gneissic layering. The structurally highly oblique and circuitous path of line 55, which was governed by limitations of the road network and was regarded at the outset of this Lithoprobe project as a distinct disadvantage, has permitted the delineation of the subsurface structure in this region over a considerable area. Most particularly, it has permitted the identification of a large-scale lateral ramp in the subsurface, the presence of which was probably the principal reason for the widespread presence of formerly deeply buried rocks at high structural levels in the Manicouagan re-



gion. The close association of the high-pressure rocks with the subsurface ramp and the tendency for lineations in the MIZ to trend east-southeast (Indares et al. 2000) at a high angle to the ramp, in contrast to southerly to southeasterly trends elsewhere in the region, illustrates the importance of the ramp in controlling the deformation movement-pattern. It also suggests criteria that may permit the identification of lateral-ramp structures at depth elsewhere in the Grenville Province and in other deeply exhumed orogens.

The out-of-the-vertical character of many of the reflections observed on line 55 dictates that a line drawing from the reflection record is far removed from a vertical section and is, in fact, not a section at all. Many of the large structures apparent on the line drawing are, furthermore, artifacts of the disposition of the line. Only in the northern part of the line, where its trend runs perpendicular to regional strikes for considerable distances, does the seismic record bear a simple relationship to a structural section. Such problems are to be expected in regions of structural complexity and where road access is not optimal. Some of these problems may be circumvented, and perhaps even used to advantage, as was the case here, but they also demonstrate the dangers of seismic interpretation in structurally complex terranes without good control on the regional attitudes of subsurface reflectors.

This study has provided evidence in support of the proposal of Indares (1997) for the existence of a High Pressure belt in this part of the Grenville Province, located between the Parautochthonous and Allochthonous belts, as defined by Rivers et al. (1989). The presence of Pinwarian rocks and chemically similar mafic dykes in both Tshenukutish terrane (High Pressure belt) and Hart Jaune terrane (Allochthonous Belt) implies lithological linkages between the two belts, suggesting that they may not have been widely separated before the Grenvillian Orogeny. There is, however, a profound contrast in their metamorphic evolution, in terms of both the *P–T* regime and the timing of metamorphism. In this part of the Grenville Province, the high-pressure rocks underwent their tectonic evolution in the lower part of a crustal-scale orogenic wedge during the Ottawan pulse of the Grenvillian Orogeny. In contrast, the main metamorphism in the respectively underlying and overlying Parautochthonous and Allochthonous belts took place in two mid-crustal thrust wedges during the Rigolet pulse.

The Grenvillian Orogeny in eastern Quebec appears to have produced two distinct episodes of crustal thickening, separated by an inversion during which the major unroofing of the high-pressure terranes took place. Although both episodes of crustal thickening were followed by normal faulting, in neither case does normal faulting appear to have been the major agent by which the crustal thickening was reversed. In the first case, the lower crust appears to have been extruded, in a presumably compressional environment. In the second case, normal faulting appears to have effected only small (<6 km) throws at a comparatively early stage of adjustment, when the rocks now exposed at the surface were still relatively deeply buried.

The high temperatures of metamorphism evident in both the high-pressure metamorphosed rocks of the MIZ and the overlying less deeply buried rocks are consistent with detachment of the lithospheric mantle beneath the zone of maximum thickening after ca. 1050 Ma. The resultant heating of the region of thickened crust may have been important in effecting the large-scale structural adjustment necessary to transport the deeply buried rocks to high structural levels. This structural adjustment appears to have resulted in an essentially flat Moho at a depth of about 50 km beneath the zone of crustal thickening, after the extrusion of the high-pressure rocks. Following the second thickening event, the rocks cooled slowly and reached their closure temperatures for Ar in hornblende by 968–905 Ma (Dallmeyer and Rivers 1983), implying that cooling after the Rigolet pulse was controlled primarily by erosion resulting from isostatic rebound. Near the Grenville Front, however, crustal thicknesses remain near 50 km, perhaps reflecting continued flexural strength of the lithosphere outside the region of thermal weakening.

Finally, we note that we have imaged Mesoproterozoic crustal reflectors beneath the Triassic Manicouagan impact crater. This implies that the reflectors retain seismic coherence and are not significantly offset or affected by the impact. Coupled with the observation that basement exposures on the outer shore of Manicouagan Reservoir show remarkably little evidence of the impact, we infer that the current erosion surface must lie near or beneath the base of the impact cone.

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