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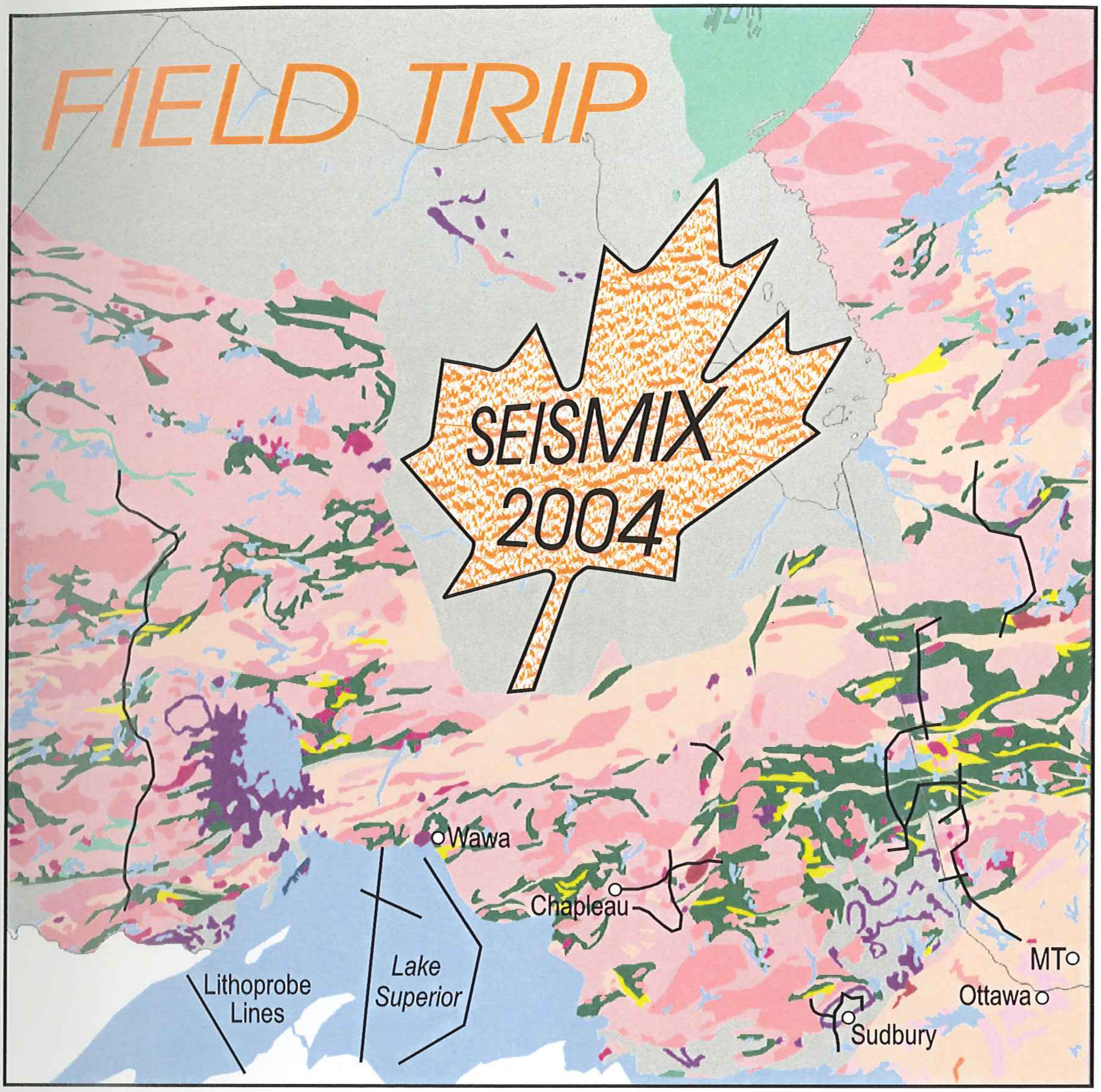

11th International Symposium on Deep Structure of the Continents and their Margins

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SEISMIX 2004

SEISMIX 2004

The 11th International Symposium on

Deep Seismic Profiling of the Continents and their Margins

Post-conference excursion

Field Guide

2 – 5 October 2004
Chapleau – Timmins – Sudbury, Ontario

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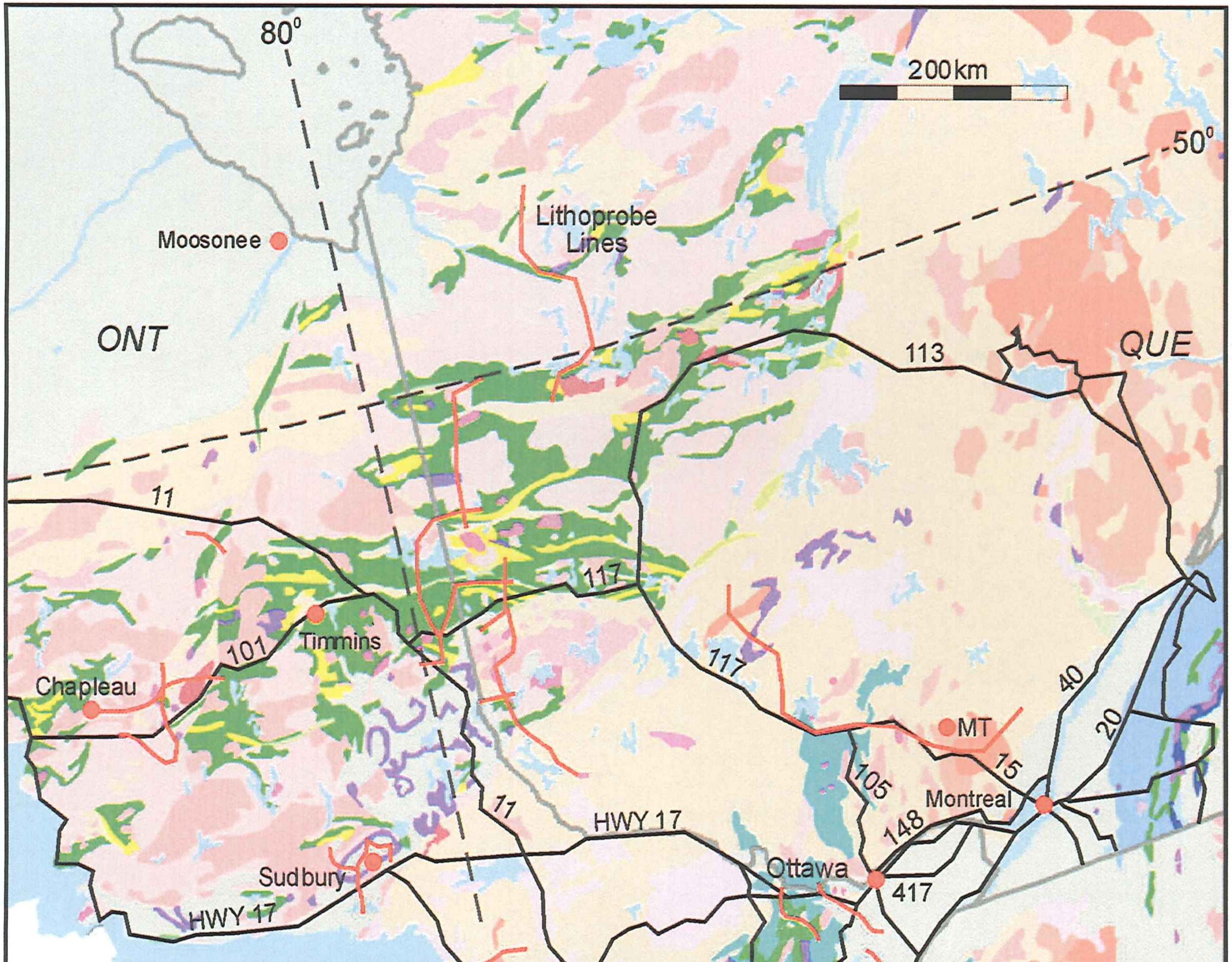

Day 1:
Mont Tremblant
to Chapleau
Abitibi

Day 2:
Chapleau to
Timmins
Kapuskasing Transect

Day 3:
Timmins to
Sudbury
Sudbury Basin

Day 4:
Sudbury to
Montreal
Grenville Front

Part 1.

Field Guide to the Kapuskasing Uplift, Chapleau-Foleyet Transect: a window on the deep crust

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Introduction

The Kapuskasing uplift, a major geophysical and geological anomaly of the central Canadian Shield (Fig. 1), was chosen as one of the first LITHOPROBE transects when it was recognized to be one of a few exposed cross-sections of the continental crust (*cf.* Fountain and Salisbury, 1981; Percival *et al.*, 1992) and thus presented an opportunity to explore the composition, structure and evolution of the continent at depth. LITHOPROBE studies (1984-1994) included geophysical (seismic reflection and refraction, electromagnetic, gravity, paleomagnetic, heat flow, modeling), geological (mapping, structural analysis, thermobarometry, petrology, fluid inclusions), geochronological (U-Pb zircon, titanite, monazite, ^{40}Ar-^{39}Ar, Rb-Sr) and geochemical (major, minor, trace elements; Nd-Sm, Pb, O isotopes) work that culminated in a Special Issue of the Canadian Journal of Earth Sciences (Percival, 1994; Percival and West, 1994). During this trip we will be visiting exposures relevant to several themes of current interest:

(1) Nature of lower crustal reflectors

Considerable ambiguity exists in the interpretation of seismic reflection profiles owing to uncertainty about the cause of deep reflectivity. Each profile requires interpretation in light of all available constraints, but a limited number of possibilities have been suggested for the subhorizontal, discontinuous reflectivity observed at the base of many continental sections. They include ductile thrust faults analogous to shallow thin-skinned structures, extensional detachments, mafic sills, igneous layering, intrusive complexes, and overpressured fluid zones. During the trip we will examine the transition between relatively homogeneous granitic and tonalitic rocks at mid-crustal levels, to a reflective package of layered mafic granulite and tonalite in the deep crustal part of the section *(Stops 1-1, 1-2, 1-5, 1-6, 1-9).*

(2) Upper Crust – Lower Crust transition: steep to flat structures

An enigmatic feature of many regions is the predominance of steep structures in surface exposures and underlying subhorizontally reflective crust. Possibilities for the transition include listric structures, allochthonous superstructure, unseen sills or intrusions, and structural overprinting. In the Chapleau area we will demonstrate that the onset of reflectivity coincides with the development of subhorizontal, extensional high-strain zones that overprint an earlier generation of steep structures *(Stops 1-1, 1-2).*

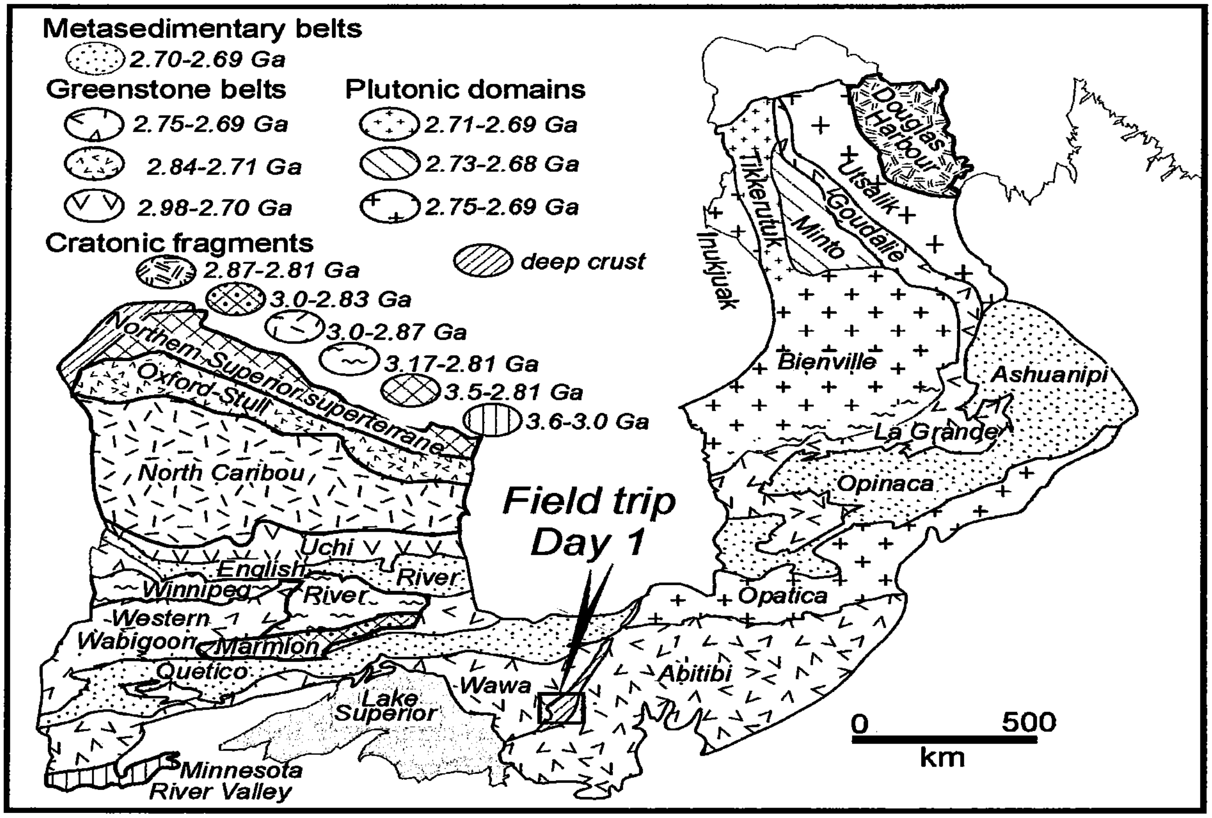


Figure 1. Tectonic map of the Superior Province, central Canadian Shield, showing distribution of cratonic fragments and greenstone belts.

(3) Vertical evolution of the crust

A key discovery related to (2) is that the deep crust remained hot, ductile and structurally mobile long after the upper crust had frozen into its present configuration. In this region the major orogenic event recorded at high structural levels (Abitibi, Wawa greenstone belts) peaked at about 2690 Ma. In the Chapleau area the ductile extensional strain has been dated at about 2650 Ma *(Stops 1-1, 1-2)* and in the deepest part of the section, the peak of granulite-facies metamorphism occurred between about 2640 and 2620 Ma *(Stops 1-7, 1-9)*, as much as 70 m.y. after stabilization of the upper crust. The cause of this phenomenon, which is also apparent from xenolith studies in other cratons, remains poorly understood.

(4) Moho topography

One of the geophysical anomalies that characterizes the Kapuskasing structure is a ca. 8 km bulge in the Moho, to 53 km from background levels of 40-45 km. The Moho topography likely formed in response to regional shortening at ca. 1900 Ma, which brought the deep crustal levels of the Kapuskasing uplift to the surface along a northwest-dipping thrust fault *(Stop 1-8)*. This brittle translation and uplift was balanced by ductile flow of material into a crustal root. The root exhibits refraction velocities in the 7.5 km/s range, much higher than those of surrounding lowermost crust (6.8-7.0 km/s). Coupled with the observation of a positive gravity anomaly over the zone of thick crust, these constraints suggest that eclogite was produced when the root formed (Fig. 2).

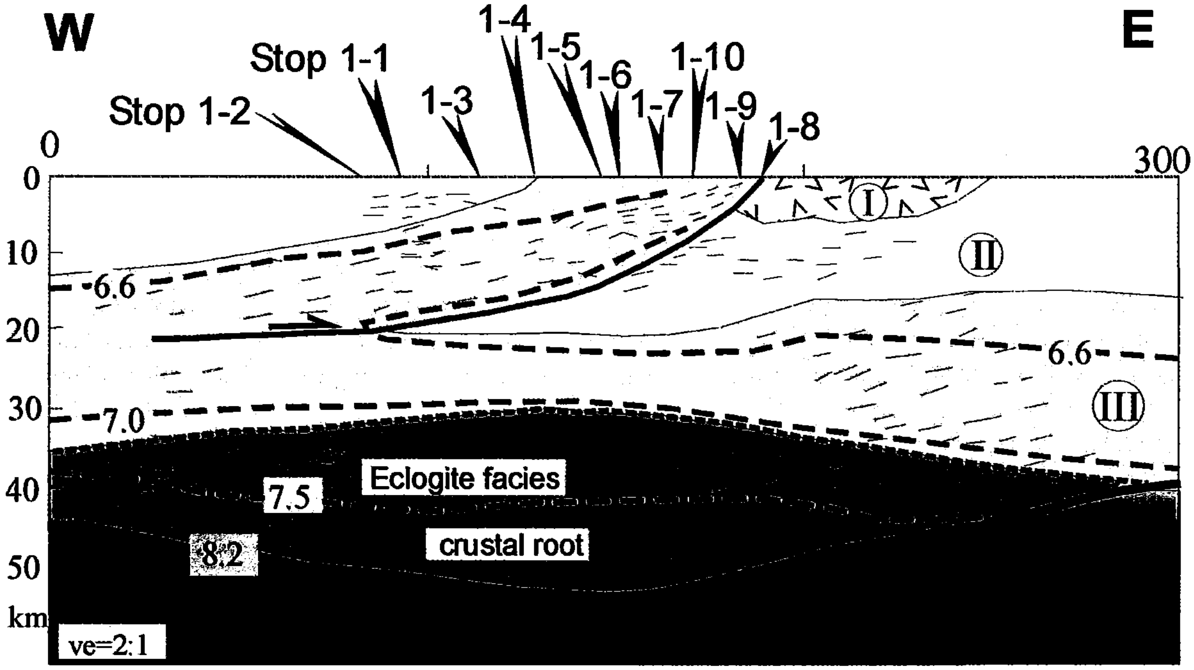


Figure 2. Generalized west-east cross section illustrating the layered nature of southern Superior Province crust, and the structure of the Kapuskasing uplift. I-Greenstones comprise a layer up to 10 km thick of greenschist-facies, dominantly mafic rocks with upright structures. Structurally below is: II- Tonalitic gneisses up to 10 km thick, metamorphosed to amphibolite-facies and characterized by increasingly pervasive low-dipping extensional shear zones at depth. This layer grades with increasing depth to: III- Layered gneiss (0.1 - 10 km scale), metamorphosed to amphibolite and granulite facies, with low dips and high Vp (> 6.8 km/s).

Geological overview

The Superior Province (Fig. 1), Earth's largest exposed Archean craton, records a dramatic series of continent-building events about 2.7 billion years ago (2.7 Ga). It contains rocks ranging in age from ~3.8 to 2.6 Ga that were stabilized by about 2.5 Ga. Proterozoic tectonism at the margins and intrusion of mafic dyke swarms have only slightly modified its cratonic nature. Debate continues as to the nature of tectonic processes during the Archean (*cf.* Hamilton 1998; 2003; de Wit 1998), although accretion of diverse oceanic and microcontinental terranes in a modern plate-tectonic framework appears to best suit observations from most of the Superior Province (e.g. Card 1990; Williams *et al.* 1992; Calvert *et al.*, 1995; Stott, 1997; Percival *et al.* 2001; White *et al.*, 2003). The rich mineral wealth of the Superior Province has inspired intense geological investigation over the past twenty years, including three Lithoprobe transects (Percival and West 1994; Ludden and Hynes 2000; White *et al.* 2003), the western Superior NATMAP project (e.g. Percival *et al.* 2000), Operation Treasure Hunt, major mapping projects in northern Quebec (e.g. Leclair *et al.* 1998), studies of the geodynamic setting of greenstone belts (e.g. Kerrich *et al.* 1999), and many private exploration initiatives aimed at base, precious metal and diamond targets (e.g. Discover Abitibi).

Significant advances in understanding the geology of the southern Superior Province have been facilitated through the integration of bedrock mapping with Lithoprobe seismic lines and high-resolution U-Pb geochronology[1]. The present field excursion across the Kapuskasing uplift will illustrate some of these advances, particularly in relating seismic features to geological structures of known age and origin.

Well preserved supracrustal sequences of the Superior Province were studied as early as the late 1800's and have been the source of many fundamental ideas on Archean geology (e.g. Lawson 1913). A first-order feature of the Superior Province is its linear subprovinces of distinctive lithological and structural character, accentuated by subparallel boundary faults (e.g. Card and Ciesielski 1986). Trends are generally east-west in the south, WNW in the northwest, and NNW in the northeastern Superior (Fig. 1). Recent work based on isotopic and zircon inheritance studies has provided a means of "seeing through" the latest structural and magmatic events, particularly in regions dominated by granitic rocks, revealing fundamental age domains across the Superior Province (Fig. 1).

At least five microcontinental fragments of Mesoarchean age are recognized through pervasive Neoarchean magmatism, metamorphism and deformation. Their shape mimics the province-scale structural pattern, swinging to north-south trends in the northeast. The oldest continental crust (up to 3.8 Ga) occurs in the Northern Superior superterrane (NSS; Skulski *et al.* 2000). To the south, the North Caribou- La Grande – Goudalie superterrane (NCLGG; Thurston *et al.* 1991; Stott 1997) represents a large region of ca. 3.0 Ga crust that may have acted as a continental nucleus during assembly of the Superior Province (Goodwin 1968; Williams *et al.* 1992). Further south, the Winnipeg River (WR) and Marmion (MM) terranes are relatively small continental fragments dating back to 3.4 and 3.0 Ga respectively (Beakhouse 1991). In the far south, the Minnesota River Valley terrane (MRVT) of unknown extent contains remnants up to ca. 3.6 Ga (Goldich *et al.* 1984; Gibbs *et al.*, 1984). Domains of oceanic ancestry, identified by juvenile isotopic signatures and lack of inherited zircon, separate some of the continental fragments. These dominantly greenstone-granite terranes generally have long strike lengths and record geodynamic environments including oceanic floor, plateaux, island arc and back-arc settings (e.g. Thurston 1994). Examples include the Oxford-Stull terrane in the north, the western Wabigoon in the west, and the Abitibi-Wawa subprovince in the southeast.

Still younger features, the metasedimentary belts (Breaks, 1991; Williams, 1991), separate some of the continental and oceanic domains. Extending across the entire province, these 50-100 km wide belts of metagreywacke, migmatite and derived granite appear to represent thick syn-orogenic sequences, deposited, deformed and metamorphosed during collisional orogeny.

The Abitibi subprovince represents the upper crustal part of the transect. Large parts have well preserved primary features, having been metamorphosed only to sub-greenschist facies (Jolly, 1978; Powell *et al.*, 1993). Many theories have been advanced on the tectonic setting of Abitibi volcanism (2747-2698 Ma), although during the past decade models have converged on oceanic plateau and arc regimes (e.g. Wyman *et al.*, 2002; Kerrich *et al.*, 1999; Ayer *et al.*, 2002). Volcanic and associated plutonic rocks were deformed prior to deposition of the unconformable Timiskaming Group (ca. 2675 Ma), which is preserved in broad deformation zones such as the Porcupine-Destor fault. This steep structure, famous for its associated lode gold deposits, is not imaged directly on seismic reflection profiles, but can be traced through offset of adjacent reflectors, to depths of about 15 km (Green *et al.*, 1990; Discover Abitibi profiles).

Until 1980 the Kapuskasing zone was enigmatic to the extent that proposed interpretations ranged from an Archean continental suture zone (Wilson, 1968), through a sinistral (Watson, 1980) and dextral (Goodings and Brookfield, 1992) transcurrent fault zone of Paleoproterozoic age, to a Mesoproterozoic tensional horst (Burke and Dewey, 1973; Thurston *et al.*, 1977).

[1] This dating technique provides ages for deposition, intrusion and metamorphism of rocks with precision commonly on the order of 0.1% (±2-3 million years on rocks 2700-3000 million years old). These dates allow calibration of Precambrian time with resolution similar to that provided by fossils in Phanerozoic strata.

Subsequent interpretations have converged on a southeast-verging thrust model, exposing an oblique crustal cross section (Fig. 2). Correlation of subprovinces east and west of the Kapuskasing uplift is important in distinguishing between models of the Kapuskasing zone as an Archean suture or an intracratonic structure. Percival and Card (1983; 1985) argued for correlation of the Abitibi and Wawa subprovinces on the basis of similarity in age of felsic volcanic (2745-2698 Ma) and plutonic (2740-2680 Ma) rocks.

The Kapuskasing uplift (Figs. 2, 3) exposes an oblique depth section through the Abitibi-Wawa and Quetico subprovinces. Three megalayers observed at surface across the profile have been correlated geophysically with features of the crust beneath the Abitibi-Wawa belt (Fig. 2):

(1) Greenstones (up to 10 km of supracrustal rocks deposited mainly 2.75-2.70 Ga; deformed and metamorphosed to greenschist-facies (up to 450°C at 2-3 kb) by 2.66 Ga), characterized by mafic compositions, upright structural features, weak reflectivity and moderate Vp (6.3-6.6 km/s). This layer is intruded from beneath by:

(2) Tonalitic Gneiss (up to 10 km of gneiss and granite crystallized mainly 2.71-2.66 Ga; deformed and metamorphosed to amphibolite-facies (600-700°C at 4-6 kb) by 2.645 Ga), with lower Vp (to 6.5 km/s) and increasing reflectivity with depth. This grades with increasing depth to:

(3) Layered Gneiss (interlayered tonalitic, mafic and paragneisses and anorthosite on 0.1-10 km scale, crystallized mainly 2.76-2.62 Ga and metamorphosed to amphibolite and granulite facies (700-850°C at 8-12 kb) at 2.70-2.60 Ga), with low dips, bright reflectivity and higher Vp (*ca.* 6.9 km/s). Material of this character could occupy the lower crust of the Abitibi belt, whose velocity near the Moho is *ca.* 7 km/s and xenoliths of which are garnet granulite.

This crustal structure was in place by *ca.* 2.6 Ga, followed by cooling and modest erosion to establish the Superior Province as a craton by ca. 2.5 Ga. Rifting at its southern margin to form the Huronian basin began shortly afterward (2.49-2.45 Ga), accompanied by emplacement into the craton of the radiating Matachewan dyke swarm (Halls *et al.*, 1994). Subsequent dyking events at 2.22, 2.17, 2.14 and 2.04 Ga mark plume impingement related to distal rifts (Ernst and Buchan, 2001a,b). Several lines of evidence indicate that all of these dyke swarms were affected by the main Kapuskasing uplift event: geometric relations among dyke swarms and faults; paleomagnetic and geobarometric observations of dykes crystallized at depth in the Kapuskasing zone; and analysis of the distortion of the Matachewan swarm in the hanging wall of the Kapuskasing uplift which indicates dextral transpression (Ernst and Halls, 1984). The Kapuskasing uplift event has not been dated directly, but present constraints suggest a final uplift age of *ca.* 1.9 Ga.

Geophysical constraints on lithosphere structure

Crustal refraction experiments (e.g. Halls, 1984) were the first to show an anomalous seismic velocity structure associate with the Kapuskasing structure (at that time known mainly as a positive gravity anomaly). In 1984 a regional refraction survey was conducted of a roughly 300 km square region centred on the Chapleau area (Boland and Ellis, 1989; Wu and Mereu, 1990). Both studies substantiated the presence of thick crust beneath the Kapuskasing uplift, as well as noting anomalously high velocities in the upper crust of the uplift. In an east-west profile, the depth at which high velocities (>6.5 km/s) are reached, rises from ~20 km beneath the Wawa area, to ~ 5 km beneath the Kapuskasing structure.

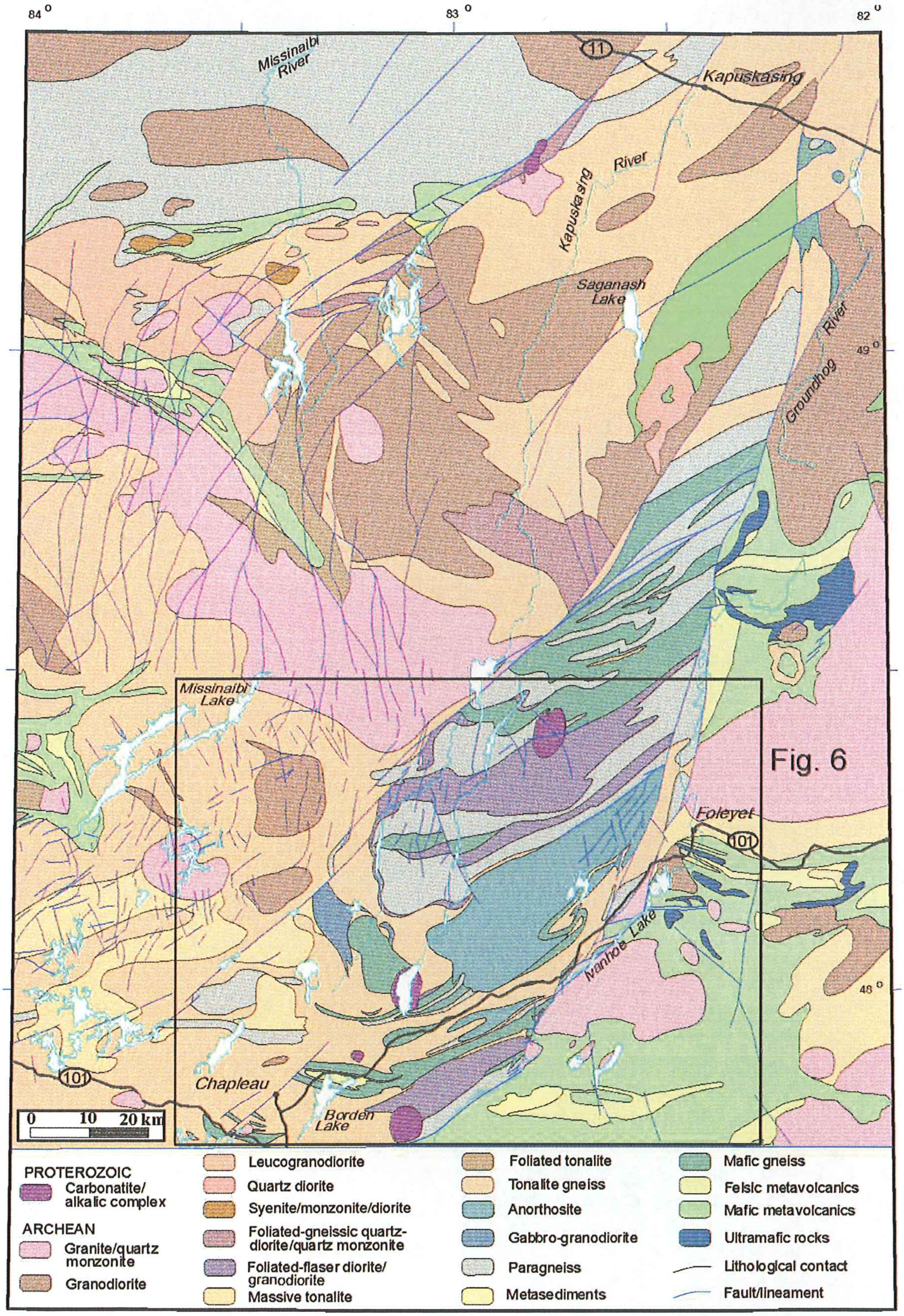


Figure 3: Geological map of the Kapuskasing uplift (after Percival and West, 1994).

Boland and Ellis (1989; 1991) concluded that the mantle velocity is 8.1 km/s at the Moho, and that the crust is thicker by ~8 km beneath the southern Kapuskasing structure than background levels of 40-45 km (Fig. 4), although resolution is less than optimal (White and Boland, 1992). The structure is internally consistent between profiles and can be reconciled with the Bouguer gravity field (Boland and Ellis, 1991). Wu and Mereu (1990; 1992) inferred a simpler structure for the upper crust and projected the high velocity zone and region of thick crust further northeast.

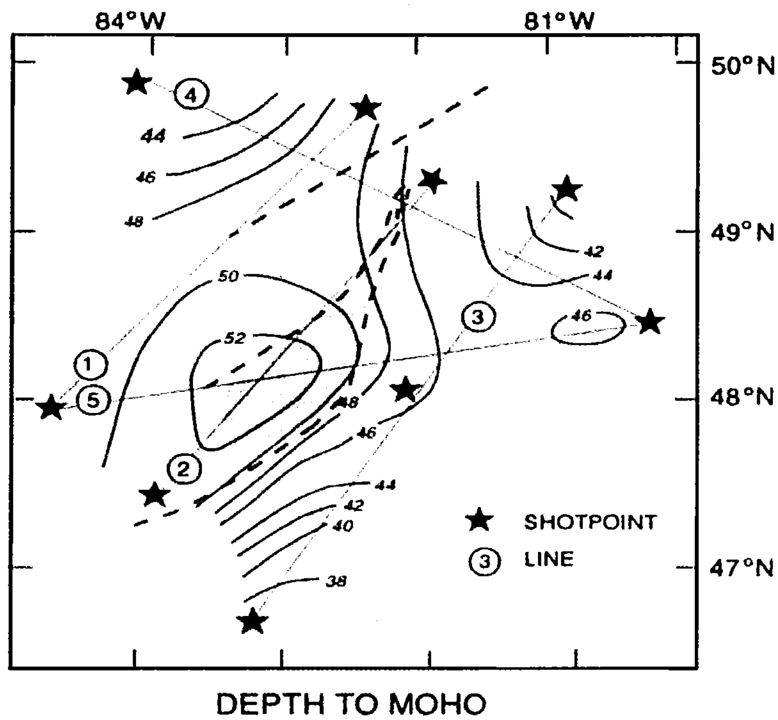


Figure 4. Compilation of seismic refraction results, showing shotpoints, receiver lines, and contours of depth to the Moho (after Boland and Ellis, 1989).

A variety of electromagnetic studies have been carried out in the Kapuskasing region. A reconnaissance investigation showed that there is no significant regional-scale anomaly associated with the Kapuskasing zone, although a modest increase occurs at mid-crustal depths, as in many other parts of the Shield (Bailey *et al.*, 1989; Kurtz *et al.*, 1989; Jones *et al.*, 1994). Based on observations of graphite grain-boundary films in Kapuskasing rocks, Mareschal *et al.* (1992) related the conductivity structure to the presence of graphite, the conductive effects of which were presumably destroyed when the films cracked during exhumation.

Lithoprobe acquired 358 line km of reflection data in 1987-88, largely in the southern part of the Kapuskasing zone (Fig. 5). All of the data exhibit strong reflectivity in the upper crust. At greater depth, the strength and continuity of reflectors generally fades. No events associated with the Moho are recognized. A striking characteristic of the data is the variability of reflectivity on N-S *vs* E-W profiles (e.g. continuous reflectivity on E-W line 2-3-4 *vs* short discontinuous events on north-south line 1). The asymmetric pattern suggests that most reflections arise from anticlinal tops of moderately corrugated surfaces striking roughly east-west.

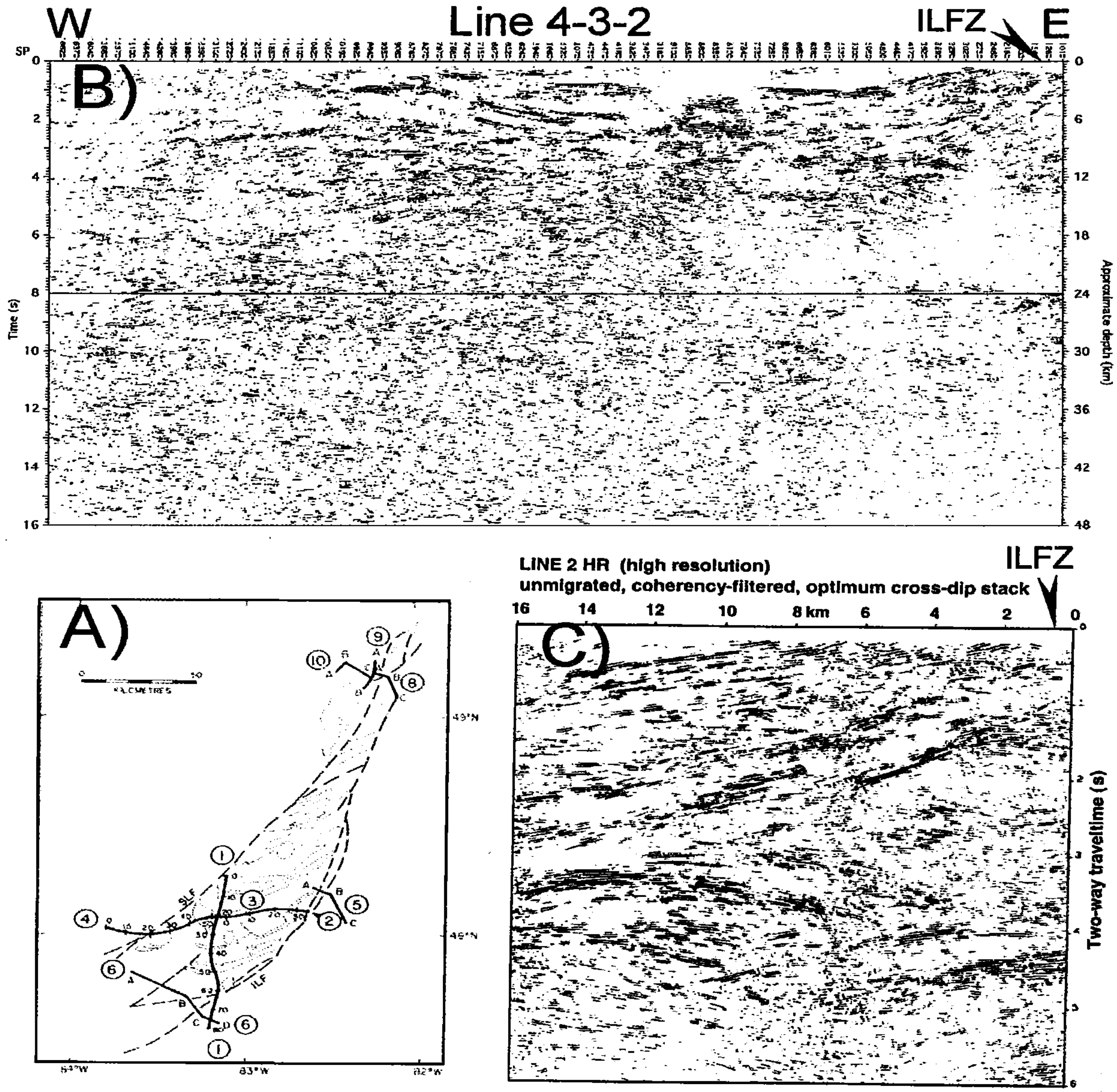
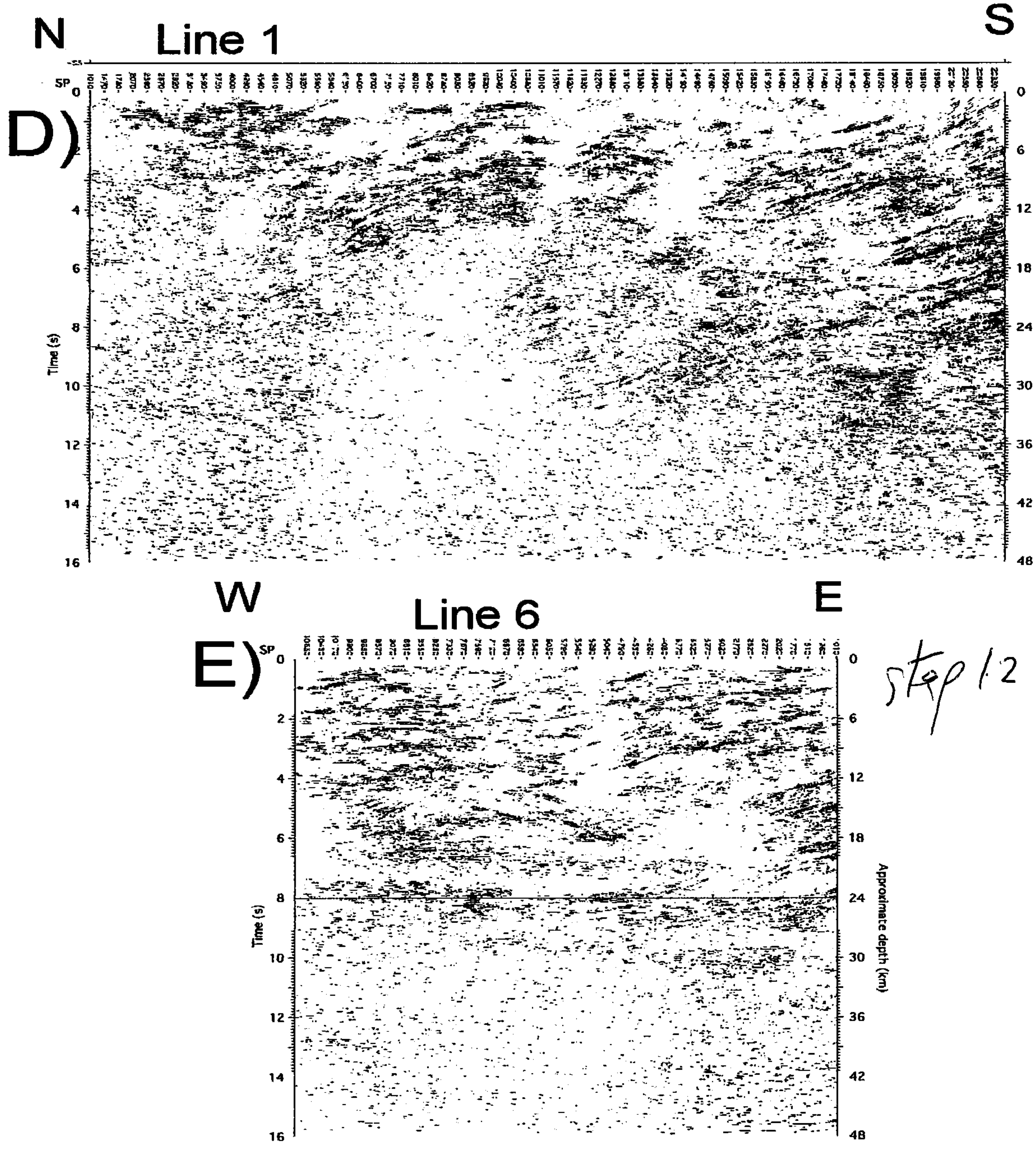


Figure 5. Seismic reflection profiles of the southern Kapuskasing uplift: a) map showing location of profiles; b) Line 4-3-2 (east-west); c) Line 2 high resolution; d) Line 1 (north-south); e) Line 6 (northwest-southeast).

A special effort was made to image the Ivanhoe lake fault zone, through high-resolution profiling (Line 2HR)(Fig. 5c). The general pattern of shallowly dipping reflectivity seen on the regional line steepens markedly (west apparent dip) near the fault zone, and a weak but continuous event dipping about 45°W may correspond to the easternmost fault surface itself (Geis *et al.*, 1990). Alternatively, it is possible that the Ivanhoe Lake fault zone is itself a steeply dipping feature that has no direct seismic expression, and the west-dipping reflections represent folds of earlier surfaces in and against the fault zone.

Line 2HR contains a prominent, shallowly west-dipping laminar band of reflections at very shallow depth. Traceable to within 1 km of the surface (Percival *et al.*, 1989; 1991; Milkereit *et al.*, 1991), the reflectors have a true dip of 17°W (Kim *et al.*, 1992) and project to surface outcrops of interlayered tonalitic and mafic gneiss that have appropriate scale and velocity contrast to produce the observed reflection response (White *et al.*, 1992)(*See Stop 1-9*).



step 12

Figure 5 (continued)

Geological and geophysical elements of the crustal structure are portrayed in an east-west cross section in figure 2. Salient points include: 1) a ca. 8 km thick bulge in the Moho; 2) anomalous (>7.0 km/s) refraction velocities within the crustal root; 3) disrupted lower crustal reflectivity within the root; and 4) a positive Bouguer gravity anomaly associated with the Kapuskasing uplift. The combined reflection and refraction data, along with surface geobarometric constraints, support a deformation model involving 27 km of horizontal (NW-SE) shortening, accommodated by brittle thrusting in the upper 20 km of the crust and by ductile flow at depth into the crustal root. Root formation, inferred to have occurred at about 600°C (ca. 10^{21}Pa's "dry" granulite rheology), destroyed typical ambient reflectivity observed on adjacent profiles and left a 70-km-wide, 15 km-thick zone with high Vp (7.0-7.8 km/s). The presence of this dense material in the root accounts for the absence of a long-wavelength negative gravity anomaly which should normally accompany thick crust, and may contribute to the 30-60 mGal positive anomaly which defines the Kapuskasing "high" over 500 km of strike length, independent of the presence of high-grade rocks at the surface (*cf.* Percival and McGrath, 1986). Although imbrication of lower crust and lithospheric mantle could yield velocities in this range, there is no evidence on reflection profiles of such geometry. Alternatively, eclogite is stable at lower crustal conditions (600°C, 15-20 kb) and mixed mafic and felsic lithologies in the eclogite facies would have appropriate bulk velocities. In general, the conversion of garnet granulite to eclogite is kinetically prohibited in the absence of fluids or strain. In the Kapuskasing root zone, the high-velocity material could be eclogitic assemblages which formed through ductile strain during the crustal thickening event.

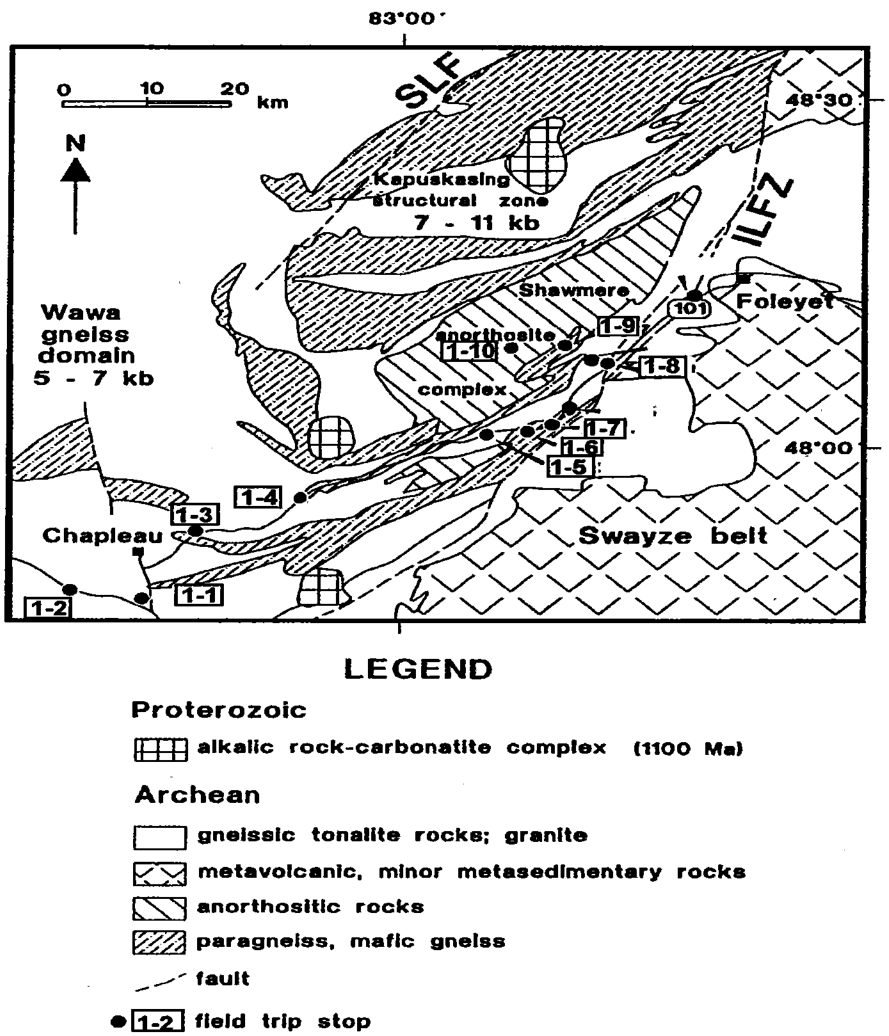


Figure 6. Location of field trip stops in the southern Kapuskasing uplift
SLF: Saganash Lake fault; ILFZ: Ivanhoe Lake fault zone.

Field Trip Stops (Fig. 6):

The first two stops will be in a large "sea" of tonalitic gneisses that crystallized and were deformed at mid-crustal (15-20 km- 5-7 kbar) levels. Here we will examine the origin of seismically reflective subhorizontal structures. The remaining stops along highway 101W traverse the Kapuskasing uplift from west to east, providing views of successively deeper structural levels (to 30-35 km- 10-11 kbar).

Stop 1-3 illustrates the nature and age of sedimentary rocks buried to mid-crustal depths and provokes questions as to their pathway from the surface. At stop 1-4 we will examine the amphibolite facies to high-pressure granulite facies transition in mafic and tonalitic rocks, taken to coincide with a mid-crustal (Conrad) refraction velocity discontinuity. Stop 1-5 demonstrates the nature of deep-crustal, ductile, high-strain zones. At stop 1-6, typical high-pressure granulite-facies gneisses illustrate processes of melt extraction from mafic rocks, as well as their complex structural history. Stop 1-7 is the outcrop made famous in Krogh's 1993 study of the timing of deep-crustal metamorphism. At the Ivanhoe Lake fault zone (Stop 1-8) we will see evidence for ca. 1.9 Ga brittle deformation related to about 17 km of thrust uplift of the Kapuskasing zone rocks. We will then back-track on the Warren-Carty forest access road to examine exposures at stop 1-9 coinciding with a zone of bright reflectivity traced to within 300 m of the surface. These outcrops were stripped and mapped carefully in the early 1990s as part of a planned, but not implemented, continental drilling initiative. Finally, at stop 1-10 we will visit the Shawmere anorthosite complex: a large, partly layered intrusion with preserved primary features.

Stop 1-1:

Tonalite gneiss with post-2660 Ma extensional overprint

(321987; 5293386) (NAD83)

The vertical face of this roadcut presents one of the clearest examples of cross-cutting relationships between early, upright foliations and spaced, gently dipping high-strain zones (Fig. 7). The outcrop consists of foliated to gneissic tonalite, foliated xenoliths of amphibolite, as well as cross-cutting tonalitic veins and granitic pegmatite dykes. All of these units are deformed within high-strain zones at the top and bottom of the outcrop. Mafic xenoliths in the central, steeply dipping part of the outcrop contain weak foliations and are generally irregular in shape, whereas in the high-strain zones the xenoliths are flattened and elongated parallel to a strong east-west trending lineation (visible at the lower southern end of the outcrop). The tonalite gneiss has a crystallization age of 2675±2 Ma whereas the tonalite veinlets crystallized at 2661±1 Ma (Moser *et al.* 1996). Thus the gently dipping shear zones in this outcrop formed some time after 2661±1 Ma.

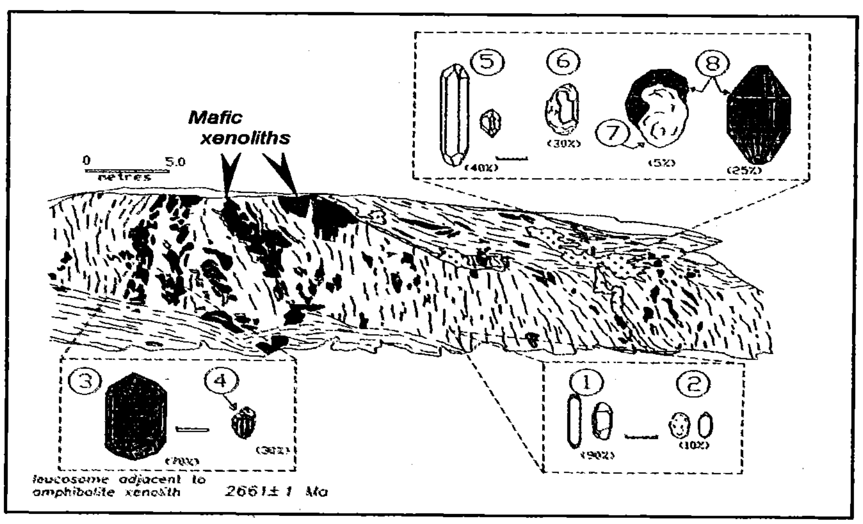


Figure 7. Sketch of road-cut at Stop 1-1 as viewed looking westwards. Sketches of different zircon types at different locations are shown. Note that the early, steep foliation in the tonalite gneiss, the xenoliths and the tonalitic leucosomes are deformed within the gently dipping shear zones at the top and bottom of the outcrop (after Moser, 1994).

Stop 1-2:

Tonalite gneiss with extensional lozenges

(313050; 5294592)

This outcrop area illustrates variable extents of extensional overprinting related to bulk rock composition. The tonalite gneiss at the western end of the outcrop area has a strong subhorizontal fabric and associated east-west lineation. The gneissosity here is folded by late, open, north-trending folds (Moser, 1994).

To the east, large roadcuts through amphibolitic gneisses expose a complex network of anastomosing high-strain zones. Between these extensional deformation zones is preserved an earlier steeply-dipping fabric. The age of the anastomosing fabrics in the amphibolite is constrained by the U-Pb zircon age (2660±2 Ma) of a west-dipping sill of tonalite gneiss exposed in the southern roadcut. The cross-cutting sill carries the extension-related fabrics and therefore provides a maximum age for the deformation.

Interpretation:

These outcrops likely represent the transition zone between non-reflective shallow crustal levels, characterized by homogeneous rock types and steeply dipping structures, and subhorizontally reflective, deep crustal levels with pervasive flat structures. Seismic images of this area (Lithoprobe line 6) exhibit subhorizontal reflectivity at shallow levels (Fig. 8). Key information acquired from these rocks includes:

(1) the subhorizontal structures are an overprint, formed late (post-2660 Ma; Moser *et al.*, 1996) in the regional tectonic history, rather than being related to accretionary processes (i.e. thrust faults); and

(2) the structures can be attributed to the effects of regional extension acting on the ductile deep crust. This process may be related to tectonic escape (orogen-parallel flow) or crustal collapse (*cf.* Sandiford, 1989).

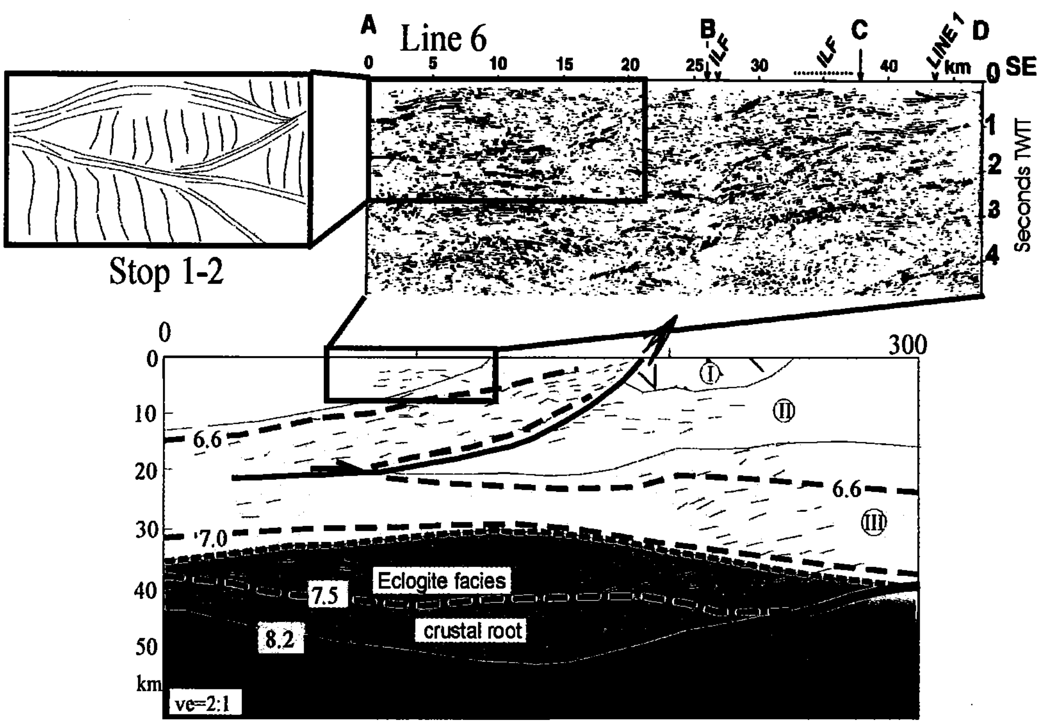


Figure 8. Interpretative section showing the correlation between the low-dipping shear zones and seismic reflectivity observed on Line 6.

Set odometer to zero at Hwy 129-101 junction

12.0 km

Stop 1-3:

Borden Lake conglomerate

(330337; 5304922)

The Borden Lake conglomerate is part of a belt of supracrustal rocks (pillow basalt, rhyolite, sandstone, conglomerate) exposed in the 3 km-wide Borden Lake synform. At this location the 30 m-thick unit consists of cobbles of quartz porphyry, granodiorite, plagioclase-porphyritic andesite and amphibolite. With the exception of the more competent granitoid cobbles the clasts are elongated parallel to minor fold axes and flattened parallel to the gently north-dipping foliation. Matrix garnet and hornblende indicate amphibolite-facies metamorphic conditions.

U-Pb dates of individual cobbles provide a range of ages between 2690±2 and 2667±2 Ma (Krogh, 1993). The youngest clast places a maximum age on deposition, making this supracrustal unit the youngest known in the Superior Province, post-dating Timiskaming sequences by about 10 m.y. Based on these data, the conglomerate must have been transported to mid-crustal depths after stabilization of overlying greenstone-belt crust. The age of strain in this area is constrained by zircon (2603±2 Ma) in pegmatite filling a strain shadow (Krogh, 1993).

Interpretations:

Percival and West (1994) hypothesized that the Borden Lake belt is akin to Timiskaming sequences, and was deposited and buried within a transcurrent fault zone. Moser *et al.* (1996) proposed that the Borden Lake belt rocks were underthrust during an episode of structural underplating (2669-2660 Ma).

25.3 km

Stop 1-4:

Mafic gneiss with 2660 Ma metamorphic zircons

(340718; 5310290)

Major lithological units within the high-grade metamorphic Kapuskasing zone include mafic gneiss and paragneiss, thought to represent a mafic volcanic – greywacke supracrustal package, and intrusive tonalite, diorite and anorthosite. Regionally, mafic gneiss and paragneiss yield Nd model ages in the range 2.75-2.70 Ga (Shaw *et al.*, 1994), similar to those of supracrustal rocks of the Abitibi belt. Depositional and intrusive ages are poorly known for supracrustal and intrusive units, although metamorphic zircons are abundant. Outcrops in this area provide insight into processes at the amphibolite-granulite-facies boundary.

This outcrop illustrates relationships between mafic gneiss enclaves and tonalitic host rocks observed throughout the Kapuskasing zone. Within enclaves, mafic gneiss consists of hornblende – plagioclase ± garnet ± clinopyroxene yielding P-T estimates of ~7 kbar, 700°C. Zircons within cores of enclaves, inferred to be of metamorphic origin, are 2660 Ma (Krogh, 1993), and thought to have formed as a byproduct of the reaction (Percival, 1983):

Hornblende + plagioclase = garnet + clinopyroxene + tonalitic melt

The margins of the enclaves consist of hornblende-plagioclase assemblages, inferred to have developed by ingress of water from crystallizing magmas:

Garnet + clinopyroxene + water = hornblende + plagioclase

Zircon from these reaction zones is high in uranium and yields ages of 2642 Ma, similar to ages obtained from dykes (2640 Ma; Krogh, 1993).

Outcrops with similar relationships between metamorphic assemblages and plutonic phases occur to the east, at deeper structural level. Krogh (1993) observed similar zircon morphologies and relative ages, but with younger absolute ages (2640, 2630 Ma) (*See Stop 1-7*).

Interpretation:

Several changes occur within the transition zone from mid- to deep crustal levels:

(1) structural attitudes change from steep to flat through an extensional overprinting event;

(2) metamorphic grade changes from amphibolite to granulite facies, with complexity introduced by production and migration of partial melts and water;

(3) the lithologic assemblage changes from tonalitic with mafic enclaves at mid-crustal levels to the west, to a higher proportion and larger bodies of mafic material, and a large anorthosite complex, at deep crustal levels.

Each of these factors contributes to predicted higher seismic reflectivity at deeper levels, 1) by providing favourable (subhorizontal) orientations of layering; 2) by increasing the density contrast between mafic (garnet-clinopyroxene *vs* hornblende-plagioclase) and felsic rock types; and 3) by introducing >100m-scale, semi-continuous compositional layers. There is probably also a bulk density increase through this zone based on the lithological change, raising the possibility of an exposed mid-crustal (Conrad) seismic velocity discontinuity (e.g. Percival, 1986).

36.1 km

Stop 1-5:

Tonalitic straight gneiss with ~1144 Ma diatreme breccia

(350486; 5313936)

This outcrop represents a high-strain zone derived from sparsely xenolithic garnet-hornblende-biotite tonalite. Highly attenuated, moderately dipping dark layers, traceable for several metres, may be stretched mafic xenoliths. Small rootless folds and a weak subhorizontal lineation are evident locally. Garnet up to a few mm across appears to have overgrown the foliation in the otherwise fine- to medium-grained gneiss. Kinematic indicators have not been observed here or elsewhere along the high-strain zone.

The western end of the outcrop contains a 20-cm-wide diatreme breccia dyke with lamprophyric matrix. Similar lamprophyres to the east have yielded Ar-Ar ages of 1144 Ma (Hanes *et al.*, 1994). Xenoliths include spinel lherzolite of presumed mantle origin and massive granite thought to derive from the Swayze belt located at a depth of approximately 10 km beneath the gently northwest-dipping Ivanhoe Lake fault zone.

Interpretation:

Leclair *et al.* (1993) suggested that major high-level structures such as the Kirkland Lake – Larder Lake break penetrate to the mid – to lower crust and may be exposed in the Kapuskasing uplift. High-strain zones such as this were inferred to be the expression of such ductile fault zones.

51.5 km

Stop 1-6:

Structural style of Kapuskasing mafic and felsic gneiss

(364229; 5319432)

This outcrop illustrates the moderately northwest dipping, gently rolling geometry of foliation and layering typical of the Kapuskasing zone. Mafic gneiss in this exposure exhibits modal layering on the 2-5 cm scale: alternating garnet-, clinopyroxene-rich and hornblende-rich compositions that may be an inherited primary compositional feature. It is possible that the compositional domains reflect variable degrees of seawater alteration of mafic volcanic rocks, although the mantle-like oxygen isotopic signatures of Kapuskasing mafic gneisses preclude much seawater interaction (Li *et al.*, 1991; Puris and Wickham, 1994). Metamorphic conditions in this area reached 10 kbar (30 km depth), 800°C (Mader *et al.*, 1994). Numerous examples illustrate leucosome coalescing from concordant, mm-scale seams into wider, discordant veins.

The mafic bodies are enclosed in hornblende-biotite tonalite gneiss, forming lozenge-like structures attributed by Bursnall *et al.* (1994) to D3 deformation. In this part of the Kapuskasing zone D1 forms gneissic layering in mafic rocks; D2 structures are intrafolial, reclined folds, and D3 represents the predominant outcrop-scale fabric (Bursnall *et al.*, 1994). The D3 structures resemble the structural style of the gneisses at stops 1-1 and 1-2 (Moser, 1994) except that the foliation in mafic gneisses is concordant to that in enclosing tonalite gneiss, suggesting that the higher deformation temperatures increased ductility of both mafic and tonalitic rocks.

54.9 km

Stop 1-7:

Kapuskasing mafic gneiss with 2640 Ma metamorphic zircon; 2630 Ma retrogression

(367690; 5319344)

At this outcrop mafic and ultramafic (garnet-hornblende-clinopyroxene-orthopyroxene) rocks are cut by tonalitic dykes with hydration selvedges. U-Pb dates on low-U zircons in mafic granulites indicate metamorphism at 2640 Ma, whereas high-U zircons in hydration selvedges have ages in the 2630 Ma range (Krogh, 1993). A cross-cutting pegmatite yielded an age of 2584 Ma (Krogh, 1993), the youngest date yet reported from the southern Kapuskasing zone.

Age relationships between prograde and retrograde metamorphism are clearly illustrated in this outcrop and the relative order of events is faithfully reflected in absolute zircon ages. Similar relationships observed at Stop 1-3 have correspond ages older by 20 and 10 m.y. At face value, this pattern suggests that similar processes occurred at progressively deeper levels with time. Two explanations have been offered to account for the observation of apparent downward younging (Fig. 9):

(1) Krogh (1993) proposed that mafic rocks were successively underplated after 2660 Ma and the prograde zircon growth records the emplacement and heating of these rocks in the lower crust. By implication, much of the lower crust has been subcreted following stabilization of high crust-level greenstone terranes.

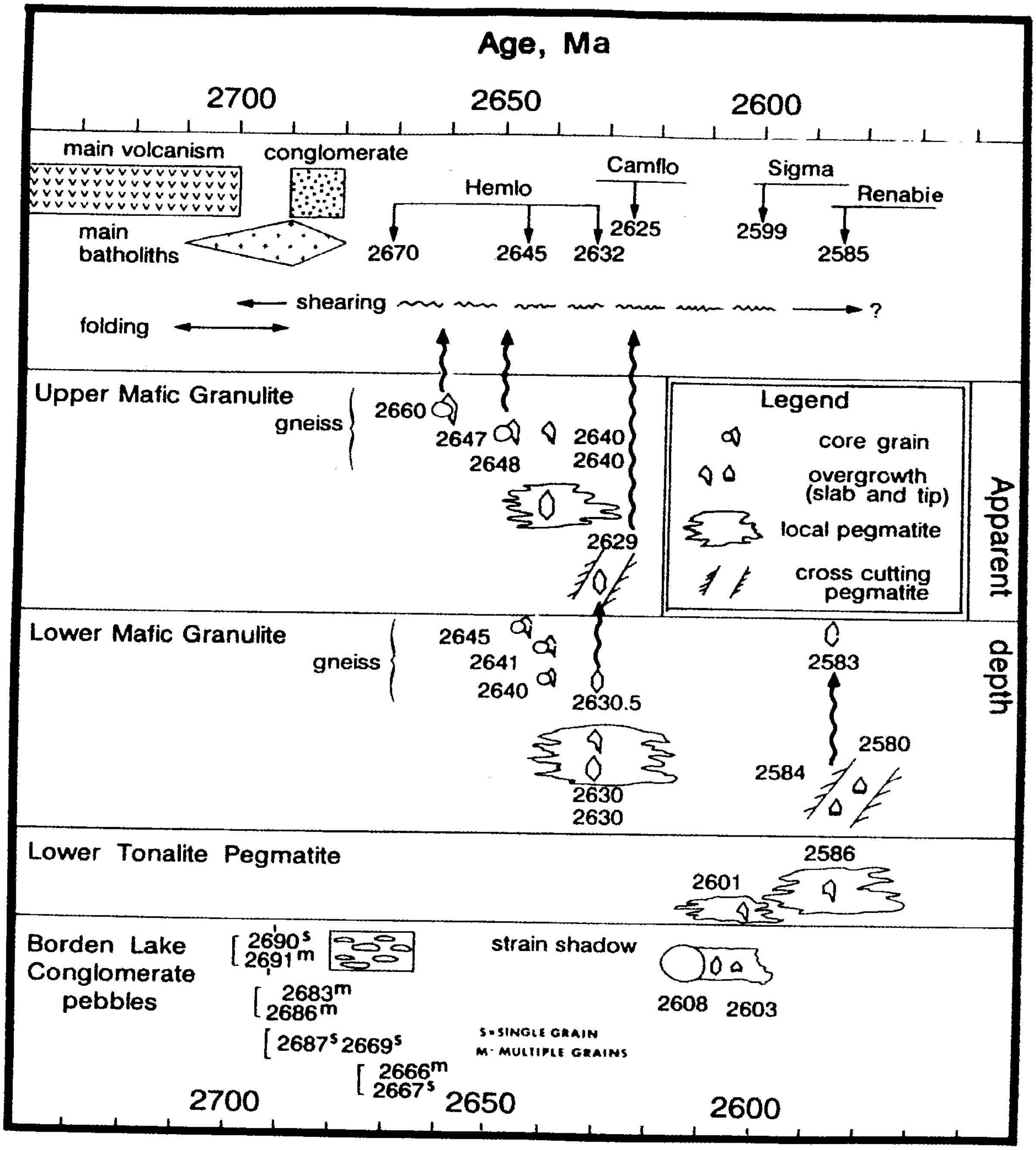


Figure 9. Summary diagram showing the distribution of U-Pb zircon ages with apparent depth in the Kapuskasing uplift and their relationship to the age of high-level gold deposits and the Borden lake conglomerate (after Krogh, 1993).

(2) Percival and West (1994) interpreted the apparent younging with depth as due to a downward-migrating ductility front over the interval 2690-2500 Ma. The crust cooled slowly following major accretion, magmatic and metamorphic events at 2700-2660 Ma. While shallow levels were stabilized rapidly, progressively deeper levels remained ductile longer and were subject to continuous recrystallization during late tectonic adjustments.

67.4 km

Stop 1-8:

Cataclasite and pseudotachylite of Ivanhoe Lake fault zone

(375846; 5328323)

In this area the Ivanhoe Lake fault zone has two strands; here the exposed western strand consists of cataclasite, pseudotachylite and many small brittle faults developed in mafic and tonalitic gneiss. Lenticles of dark green pseudotachylite up to 2 cm wide are cut by later, brittle faults. Fault-related features decrease toward the western end of the outcrop.

Uplift of the Kapuskasing zone probably occurred in several steps (Percival and West, 1994), with the latest, ~17 km at ~1.9 Ga. Evidence for the timing of uplift is indirect, based on the distribution of cooling ages (Percival and Peterman, 1994; Percival and West, 1994), and rotation and uplift of both 2.45 Ga Matachewan (*cf.* Halls *et al.*, 1994) and 2.04 Ga Kapuskasing (Lee *et al.*, 1990) dyke swarms (Percival *et al.*, 1994). Direct Ar dating of whole-rock pseudotachylite produced an average age of about 1750 Ma, which is probably too young as a result of devitrification (Fig. 10), although the highest-temperature step is closer to 1950 Ma. Crustal temperatures at 17 km and above at ca. 1.9 Ga would have been below ~300°C, accounting for the predominantly brittle nature of fault rocks in the Ivanhoe Lake fault zone. Evidence for earlier, ductile fault movement is observed on the eastern strand of the Ivanhoe Lake fault zone.

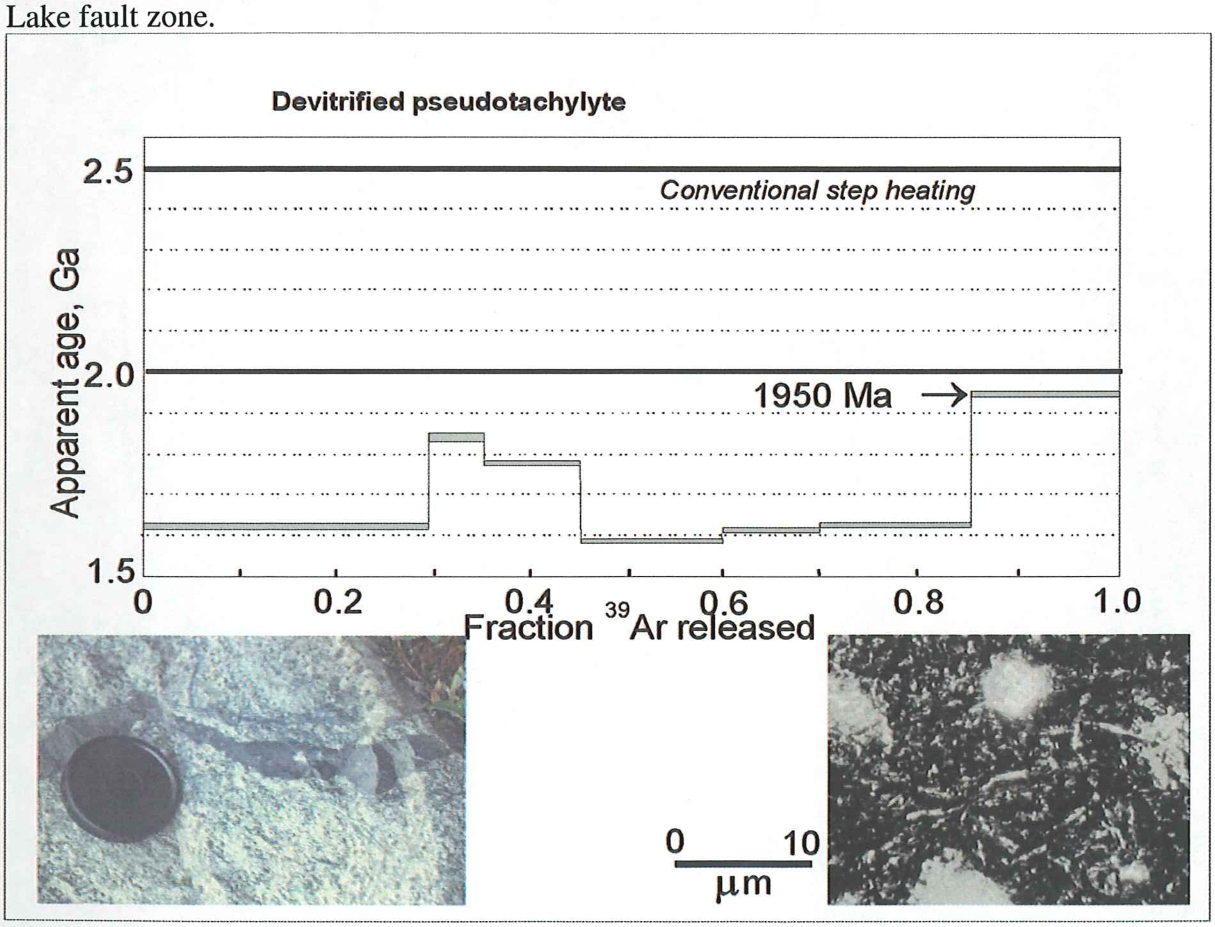


Figure 10. $^{40}Ar/^{39}Ar$ spectrum for pseudotachylite from the Ivanhoe Lake cataclastic zone. The disturbed spectrum indicates a Paleoproterozoic age, possibly around 1950 Ma, with younger resetting.

70.6 km: turn left on Warren-Carty Road (main hauling road)

78.9 km: turn right on narrow forest access road

79.5 km:

Stop 1-9:

Deep-crustal reflectors; trenched mafic-felsic gneiss sequence

(371105; 5332796)

This outcrop area was trenched to improve exposure following acquisition of Lithoprobe seismic reflection data (Percival *et al.*, 1989; Geis *et al.*, 1990), which showed typical deep-crustal reflections extending to within a few hundred metres of the surface (Fig. 5c). Natural exposure in this area is a few percent, so outcrop enhancement was performed, and additional high-resolution seismic studies were undertaken (Milkereit *et al.*, 1991; Fig. 11). The prominent reflections appear to correspond in general to interlayered mafic and tonalitic gneisses exposed over this large outcrop area. Laboratory-measured seismic velocities (Fountain *et al.*, 1990; White *et al.*, 1992) of 10-30-m-thick mafic and felsic units yield substantial reflection coefficients at an appropriate scale to generate the observed reflections.

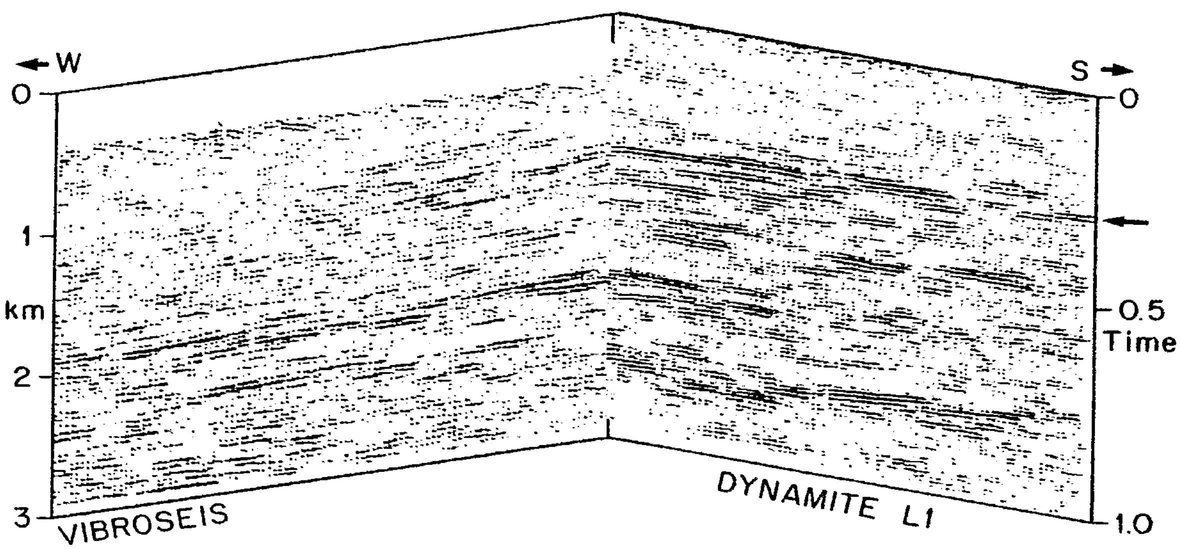


Figure 11. Fence diagram showing intersecting dynamite and Vibroseis high-resolution reflection profiles in the vicinity of Stop 1-9 (after Milkereit et al., 1991).

The outcrops consist of enclaves and layers of mafic gneiss of variable composition in hornblende-biotite ± garnet tonalitic gneiss (Fig. 12). One of the commonly inferred causes of deep-crustal reflectivity is mafic sills in felsic host rocks. This outcrop displays similar lithological diversity, but with clearly reversed age relationships (older mafic rocks in younger tonalite). Before drawing firm conclusions about the cause of deep-crustal reflectivity it is worth noting that the exact correlation between reflectors and surface geology remains somewhat enigmatic. Dips of layering at the surface are highly variable between 12 and 45° north, being folded about gently southwest-plunging, open folds associated with a prominent rodding lineation. In contrast, the reflections are very planar with a dip (17°; Kim *et al.*, 1992) closer to the attitude of the linear structures (7-20°) observed at the surface. A further unknown is the nature of structures in the exposure gaps, which are deep valleys filled with up to 100 m of overburden (Milkereit *et al.*, 1991). If these valleys coincide with faults or shear zones, these structures could also be a cause of the observed reflectivity.

Mafic rocks in this area provide P-T estimates in the 10 kbar, 750°C range (Mader *et al.*, 1994). Assemblages in D_4 ductile shears give pressures lower by ~ 0.5 kbar (Hartel and Pattison, 1996), suggesting that 1-2 km of uplift may have accompanied this deformation. Some of the youngest zircon (2615 Ma) and titanite (2493 Ma) ages are derived from mafic gneisses in this area (Krogh and Moser, 1994; Percival and West, 1994). The region is both at the deepest part of the structural section and closest to the Ivanhoe Lake fault zone, giving rise to alternate interpretations of the young ages as either related to fluid movements along the Ivanhoe lake fault zone (Krogh and Moser, 1994), or continued recording of late tectonic adjustments at depth (Percival and West, 1994).

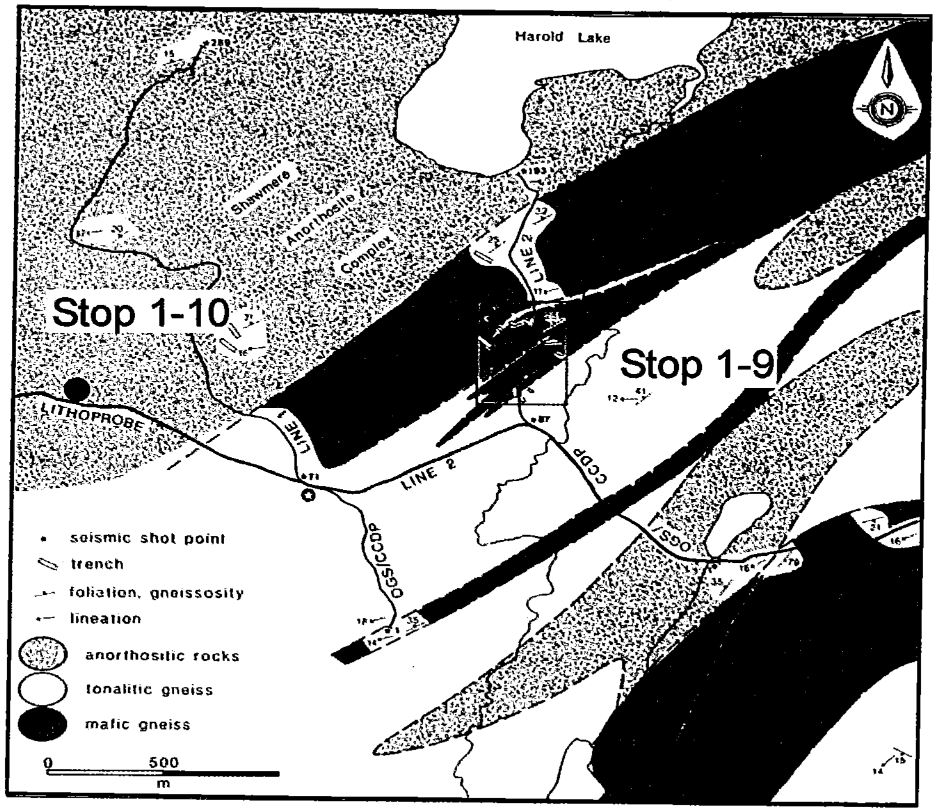


Figure 12. Detailed outcrop map of the region corresponding to the surface projection of prominent seismic reflections (after Percival et al., 1990).

Return to Warren-Carty Road; continue west for 4.6 km

Stop 1-10:

Shawmere anorthosite complex

(366937; 5332566)

The Shawmere anorthosite complex consists of a marginal layered phase and large homogeneous central phase, of which the present outcrop is representative. It consists mainly of calcic plagioclase (An_{90}) and minor hornblende. Other parts of the complex preserve igneous textures, including football-sized plagioclase megacrysts and large orthopyroxene and clinopyroxene crystals. Deformation is heterogeneous and local high-strain zones are abundant. Late alteration veins (narrow, chalky white zones) resemble brittle fractures and may relate to the Ivanhoe Lake fault zone. From mineral assemblages and fluid inclusion barometry, the veins appear to have formed at depth during isobaric cooling (Lamb and Morrison, 1997).

The age of the Shawmere complex is not well constrained. Percival and Krogh (1983) obtained a minimum U-Pb zircon age of 2765 Ma on a garnetiferous tonalite enclosed by the Shawmere complex. Zircon from a quartz-bearing phase of anorthosite yielded an age of 2650 Ma, close to metamorphic ages of mafic gneiss from the same region.

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Part 2.

Timmins area

Timmins, Porcupine Mining Camp, and the Southern Volcanic Zone of the Abitibi Greenstone Belt

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The Porcupine Camp: oblique view from helicopter towards northeast, approximately along the trace of the Destor-Porcupine Fault Zone. Dome Mine in foreground, while the village of South Porcupine and Porcupine Lake are visible in the background. Photo shows beginning of the Dome "superpit" development; the old Dome Mine headframe (center) no longer exists.

In 1909, the history of the Porcupine Mining Camp began with the discovery of visible gold on a "dome-shaped" outcrop of quartz in the right foreground of the photo.

INTRODUCTION

As part of the Deep Seismic Conference field trip, a portion of the Archean **Abitibi greenstone belt** will be visited on the way to the oblique crustal section of the Kapuskasing Zone (Day 2, Sunday), the Sudbury Structure (Day 3, Monday), and the Grenville Front (Day 4, Tuesday). This visit to the western Abitibi is timely as regional and high-resolution seismic lines were recently acquired across the Timmins area, to help resolve the geometry of this complex but prodigiously mineralized greenstone terrain. If available, preliminary sections may be shown and discussed on the field trip. Previous seismic investigations of the Abitibi along sections further to the east were published by Green et al. (1990) and Calvert et al. (1995).

The Abitibi greenstone belt (MERQ-OGS, 1983; Ludden et al., 1986; Jackson and Fyon, 1991; Corfu, 1993) is the largest preserved greenstone belt in the world, trending generally east-west across the southern **Superior craton**. In the east, it is truncated by the ca. 1 Ga **Grenville Front** tectonic zone (Day 4), whereas in the west, the Abitibi belt proper is truncated by a major Paleoproterozoic intracratonic thrust, the **Kapuskasing Zone** (Fig. 1). The latter exposes a gently west-dipping crustal section. Further west within the Kapuskasing Zone, at high structural levels within the dipping crustal section and approximately on strike, the continuation of the Abitibi greenstone belt reappears as the Wawa subprovince. Hence, the Abitibi greenstone belt is merely a remnant from an originally much larger Neoarchean granite-greenstone terrain that records the rapid crustal growth of Superior-type crust in the Neoarchean (e.g., Card, 1990; Percival et al., 2004).

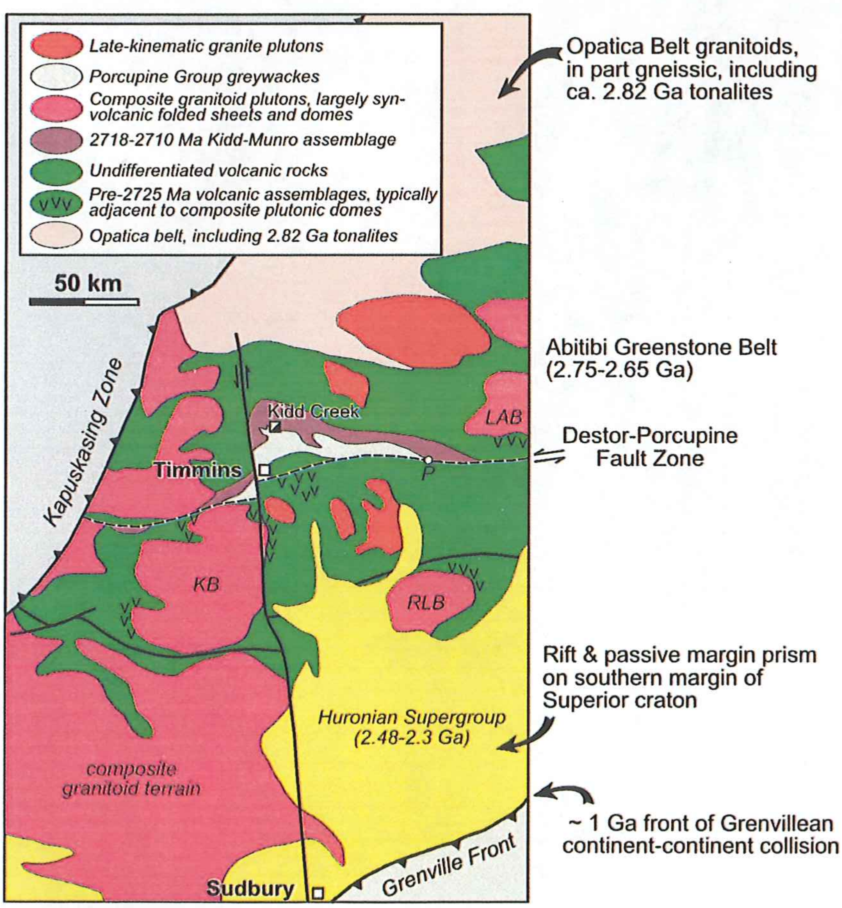
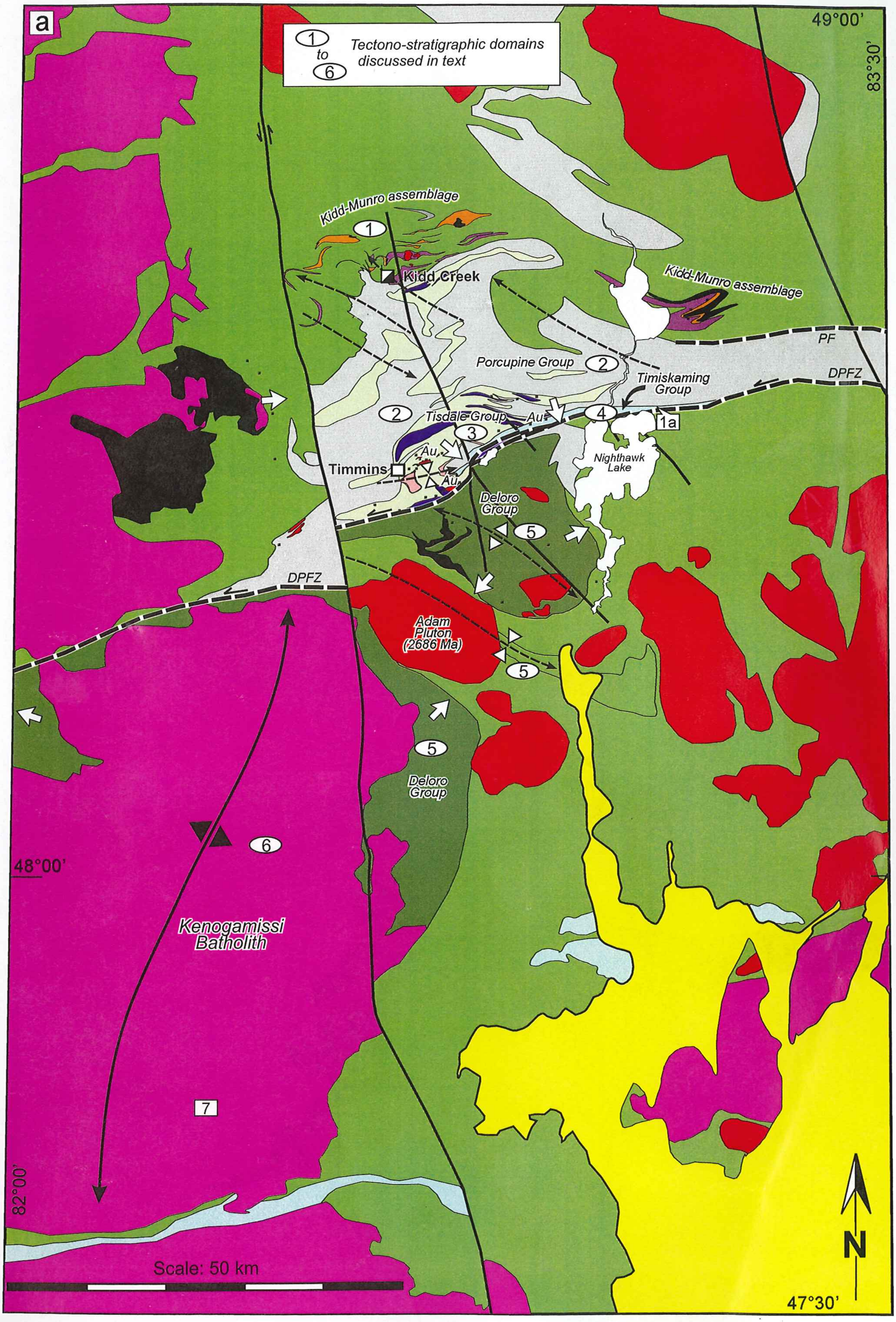
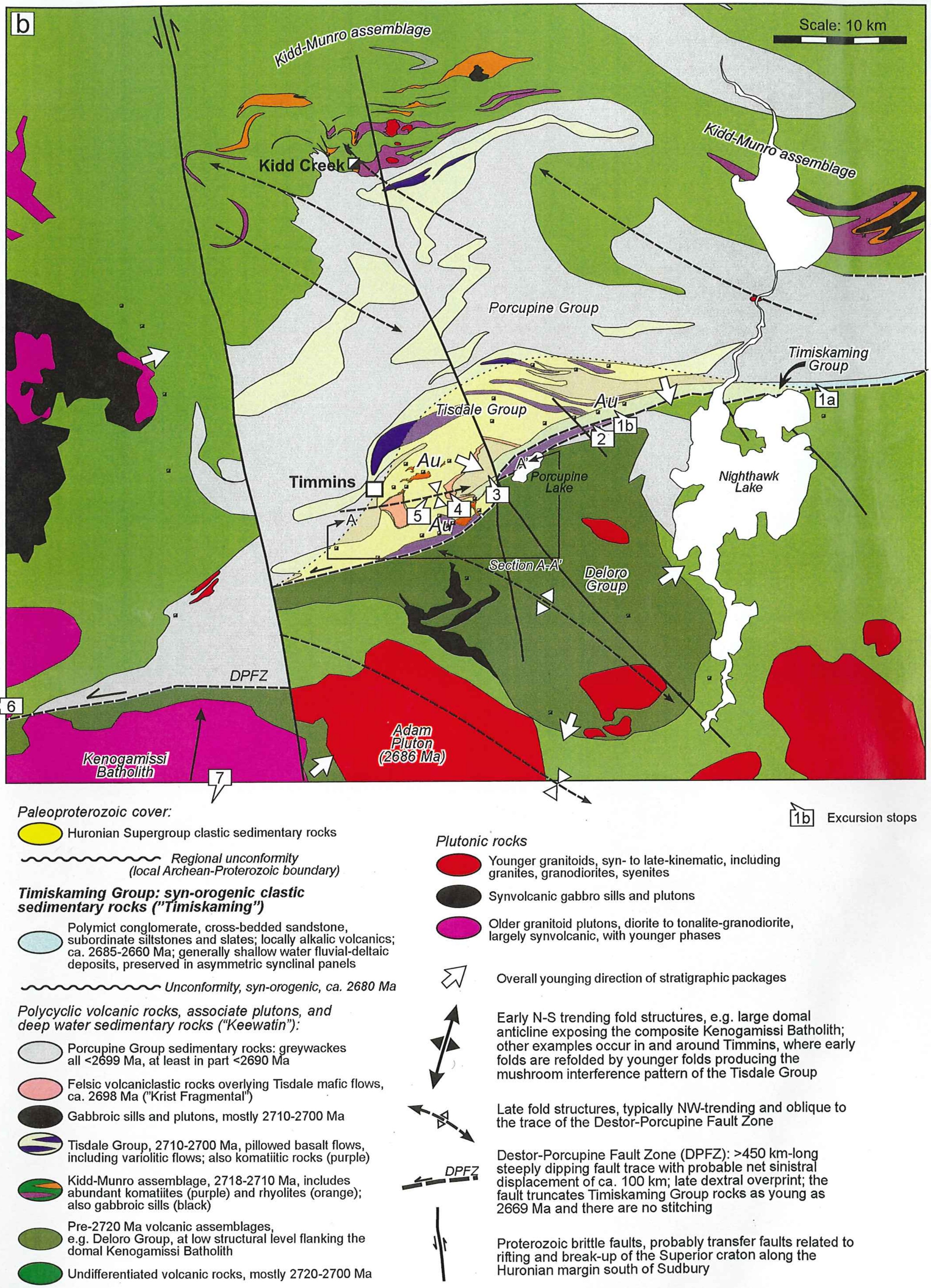


Figure 1: Schematic map of parts of the Superior craton and the Abitibi greenstone belt, from the ca. 1 Ga Grenville Front in the southeast to the Kapuskasing Zone in the northwest. The geology of the Archean Abitibi greenstone belt near Timmins is dominated by a large east-west trending fault zone, the Destor Porcupine Fault Zone. Truncation of the Porcupine Group and Kidd-Munro assemblage yields a piercing point (point P on map). The conjugate of this point, on the south side of the fault, occurs ca. 100 km to the east in Quebec (off the map). Hence, map relationships suggest the Destor-Porcupine Fault Zone is a large sinistral strike-slip fault. Abbreviations: KB, Kenogamissi Batholith; RLB, Round Lake Batholith; LAB, Lake Abitibi batholith.

The area in and around the mining town of Timmins, within the "southern volcanic zone" of the Abitibi subprovince (Ludden et al., 1986), has traditionally been referred to as the



Timmins, Fig. 2a



Timmins, Fig. 2b

Porcupine Mining Camp. This name traces its origin to the discovery of gold at the beginning of the 20th century around the shores of Porcupine Lake (see some historical notes appended at the end of this section). The camp not only hosts several **world-class gold deposits** (e.g., Hollinger-McIntyre, Dome), but 35 km to the north also one of the largest volcanogenic massive sulphide deposits in the world, the ca. 150 million tonne Kidd Creek deposit (Figs. 1, 2; Bleeker, 1999).

The area is a typical late Archean "granite-greenstone terrain", characterized by polycyclic volcanic stratigraphy, overlain by a late stage sequence of turbiditic sedimentary rocks and finally syn-orogenic clastic rocks. These rocks were intruded by numerous granitoid plutons and deformed by several phases of folding and faulting. Isotopically, the volcanic rocks are largely juvenile (e.g., Corfu and Noble, 1992; Ayer et al., 2002), although there are rare and subtle traces of ca. 2.8-2.9 Ga basement involvement (e.g., Barrie and Davis, 1990; see brief review and references in Bleeker, 2002), direct examples of which are exposed in the Kapuskasing Zone (Moser et al., 1996), and to the north of the Abitibi in the Opatica gneiss belt (e.g., Davis et al., 1995).

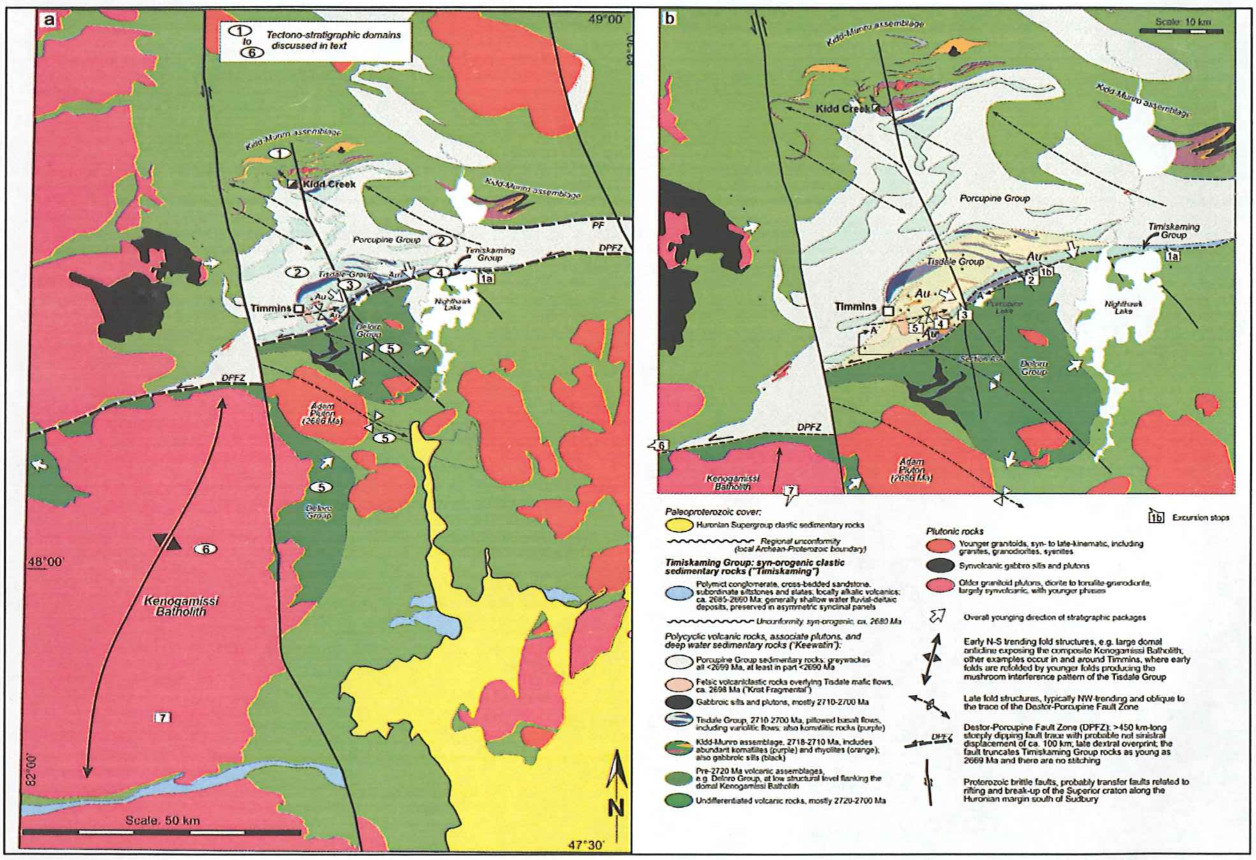
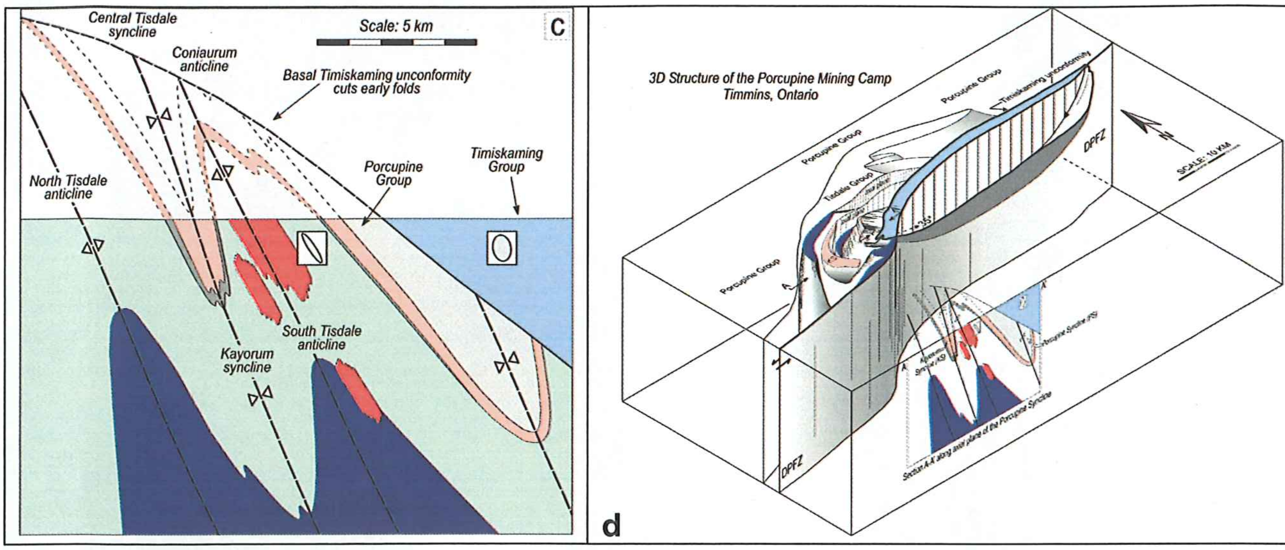


Figure 2: Simplified geological map of the Abitibi greenstone belt in and around Timmins. a) Overview map; numbers 1 to 6 are specific tectono-stratigraphic domains discussed in text. b) Enlarged portion of the map around Timmins, annotated with excursion stops 1 to 7. Section A-A', parallel to the axial plane of the Porcupine Syncline, is shown in Fig. 2c. c) West-to-east structural cross-section A-A', parallel to the axial plane of the Porcupine Syncline, to highlight the importance of early west-vergent folds in the overall interference structure. Both the Porcupine Syncline and the earlier folds predate the Timiskaming unconformity. d) Block diagram illustrating the 3D structure of the Porcupine Camp, as seen from the southwest. Note the preservation of the Timiskaming rocks in a synclinal panel along the Destor-Porcupine Fault Zone. The southern half of this synclinal structure has been removed by subsequent fault movement.



LOCAL GEOLOGY

To appreciate the geology of the Porcupine Camp and its complexities, some knowledge of the local stratigraphy is required. A brief stratigraphic outline (e.g., Fig. 3) will also help to put the excursion stops into context. It is useful, however, to first introduce one of the major fault structures in the area, as it controls the distribution of some of the stratigraphic packages.

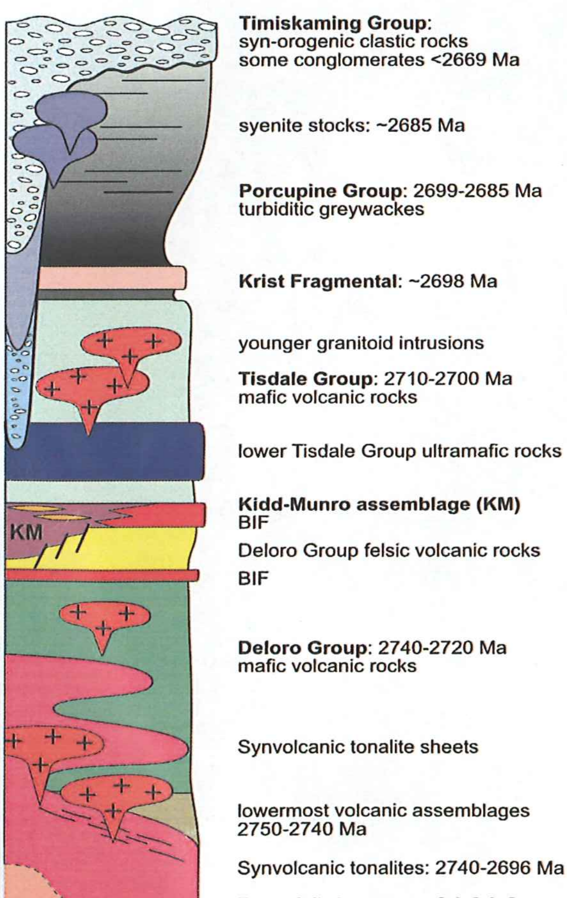


Figure 3: Simplified stratigraphic column for the Timmins area, highlighting units discussed in the text.

Destor-Porcupine Fault Zone

The geology of the "southern volcanic zone" is dominated by a major regional fault zone or structural-stratigraphic "break", which is referred to as the **Destor-Porcupine Fault Zone** (DPFZ; Figs. 1, 2). This east-west trending fault can be traced from the Grenville Front to the Kapuskasing Zone (and beyond to the west; see, for instance, Leclair et al., 1993) and is therefore >>450 km long. The characteristics of this fault zone are:

1. A well-defined fault trace that is well over 450 km long.
2. At surface, where the fault can be recognized, the structure appears generally steep.
3. There are few, if any, systematic differences in metamorphic grade across the fault (e.g., Powell et al., 1993).
4. Although broadly similar rock assemblages

occur on either side, nowhere can rocks be linked directly across the fault (e.g., see map of Figs. 1, 2), thus suggesting significant displacement.

5. Synclinally infolded panels of synorogenic clastic rocks are truncated by and occur along the fault (e.g., Timiskaming Group, Fig. 2), thus requiring that movement was late in the structural history.
6. This inference is further supported by the observation that, despite an abundance of plutonic stocks, there are no known plutons that stitch the fault; i.e. fault movement outlasted upper crustal plutonism.
7. Some late fold generations and the regional cleavage are oblique and clockwise to the fault (Fig. 2a; see also Hubert et al., 1984).
8. Gold mineralization is concentrated along the fault zone, particularly in splays of the main fault.
9. Major gold concentrations tend to occur in S-shaped bends of the main fault, where the fault swings to a southwesterly trend (Bleeker, 1995).
10. Kinematic information is conflicting; latest movement may have been dextral.

Using empirical scaling observations, the length of the fault trace strongly argues for a large-offset principle displacement fault. This is supported by general inability to correlate rocks directly across in any particular area. If these inferences are correct, the lack of systematic differences in metamorphic grade and (or) crustal level across the fault eliminates significant dip-slip movement at a late stage. Hence, straightforward structural reasoning leads to the conclusion that the Destor-Porcupine Fault Zone must represent a **major strike-slip fault**. Modern analogues would be trench-parallel strike-slip faults in plate margins with oblique convergence and fault partitioning (e.g., Jarrard, 1986).

The next question is then: what is the magnitude and sense of the total displacement? As kinematic studies along ancient crustal faults often yield ambiguous or conflicting results, the most robust indicators of net displacement are markers on either side of the fault that can be correlated with confidence. A search for such markers (piercing point P, Fig. 1, and its conjugate 100 km to the east) yields ca. 100 km of net sinistral displacement on the Destor-Porcupine Fault Zone (Bleeker, unpublished data). A displacement of this magnitude qualitatively satisfies many of the observations listed above (scaling arguments, no easy correlation across the fault, oblique en-échelon folds) and would suggest that major gold mineralization of the Porcupine camp occurs in a dilational bend of the Destor-Porcupine Fault Zone (Fig. 2b). This interpretation does not preclude that the main fault zone may have utilized parts of earlier thrusts, or may have been overprinted locally by younger dip-slip faults. In many areas, particularly in highly fissile schists along the fault zone, there is evidence for younger northeast-trending crenulations that probably relate to a late-stage dextral reactivation of the fault zone.

Hence, the Destor-Porcupine Fault Zone, similar to many other ancient crustal fault zones, likely had a complex history. Nevertheless, first-order regional relationships strongly argue for large, sinistral, strike-slip movements late in the overall structural evolution of the Abitibi greenstone belt.

Tectono-stratigraphic domains and regional stratigraphy

It is easiest to discuss the general and simplified stratigraphy in terms of about six structural-stratigraphic domains. From north to south, these are (see also Fig. 2a):

1. An east-west trending belt of 2720-2710 Ma "trimodal" volcanics (komatiites, basalts, rhyolites), the **Kidd-Munro assemblage**. North of Timmins, these rocks host the giant Kidd-Creek Cu-Zn-Ag deposit associated with a long-lived 2717-2710 Ma rhyolite volcanic center (Bleeker and Parrish, 1996; Bleeker, 1999; Bleeker et al., 1999).

2. An east-west trending belt of regionally extensive greywacke turbidites, the ca. 2699-2685 Ma **Porcupine Group** (Bleeker and Parrish, 1996 and reference therein). These deepwater sedimentary rocks are thought to stratigraphically overlie the Kidd-Munro assemblage as well as other volcanic rocks in the area as the greywackes contain detrital zircons from the various volcanic packages. To the east, these rocks are truncated by the Destor-Porcupine Fault Zone (Fig. 1) and are thought to reappear in Quebec as the Kewagama Group, ca. 100 km further east. This robust correlation constrains the net offset on the fault as ca. 100 km, sinsitral.

3. A thick ultramafic to mafic flow sequence, the ca. 2710-2700 Ma **Tisdale Group**, which is (disconformably?) overlain by ca. 2698 Ma intermediate volcaniclastic rocks of the **Krist Fragmental** (Pyke, 1982). These intermediate volcaniclastic rocks occur around the Porcupine Syncline and in reappear in the core of the Kayorum Syncline (Fig. 2b, c and d). In turn, the Krist Fragmental unit is conformably overlain by the turbiditic greywackes of the Porcupine Group.

4. A thin, south-younging panel of coarse clastic sedimentary rocks, immediately north of the Destor-Porcupine Faultz Zone. This sequence is known as the ca. 2685-2660 Ma **Timiskaming Group**. The panel of Timiskaming clastic rocks preserves parts of a deeply infolded syncline that on its southern side appears truncated by the fault (Fig. 2b, d). Along the northern boundary of the stratigraphic panel, the Timiskaming Group clastic rocks unconformably overlie Tisdale Group volcanic rocks and Porcupine Group greywackes that were previously folded. Although Timiskaming-type sedimentation probably started around 2685 Ma in some areas (Ayer et al., 2002), the youngest concordant detrital zircon from Timiskaming conglomerates east of Timmins has been dated at 2669 Ma (see Fig. 4; Bleeker, unpublished data). Hence, these "syn-orogenic" clastic rocks were deposited at least in part after 2669 Ma. Significant gold mineralization is hosted by the Timiskaming Group (e.g., Fig. 5b) and therefore gold lode formation must largely post-date deposition of these syn-orogenic clastic rocks.

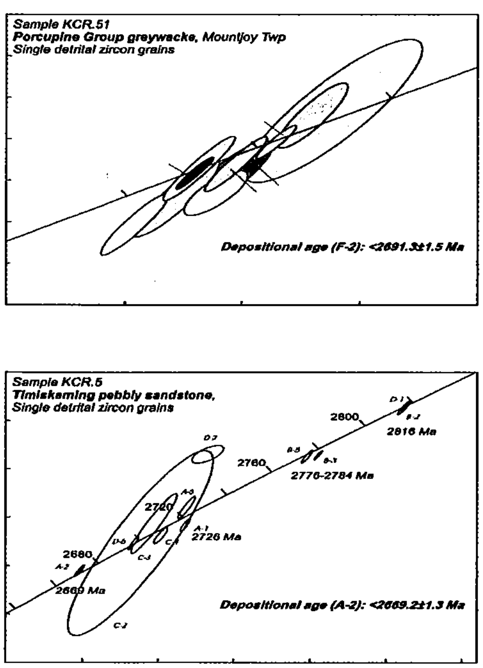


Figure 4: Single-grain detrital zircon U-Pb age determinations for two important samples from the Timmins area (Bleeker, unpublished data). a) 10 analyses from a sample of Porcupine Group greywacke, from just west of Timmins. Note the typical pattern seen in Porcupine Group samples (Bleeker and Parrish, 1996; Bleeker, et al. 1999) of a variety of zircon grains showing different ages that correspond to known volcanic rocks and plutonic ages in the surrounding greenstone terrain. Youngest concordant analysis (grain F-2) yields a precise age of 2691.3±1.5 Ma, providing a maximum age of deposition for the Porcupine Group turbiditic greywackes. b) Similar data for Timiskaming Group pebbly sandstone. Note the broader age range typical for younger syn-orogenic sandstones, including grains from more deeply eroded hinterland yielding 2816 Ma ages. The youngest concordant grain (A-2) with a precise age of 2669±1.3 Ma, yields a maximum age of deposition for this Timiskaming sandstone.

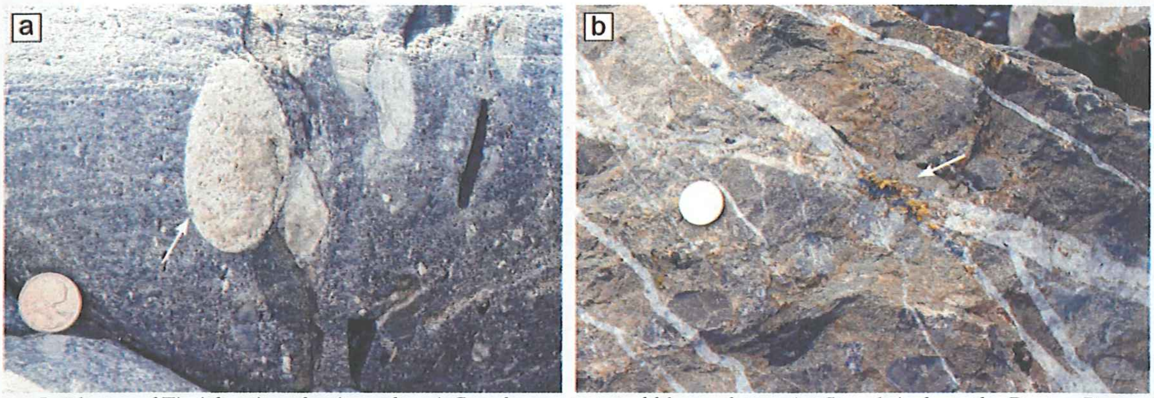


Figure 5: Photos of Timiskaming clastic rocks. a) Conglomerate to pebbly sandstone (at Stop 1a) along the Destor-Porcupine Fault Zone. Note well-rounded granitoid pebble with pre-depositional foliation (at arrow), indicating relatively deep erosion just prior to deposition of the syn-orogenic clastic rocks. b) Altered basal conglomerate (from pits near Stop 1b), with quartz veins containing spectacular free gold (at arrow).

5. The geology south of the Destor-Porcupine Fault Zone is dominated by a large domain of folded volcanic stratigraphy in which Tisdale-age equivalent rocks overlie an older, ca. 2740-2723 Ma volcanic package known as the **Deloro Group** (Pyke, 1982). The top of the Deloro Group is marked by felsic volcanics and a banded iron formation (BIF) of regional extent. These distinctive stratigraphy features are repeated on either side of the large Kenogamissi Batholith, always facing outwards, thus demonstrating preservation of regional stratigraphic relationships over large areas (see also Heather et al., 1995; Ayer et al., 2002).

6. At the lowest structural and stratigraphic levels, the Deloro Group is intruded by a multiphase tonalite-trondhjemite-granodiorite (TTG) plutonic complex, the **Kenogamissi Batholith** (Figs. 2a, 3). This batholith, with individual phases ranging in age from ca. 2725 Ma to 2682 Ma, is exposed in the core of a broad, north-south trending anticlinal structure separating the supracrustals in the Timmins area from correlative rocks further to the west. Many of the plutonic phases in the Kenogamissi Batholith are of similar age to the volcanic packages, and in part represent the plutonic roots of volcanic edifices that produced the volcanic rocks.

The main stratigraphic units and their relationships are further illustrated in the schematic column of Figure 3. Detailed examinations of the stratigraphic relationships between individual packages, aided by precise zircon dating, and dating of inherited and detrital zircons, has shown that, despite significant deformation, the inferred stratigraphic stacking of multiple volcanic packages, finally overlain by deep water sedimentary rocks, is largely primary, spanning a total age range from ca. 2750 Ma to 2685 Ma. If true, popular models of micro-terrane accretion, in which each individual volcanic package is interpreted as a remnant of an arc or an oceanic plateau (e.g., Desrochers et al., 1993), must be incorrect. As mentioned earlier, some of the synvolcanic plutonic rocks show rare evidence for involvement of somewhat older basement.

Field Trip

Purpose

Field trip stops[2] in the Timmins area were selected to provide a brief impression of some of the key stratigraphic and structural features, particularly in relation to seismic profiling. Interestingly, some of the same rocks will be observed later at deeper structural levels and

[2]Field trip stops and some of the background material are in part based on two earlier field trip guidebooks (Bleeker, 1995, 1997).

higher metamorphic grades within the Kapuskasing Zone. Hence, it is useful to compare both areas.

Fault Zone and the regional structural architecture on either side of the fault. Resolving the deep structure and regional architecture are important because a key feature of the Porcupine gold camp is that essentially all gold mineralization (60-70 million ounces) occurs in a lozenge-sh Recent seismic lines were shot to investigate the deep structure of the Destor-Porcupine aped area north of the main fault trace (Fig. 2b). This key observation (Bleeker, 1995) suggest two possible models, each with significant but different implications for future exploration:

1) *The strike-slip model*: the lozenge-shaped area may represent a dilational bend in a dominantly sinistral strike-slip fault system, further associated with favourable structural and lithological traps. If so, the asymmetry of the camp, with essentially all significant gold north of the main fault, is intriguing. Were potentially mineralized rocks south of the fault removed ca. 50-100 km to the east by late displacement on the Destor-Porcupine Fault Zone?

2) *The dipping fault model* (Fig. 6): alternatively, the major asymmetry in gold distribution may relate to a northerly dip of the fault system, with the Destor-Porcupine Fault Zone becoming listric to the north at depth. Indeed, a general north-dipping (south-vergent) structural asymmetry is also apparent from some other regional structural observations (e.g., the Timiskaming panel is steeply overturned towards the south). If so, the northern hanging wall of the main fault zone, with secondary splay faults above and tapping into a dipping plumbing system (Fig. 6), may be a more favourable target for exploration than the southern footwall.

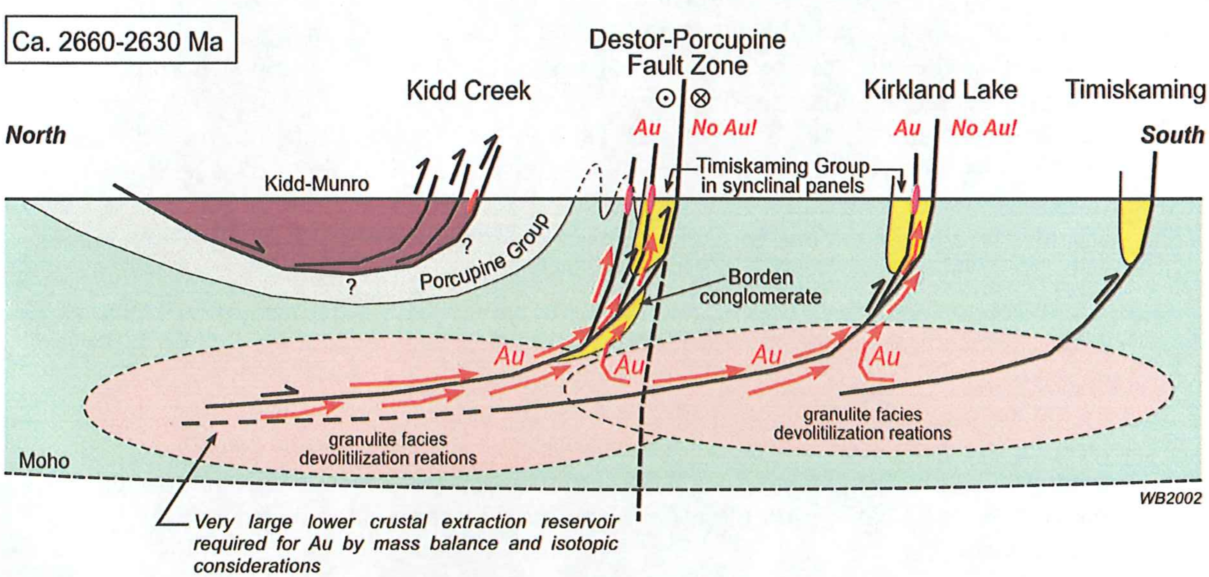


Figure 6: Cartoon cross-section, illustrating the "dipping fault model" and how it could explain the asymmetry in gold distribution across the Destor-Porcupine Fault Zone and a similar fault zone in Kirkland Lake to the southeast. Gold was mobilized by granulite facies devolatilization reactions in the lower crust (exposed in the Kapuskasing Zone), producing mixed H_2O-CO_2 fluids. These fluids were channeled and focused by major dipping fault structures, escaping to the upper crust. Here, they preferentially infiltrated along secondary structures in the hanging walls of the main faults. Note that a dipping listric fault has excellent potential to tap a large-volume lower crustal extraction reservoir. The synclinal panels of Timiskaming clastic rocks are shown in yellow. Seismic profiling can test several key elements of this hypothetical model section. Another key point illustrated by this section is that the tight infolding and overturning to the south (i.e., south vergence) must post-date deposition of uppermost Timiskaming conglomerates and thus post-date 2669 Ma. The "Borden conglomerate" of the Kapuskasing Zone (also with 2669 Ma detrital zircons!) shows that similar aged conglomerates were transported, on north dipping structures, all the way into the deep crust. This late in the structural evolution, this significant structural fabric is therefore unlikely to be related to accretion, but rather to late stage north-south shortening, perhaps as a far-field response to final collision. This has important implications for how to interpret other Abitibi seismic lines (e.g., the main Abitibi line interpreted by Calvert et al., 1995).

Of course a hybrid model, involving a dipping plumbing system overprinted by late-stage strike-slip faulting, may be equally possible. In both models, the ultimate source of the Au bearing fluids is thought to be deep crustal, regional, granulite grade metamorphic devolatilization of a thickened crustal pile, involving abundant juvenile mafic and ultramafic rocks (with Au in trace amounts of sulphides?). It is hoped that the recently acquired seismic lines will resolve some of the geometrical questions.

Another important point of potential discussion, also illustrated by Figure 6, is the tight infolding and overturning to the south of the Timiskaming conglomerates. This south vergent structural pattern must post-date Timiskaming deposition and thus post-date 2669 Ma. The very similar "Borden conglomerate" of the Kapuskasing Zone (also with 2669 Ma detrital zircons!) shows that some Timiskaming conglomerates were transported on north-dipping structures all the way into the lower crust. This significant structural event and the south-vergent structures are therefore this unlikely to be related to early accretion, but rather to late stage north-south shortening, perhaps in response to final collision. This has implications for interpretation of other Abitibi seismic lines (e.g., Calvert et al., 1995).

Field trip stops

We will try to make the following short stops driving through Timmins, on our way from Tremblant to Chapleau. Details will depend on how much time is available. Stops are indicated on the map of Figure 2b:

Stop 1a: Sandstone and pebbly conglomerate of the Timiskaming Group panel, east of Timmins.
Stop 1b: Steeply overturned Timiskaming clastic rocks just north of the Destor-Porcupine Fault Zone, near the Pamour gold mine.
Stop 2: Mafic schists along the Destor-Porcupine Faults Zone.
Stop 3: Porcupine Group greywackes in the core of the Porcupine Syncline.
Stop 4: The Krist Fragmental intermediate volcaniclastic unit, near the hinge of the Porcupine Syncline.
Stop 5: Deformed pillow lavas of the Tisdale Group in the structural saddle between the Porcupine and Kayorum synclines.
Stop 6: A superb outcrop of mafic schist along the Destor-Porcupine Fault Zone, to the west of Timmins.
Stop 7: Gneissic tonalite of the Kenogamissi Batholith along Highway 144 from Timmins to Sudbury (on Day 3).

Descriptions of field trip stops

Stop 1a:

Sandstone and pebbly conglomerate of the Timiskaming Group panel, east of Timmins

A spectacular outcrop of coarse clastic Timiskaming Group sedimentary rocks is present along the north-side of the highway, just west of the turn-off to Iroquois Falls. Polymict pebbly sandstones alternate with conglomerates. Meter-scale cross-bedding is conspicuous throughout the outcrop and provides unequivocal younging directions. The vertically dipping beds young towards the south, as is typical for the Timiskaming panel east of Timmins as a whole. A wide variety of clasts can be observed, representing volcanic, sedimentary, as well as plutonic rock types. Particularly noticeable are bright-green, fuchsite-altered mafic to ultramafic volcanic clasts. Some ultramafic clasts preserve spinifex textures. Besides a variety of porphyry clasts, medium-grained granitoid clasts are also common. At least one of the granitoid clasts shows a fabric (Fig. 5a). Coarse grained ranitoid clasts are important because they imply significant uplift, erosion, and unroofing of plutons just prior to deposition of the Timiskaming clastics. Detrital zircons from this outcrop were dated, the youngest concordant grain yielding an age of 2669 Ma (Fig. 4).

Stop 1b:

Timiskaming greywackes and overlying cross-bedded sandstones

This stop examines Timiskaming rocks further west, just past one of the major gold mines in the area, the Pamour Mine, hosted in part by quartz veins within basal Timiskaming conglomerate and overlying siltstones and greywackes. An outcrop of the thinly bedded siltstones near the Pamour Mine used to show thinly layered Timiskaming sequence greywacke turbidites, younging towards the south and folded into a sinistral reclined fold. A strong cleavage refracted through the layering and was axial planar to the fold. This cleavage is the regional cleavage in the camp and this locality demonstrated that it was axial planar to reclined folds in the Timiskaming Groups, and hence associated with late-stage fold structures (F4) with sinistral asymmetry.

Unfortunately, this outcrop may no longer be available. However, ca. 100 m south along the highway occurs a large outcrop of Timiskaming Group pebbly sandstones. Again, ubiquitous cross-bedding and pebble lags demonstrate that the steeply north dipping beds (60-75°) young to the south and are steeply overturned (see cartoon in Fig. 6). A cleavage can be seen which trends subparallel to bedding but dips shallower to the north and relates to the overturning of the beds and the overall Timiskaming synclinal structure. The sandstones show coarse foresets in places, and local pebble lags, and are interlayered with pebbly conglomerates. They must have been deposited in a high-energy, fluvial, braided stream to shallow marine deltaic environment. A variety of rock types, including granitoids, can be identified among the clasts.

Stop 2:

Mafic schists along the Destor-Porcupine Faults Zone

A small polished roadside outcrop on the north side of the highway, towards Porcupine, provides one of the few outcrops along the trace of the Destor-Porcupine Fault Zone. All that can be seen are highly fissile mafic schist, so strongly deformed that no protolith can be discerned. The foliation is essentially vertical. This outcrop is of interest in that it occurs immediately south of the Timiskaming panel of clastic rocks and thus demonstrates the spatial relationships (see Fig. 2b and d). A similar but much better exposed outcrop of the Destor-Porcupine Fault Zone occurs ca. 50 km west of Timmins.

Stop 3:

Porcupine Group greywackes in the core of the Porcupine Syncline

A small roadside outcrop in South Porcupine exposes turbiditic greywackes of the ca. 2685 Ma Porcupine Group, within the core of the east-plunging Porcupine Syncline. This quick stop helps in building the regional stratigraphic and structural picture.

Stop 2.2:
Felsic norite

The Sudbury Igneous Complex is composed of the Sublayer (contact sublayer and offset dikes) and the Main Mass. The Main Mass is further subdivided into 4 units, from base to top, these are the mafic norite, felsic norite, quartz gabbro and granophyre units. The felsic norite is a medium-grained to coarse-grained rock with hypidiomorphic equigranular texture. It is composed of 50 to 70% plagioclase (An_{50-65}), 3 to 10% potassium feldspar, 10 to 30% orthopyroxene, 5 to 15% granophyric quartz and 2 to 5% biotite, with mnor amounts of aoatite, zircon, ilmenite and sulphide minerals (Coats and Snajdr 1984).

Return to vehicle, continue southeast along Highway 144 for roughly 1.4 km to the junction of Regional Road 8. Turn left and park on the east side of the road. First, walk to the easternmost end of the high rock cut on the north side of Regional Road 8 to examine the quartz gabbro unit. Then cross the road to examine rocks of the granophyre.

Stop 4:
The Krist Fragmental intermediate volcaniclastic unit

A large outcrop on the south side of Highway 101, between South Porcupine and Timmins, exposes the intermediate Krist Fragmental volcaniclastic unit, stratigraphically between Porcupine Group greywackes (up section to the east, Stop 3), and Tisdale Group mafic flow (down section to the west, Stop 5). Angular feldpar-phyric volcanic clasts can be seen, as well as at least two cleavages. An early east-dipping fabric is axial planar to early west-verging folds (F1) in the volcano-sedimentary rocks of the Tisdale and Porcupine Groups. These early folds were refolded into a mushroom pattern by the east-trending Porcupine Syncline (F2), and both fold generations were later truncated by the basal Timiskaming unconformity. This can be easily seen from the truncation of the Krist Fragmental unit on both limbs of the Porcupine Syncline (e.g., Fig. 2b). Subsequently, the Timiskaming rocks themselves were tilted and folded in the south-verging syncline (F3) along the Destor-Porcupine Fault Zone, and this syncline was subsequently truncated. Finally, reclined folds with sinistral asymmetry further deform the already tilted Timiskaming strata (F4).

Stop 5:
Moderately deformed pillow lavas of the Tisdale Group

This stop allows examination of moderately deformed, ca. 2707 Ma, pillow lavas of the Tisdale Group, in the structural saddle between the Porcupine and Kayorum synclines. The pillows are stretched and flattened, plunging steeply to the east, parallel to the axial plane of the early east-verging folds. Another reason to stop here is to help built the stratigraphic picture and to show typical mafic pillow lavas at some distance from the Destor-Porcupine Fault Zone, in contrast with the mafic schists within the fault zone (stop 2 and next stop).

Stop 6:
Mafic schist along the Destor-Porcupine Fault Zone, to the west of Timmins

About halfway between Timmins and the Kapuskasing Zone, another rare outcrop of highly deformed mafic schists marking the Destor-Porcupine Fault Zone occurs along the north side of the highway. This outcrop allows better examination of the steeply dipping schistosity. Younger crenulations with subhorizontal axial planes and axes deform the main schistosity and may relate to late-stage horizontal relaxation and gravitational collapse. Such crenulations are seen throughout the area and may locally relate to large horizontal buckle folds.

Stop 7:
Gneissic tonalite of the Kenogamissi Batholith along Highway 144 from Timmins to Sudbury (on Day 3)

Outcrops across the road from the gas station and coffee bar along the highway near Gogama show hornblende-biotite tonalitic gneiss of the Kenogamissi Batholith with deformed gneissic foliation. This early phase of the batholith has been dated at 2725 Ma, with an inherited population of ca. 2735 Ma zircons (O. van Breemen, unpublished data). The gneissic tonalite probably represents tonalite sheets that intruded deep in the pile during Deloro Group volcanism. These tonalite sheets represent a deep structural level, transitional to what was observed in the oblique section of the Kapuskasing Zone. Here, this deep structural level is exposed because of the broad domal anticline centered on the Kenogamissi batholith. As in the Kapuskasing Zone, the interlayering of such tonalite sheets with mafic volcanics may cause much of the deep crustal reflectivity in granite-greenstone terranes.

Appendix: a few historical notes

The discovery of mineral deposits in northern Ontario has been intimately linked with construction of the "Temiskaming and Northern Ontario Railway" from the city of North Bay in the south to the clay belts of the north. The Cobalt Silver Camp had been discovered in 1903 and by 1907 was well established.

Attracted by the riches of the rapidly developing Cobalt camp, and the potential of making further discoveries in more remote areas to the north, prospectors travelling mostly by canoe reached the Nighthawk Lake area (Fig. 2) in 1905-1907. The siting of gold in 1907 on Gold Island in Nighthawk Lake, 25 km east of Timmins, is regarded as the first significant discovery in the camp (Doherty, 1986), although scattered reports of gold in the area existed prior to this date (e.g., see Barnes, 1991). A small staking rush followed and, soon after, gold was discovered on the northeast shore of Porcupine Lake. An increasing number of prospectors followed, making a number of smaller gold finds.

In the year 1909, a party of prospectors headed by John S. Wilson found a spectacular showing of gold on a dome-shaped outcrop of quartz, approximately 3 km southwest of Porcupine Lake. This discovery established the Dome Mine, which has been in continuous operation ever since 1910. To the end of 1993 the Dome Mine had produced 12.2 million ounces of gold and 2.2 million ounces of silver (Dome Mine geological staff, pers. comm., 1995). Since then, through development of the "superpit" and mining of remaining high grade gold lodes underground, several more million ounces have been mined.

Shortly following the Dome Mine discovery, in late October 1909, prospectors Benny Hollinger and Alec Gillies discovered significant gold showings a few kilometres further to the west, and staked what was to become the giant Hollinger Mine. Staking of the property that became the McIntyre Mine followed immediately after the Hollinger Mine discovery (Doherty, 1986). It is worth noting, however, that gold-bearing quartz veins in the general vicinity of the Hollinger Mine had been found earlier, in 1906, by prospector Reuben D'Aigle. He staked seven claims and, in 1907, he and his partners proceeded to excavate a number of small test pits. Although they spotted some gold, they evidently missed the spectacular free gold showings found in 1909 and decided to abandon their efforts and move on (Barnes, 1991).

Due to the large number of prospectors in the area, each significant discovery was immediately followed by a staking rush of the surrounding land. This led to a large number of independent properties and mines, some of which were amalgamated over the years. Between 1910 and 1924, between two and nine mines were in production. Disaster followed in 1911, when much of the infrastructure in the mining camp, including the community of South Porcupine, was destroyed by fires. As part of the reconstruction process, the town of Timmins was created just west of the Hollinger Mine. The new town was named after Noah Timmins, financier and co-owner of the Larose Mine in Cobalt, who shortly after the Hollinger Mine discovery acquired a controlling interest in the property (see Barnes, 1991).

Ever since these early days of discovery, Timmins has lived through ups and downs associated with events affecting the gold market: wars, the depression, instatement or re-instatement and eventual abandonment of the gold standard among others. In the early 1960s, gold mining in Timmins was in a down turn. Post-war developments of declining prices and increasing labour costs forced many of the gold mines to close down. For instance, the Buffalo Ankerite Mine closed in 1953, the Coniaurum Mine in 1961, the Delnite Mine in 1964, and the Paymaster Mine in 1966. The Hollinger Mine was nearing the end of its life and finally closed in 1968.

Against this background, the announcement of the Kidd Creek discovery, made just 24 km north of Timmins, in early 1964, was even more spectacular. This discovery, by an exploration team of Texas Gulf Sulphur Company, represented the culmination of a more than 10 year-long, progressive exploration effort (Bleeker and Hester, 1999). The discovery drill hole K55-1, started on November 8, 1963, cut an initial 26 feet of clay overburden followed by 629 feet of mineralized rhyolite and massive sulphides, averaging 1.18% Cu, 8.1% Zn, and 3.8 ounces/ton Ag. The discovery was kept secret for several months, during which period the mineral rights to the adjoining properties were secured. To detract attention from the discovery site, a decoy hole was drilled on barren ground approximately 1 km to the north. Despite the secrecy, rumours of a significant sulphide discovery started to surface in Timmins in early 1964 and when the discovery was finally announced on April 16, 1964, one of the largest staking rushes in Canadian mining history was in full swing.

Presently, Kidd Creek has mined well over a 100 million tonnes of sulphide ore and is now developing a deep down-plunge extension of the tightly folded ore body (Bleeker, 1999). It is quickly becoming one of the deepest base metal mines in the world. The giant Hollinger-McIntire mine has long since closed and barring significant new discoveries, the Dome Mine is nearing the end of its life. However, several smaller gold mines are still in operation in and around Timmins, and with rising metal prices in 2004, exploration for new mineral discoveries is in full swing.

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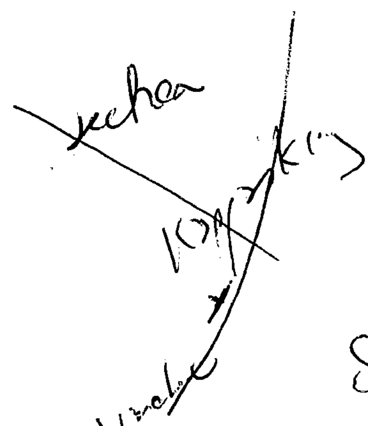
young conglomerate
2680
pulled horizontal



Small green step

Tisdale - refine scale
2700 my

2 models
- 1) Y/int
or 2) Strike slip

some stacking of packages
but ~~not~~ stratigraphic

- last igneous Kirst
then turbidites
finally overlain by Timiskaming group (conglomerates)
<2669 Ma

to get gold to form
need faults
large faults
- focus fluids
+ deposit
most deposits
with meteoric
fluids

set in synclines
within Destor-Porcupine Fault
- have been transported to mid crust
& back

Au sits
in volcanic
rocks + conglomerates
major phase of gold came
after conglomerates
deposited

some timing as high grade
metam in mid & lower crust
(amphibolite-granulite)
CO_2-H_2O

Chicxulub 65 my Mexico slightly older than Sudbury

Africa – Vredefort eroded very deep

Sudbury 1850 my eroded to 8 km depth

Part 3.
Sudbury area

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Introduction

Figure 1 illustrates the regional geological setting of the Sudbury area. It lies at the juncture of three structural province of the Canadian Shield; namely the Archean Superior Province, the Paleoproterozoic Southern Province and the Mesoproterozoic Grenville Province, and includes the 1850 Ma Sudbury Structure. It should be noted that rocks of the Southern Province are underlain at depth by rocks of the Superior Province, that that the Grenville Province formed well after the Sudbury Structure, consequently, the juxtaposition of these 4 major geological domains is not unusual or unique.

Table 1 provides a summary of the major geological events and the age of the main rock units within the Sudbury area. Recent summary articles on the regional geology include: Southern Province: Bennett *et al.* (1991). Sudbury Structure: Dressler *et al.* (1991). Grenville Province: Easton (1992), Davidson (1998b). Superior Province in Sudbury area: Easton (2000).

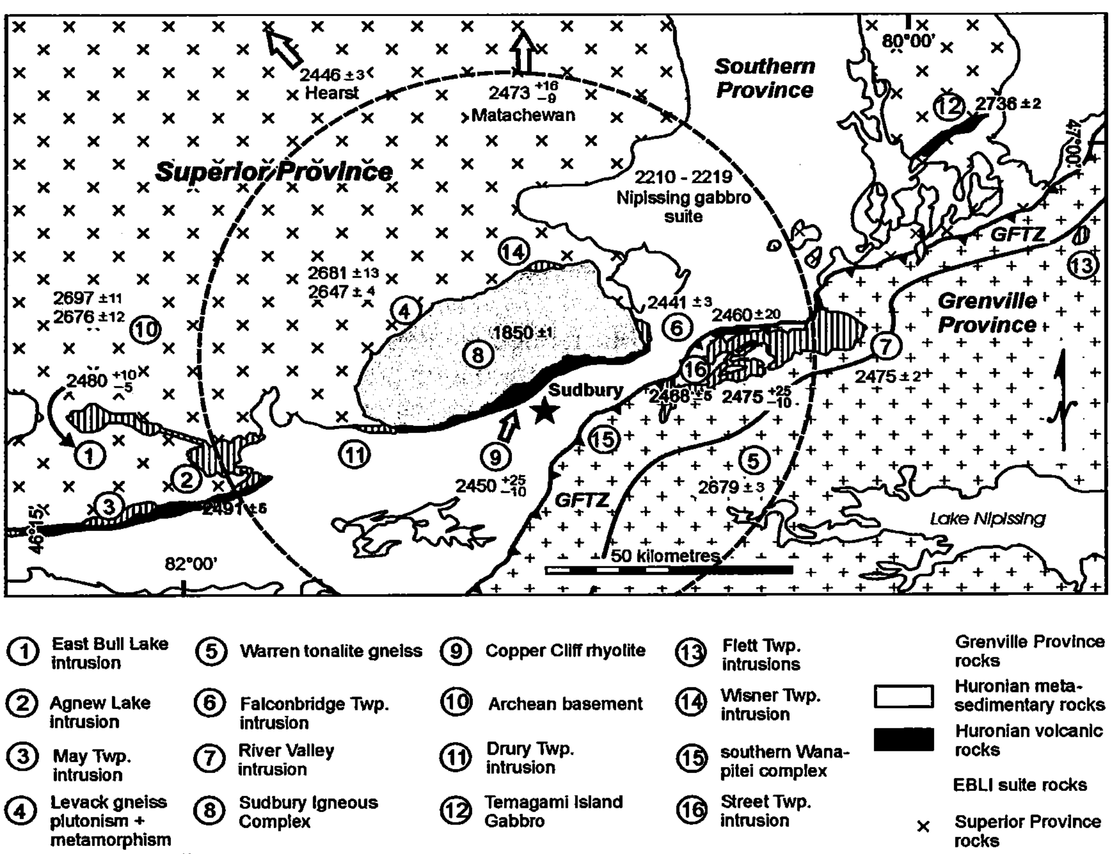

Figure 1. Map of the Sudbury region.

Proterozoic rocks in the Sudbury area

Proterozoic rocks in the Sudbury area are assigned to either the Paleoproterozoic Southern Province or the Mesoproterozoic Grenville Province. The Southern Province in Ontario comprises Paleoproterozoic metasedimentary and metavolcanic rocks of the Huronian Supergroup and gabbroic intrusions of the Nipissing gabbro suite. Also included in the Southern Province are the Sudbury Igneous Complex and the Whitewater Group; plutonic and minor volcanic rocks of the Killarney Magmatic Belt; and rocks of the Sudbury diabase dike swarm (Bennett *et al.* 1991).

The Huronian Supergroup was deposited unconformably on Archean plutonic and supracrustal rocks of the Superior Province. The lowest unit, the Elliot Lake Group, consists of both metavolcanic and metasedimentary rocks. In the Sudbury area, the metavolcanic units include tholeiitic basalts of the Elsie Mountain Formation, evolved tholeiitic basalts, dacites, and metasedimentary rocks of the Stobie Formation, and dacites and rhyolites of the ~2450 Ma Copper Cliff Formation. The latter are likely coeval with the Murray and Creighton granites (Bennett *et al.*, 1991). The metavolcanic units interfinger with, and are overlain by the Matinenda Formation in the west and the McKim Formation in the east. Geochemical data reported by Innes (1972, 1977) and Easton (1998) indicate a tholeiitic affinity for the Stobie Formation, whereas felsic metavolcanic rocks of the Copper Cliff Formation and the Murray and Creighton granites show calc-alkalic signatures.

At the base of the Huronian Supergroup in the Elliot Lake, Agnew Lake and Sudbury areas are several layered gabbro to anorthosite intrusions referred to as the East Bull Lake intrusive suite (Peck *et al.*, 1993; James *et al.* 2002; Easton 2003; Easton *et al.* 2004). These bodies have been dated at 2491 to 2475 Ma (Krogh *et al.*, 1984), and appear to be slightly older than the rocks of the Elliot Lake Group.

Current geochronological data suggests 2 main pulses in magmatism:

(1) A predominantly mafic pulse at ~2475 Ma, represented by the East Bull Lake intrusive suite, Matachewan dikes, and the mafic part of the Elliot Lake Group.

(2) A felsic-dominated pulse at ~2450 Ma, represented by felsic metavolcanic rocks of the Elliot Lake Group, related granitic intrusions, and Hearst dikes.

The rifting event produced an east-trending, southward-deepening basin that likely resulted in formation of oceanic crust. The geometry of the Matachewan and Hearst dikes and their distribution with respect to the metavolcanic rocks of the Elliot Lake Group is consistent with the north-northwest trend of the dikes reflecting the orientation of a failed arm of the rift (e.g., Fahrig, 1987; Ernst and Buchan 2001).

The collisional event at 1870 to 1835 Ma, termed the Penokean Orogeny, is believed to be responsible for most of the metamorphism and deformation present within the Huronian Supergroup. The scale and intensity of the Penokean Orogeny remain the subject of debate (Davidson *et al.*, 1992; Card, 1992), in part because the Penokean Orogeny has no associated plutonism in Ontario. In contrast, Riller *et al.* (1999) attributed deformation and peak metamorphism of the Huronian Supergroup to the 2470 to 2220 million-year-old Blezardian Orogeny, with subsequent transpressional deformation during the Penokean Orogeny. These divergent views reflect the lack of constraints on the age of Huronian metamorphism and deformation.

The Sudbury Igneous Complex was emplaced at 1850 Ma (Krogh *et al.*, 1984) and consists of a lower, ore-bearing sublayer, a main mass of norite, and an upper granophyre (e.g., Dressler *et al.*, 1991). Associated with the Sudbury Igneous Complex are brecciated rocks, termed the Sudbury breccias (e.g., Dressler *et al.*, 1991), consisting of randomly oriented blocks of country rock in a fine-grained, pseudotachylite matrix. The breccias occur up to 200 km from Sudbury, but are most abundant near Sudbury. The Sudbury Igneous Complex and related rocks has been variously interpreted as originating from meteorite

Table 1. Timing of major geological events and summary of age constraints on the main rock units present in the Sudbury area (*from* Easton 2003).

Event and/or Map Unit	Age Constraint (in Ma)	Comment and/or Source
Grenville dike swarm	586±4	Kamo, Krogh and Kumarapeli (1995)
Pegmatite vein emplacement	989±2	Corfu and Easton (2000)
Age of peak metamorphism in the hangingwall of the Grenville Front tectonc zone	1000 to 990	Corfu and Easton (2000), Easton and Kamo (2003)
Age of peak Grenvillian metamorphism in the Central Gneiss Belt	1040 to 1030	Carr *et al.* (2000)
Sudbury dike swarm	1238±4	Emplaced in or along northwest-trending faults in the Southern Province. Deformed and metamorphosed within the Grenville Province. Krogh *et al.* (1987).
Killarney magmatic belt second-stage magmatism, coincident with magmitism in the Eastern Granite Rhyolite Province and in the Central Gneiss Belt	1471±3	van Breemen and Davidson (1988)
Regional albitization metasomatic event	1701±4	U/Pb monazite, Schandl, Gorton and Davis (1994); fluid focussed along northwest-trending faults
Killarney magmatic belt volcanism and high-level plutonism	1740, 1747±3, 1749±12	van Breemen and Davidson (1988); Sullivan and Davidson (1993); Davidson and van Breemen (1994)
Northwest-trending regional faults	Pre-1700, post-1850	Faults cut Sudbury Structure
Penokean orogeny (folding and metamorphism of Huronian Supergroup rocks)	~1835	Peak metamorphism. Holm *et al.* (2001)
Impact event and formation of Sudbury breccia	1850±1	Krogh, Davis and Corfu (1984)
Thrust faulting	post-F_2 pre-regional faulting	Sudbury breccia localized along these faults, suggesting they are pre-Sudbury Structure
F_2 folding	post-2200, pre-1700, pre 1850?	Pre-regional faulting, Nipissing sills folded, relationship to Sudbury Structure uncertain
F_1 folding	pre-2200	Nipissing sills in Street Township appear unaffected by this folding
Emplacement of Nipissing gabbro sills	2219±4 to 2210±4	Corfu and Andrews (1986); Noble and Lightfoot (1992)
Huronian sedimentation	>2220 but <2460	
Huronian felsic volcanism and related plutonic rocks	~2477 to 2375 (2450±25, 2460±20, 2477±9, 2415±5, 2376±2)	Krogh, Davis and Corfu (1984), Corfu and Easton (2000), Krogh, Kamo and Bohor (1996), Smith (2002)
Emplacement of East Bull Lake intrusive suite rocks	2475±2	Heaman (geochronologist, University of Alberta, personal communication, 1999)
Emplacement of orthopyroxene hornblendite bodies	2468±5	Corfu and Easton (2000)
Emplacement of alkali feldspar granite bodies in Crerar and Dana townships	~2660	Bodies intrude Crerar and Pardo gneiss, Easton and Kamo (2003)
High-grade Archean metamorphism and migmatization	2647±4	Krogh, Davis and Corfu (1984); Wodicka and Card (1995)
Emplacement of megacrystic granodiorite bodies in Crerar gneiss association	~2665	Easton and Kamo (2003)
Emplacement ages of Archean units in the Sudbury area	2711±7 to 2642±1	Krogh, Davis and Corfu (1984); Wodicka and Card (1995); Chen, Krogh and Lumbers (1995); Meldrum *et al.* (1997)

impact, impact-induced plutonism and volcanism, and volcanism (*see* Muir, 1984, Peredery and Morrison, 1984). The southern part of the Sudbury Igneous Complex was weakly metamorphosed by an event that also retrograded metamorphosed rocks of the Huronian Supergroup. Regional sodium and potassium metasomatism and silicification has intensely altered rocks locally within the Huronian Supergroup, especially along faults, at ~1700 Ma (Gates, 1991; Schandl *et al.*, 1994; Easton and Murphy 2002; Fedo *et al.*, 1997). Significant magmatism occurred at 1750 to 1730 Ma and again at 1500 to 1450 Ma in the Killarney Magmatic Belt (van Breemen and Davidson, 1988; Davidson and van Breemen, 1994).

The last major magmatic activity in the Southern Province occurred at 1240 Ma with the emplacement of the northwest-trending Sudbury diabase dike swarm (Krogh *et al.*, 1987). This event is noteworthy, as rocks of this dike swarm can be traced across the Grenville Front into the Grenville Front tectonic zone, providing an important marker horizon (e.g., Bethune, 1997).

Significant changes in thickness within the Huronian Supergroup occur east and west of a line roughly coincident with the trace of the north-trending Upper Wanapitei River fault. Debicki (1990) estimated the thickness the Huronian Supergroup in the Sudbury-Coniston area to be roughly 10,350 m, 85% of which consists of the lower 3 groups. In contrast, east and northeast of Wanapitei Lake, the thickness of the Huronian Supergroup is roughly 6,250 m, 75% of which consists of the Cobalt Group. The Elliot Lake, Hough Lake and Quirke Lake groups are all considerably thinner east of Wanapitei Lake, and the McKim and Ramsey Lake formations are apparently absent (Easton and Murphy 2002).

Major Fault Systems

Upper Wanapitei River fault

The Upper Wanapitei River fault has had a protracted deformation history, exhibiting at least 7 to 8 km of left-lateral movement between 2170 and 1850 Ma (Buchan and Ernst, 1994), and at least 3 km of left-lateral movement post-1040 Ma (Easton and Murphy 2002). According to Easton (2000), the north-trending Upper Wanapitei River fault apparently divides the Archean rocks in the Elliot Lake to North Bay area into two domains, with the boundary between these domains passing through Street Township. The eastern domain, which includes the River Valley–Hagar area consists of supracrustal and metaplutonic rocks, with deeper levels in the crust being exposed, likely due to Grenville orogenesis, to the south (Easton 2003). In contrast, the western domain is pluton-dominated, with deeper levels of the crust, being exposed to the east. The amount of vertical movement across the fault is unknown. Significant vertical movement would be necessary to juxtapose granulites of the Levack gneiss complex against higher-crustal-level rocks to the east, however, if the Upper Wanapitei River fault re-activated an older listric fault system, then considerably less relative uplift across the fault may be present.

Murray fault system

Significant changes in stratigraphic thickness within the Huronian Supergroup occur across some regional faults, in particular, the Murray fault system (e.g., Bennett et al, 1991, and references therein). North of the Murray fault, the McKim Formation is tens of metres thick, whereas south of the fault, it probably exceeds 1000 m. The thickness and facies variations across the Murray fault system suggest that the faults represent south side down, syn-sedimentary, growth faults that were reactivated during compression related to the Penokean Orogeny (e.g. Zolnai *et al.*, 1984).

In addition to stratigraphic thickness variations, the Murray fault system also marks profound changes in structural style, metamorphic grade and magmatic associations (Card *et* *al.*, 1972). Deformation is more complicated and of greater intensity south of the fault. Likewise, metamorphic grade is higher immediately south of the fault (amphibolite facies transitional southward to greenschist facies) than to the north (greenschist to subgreenschist facies). South of the fault, there are several 1750 Ma and younger granitoid complexes (e.g., Cutler batholith) (Davidson and van Breemen, 1994), whereas, north of the fault, there are no such intrusions.

Northeast of Coniston, the Grenville Front boundary fault and the Murray fault system are thought to merge into the Wanapitei fault (Davidson, 1998a), which can be traced eastward into Street Township. This fault is then offset to the north by the Upper Wanapitei River fault, and continues eastward as the Ess Creek and Grenville Front boundary faults along the trend of the Kabikotitwia and Strugeon rivers (Easton and Murphy 2002).

The Grenville Front

The Grenville Front tectonic zone is a region up to 30 km across lying between the Grenville Front and the Central Gneiss Belt of the Grenville Province (*see* Figure 1). Easton (1992) divided the Grenville Front tectonic zone in Ontario into 3 lithologic segments. The western segment between Killarney and Wahnapitae comprises rocks equivalent in age, geophysical signature, and rock type to the adjacent Killarney Magmatic Belt. The central segment stretches from Wahnapitae to River Valley, and contains mafic intrusive rocks of both the East Bull Lake and the Nipissing intrusive suites, as well as several types of migmatitic gneisses, likely of Neoarchean age (Easton 2000, 2003). The eastern segment, between River Valley and the Ottawa River, includes rocks mainly derived from the adjacent Superior Province. The study area lies mainly in the central segment, although the eastern part of the study area straddles the boundary between the central and eastern segments.

The Grenville Front is a zone of southeast-dipping faults and mylonite and has generally been placed at the southeast limit of recognizable Southern Province rocks (e.g., Lumbers, 1975; Davidson, 1998a). Locally there are complications that have led to many debates concerning the identity of the Grenville Front and its distinction from other faults that intersect, merge with or are parallel to the Front (see discussion in Davidson, 1998a). In central Street Township the Grenville Front (which is coincident with the Wanapitei fault) has been displaced to the north by the younger, north-trending, Upper Wanapitei River fault by at least 850 m of sinistral and west-side-up movement (Easton and Murphy 2002). This displacement likely occurred after circa 590 Ma, as a Grenville swarm diabase dike in northern Henry and Loughrin townships is also displaced by north-trending faults.

Seismic surveys

Sudbury Structure

Two interpretations have been suggested for the lines across the Sudbury Structure. The consensus view, expressed by Milkereit *et al.* (1992), is that the structure is not funnel-shaped, as had long been previously suggested, but consisted of a series of south dipping sheets, with rocks of the Huronian Supergroup and the South Range thrust northward onto less deformed rocks of the north range along faults related to the south range shear zone. In contrast, Card and Jackson (1995) suggest instead that the seismic data can be modelled as a funnel- or basinal-shape, with some deformation occuring along the south range shear zones.

Wu *et al.* (1995), upon additional examination of the data, that thrust faults of the south range shear zone affect he Onwatin Formation, but not the overlying Chelmsford Formation, placing a relatve age constraint on the timing of thrusting. Boerner and Milkereit (1999) note that there is a strong reflection contrast between the contact of the Sudbury Igneous Complex and its footwall, and present evidence for distinguishing impact from non-impact related reflectors within the reflection data. Boerner *et al.* (2000) provide a synthesis of the seismic studies of the Sudbury Structure.

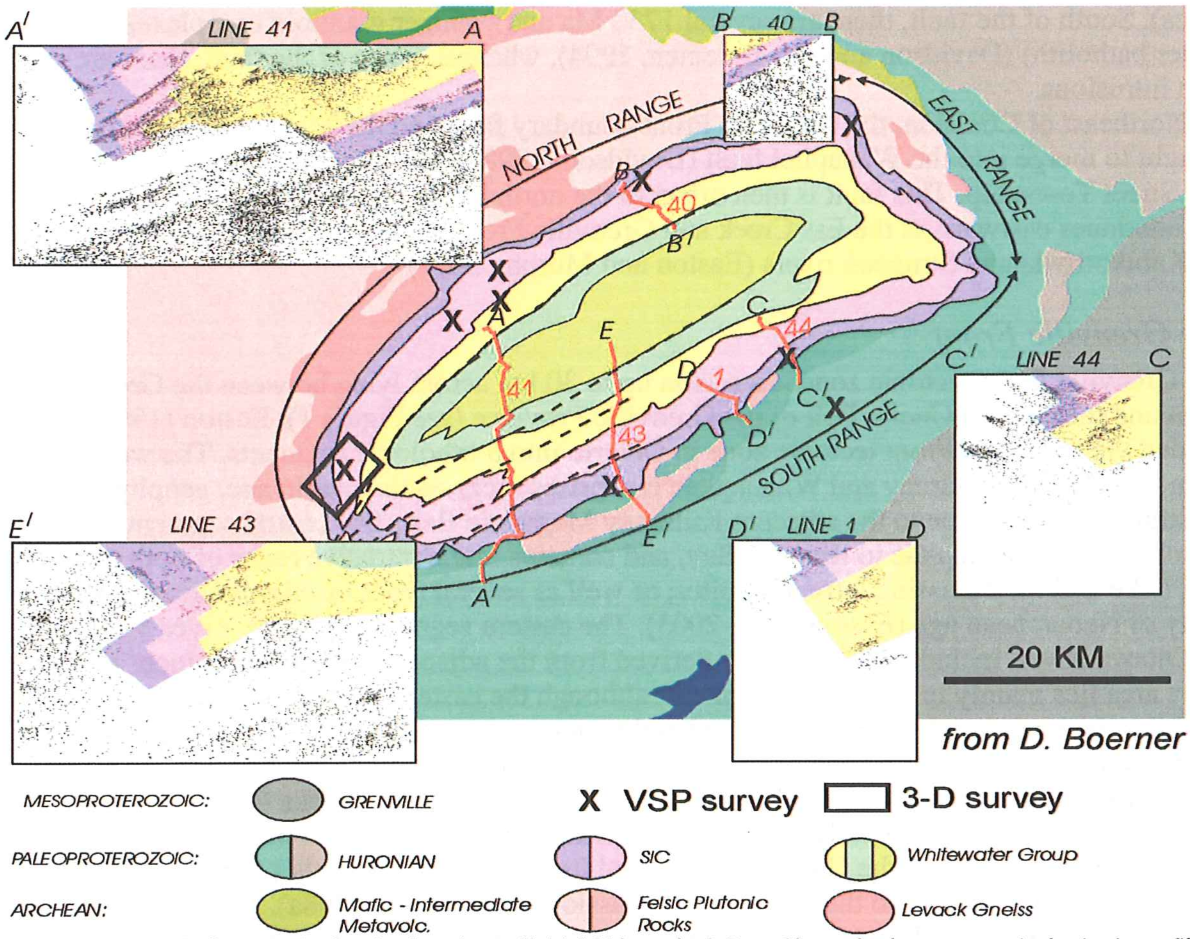

Figure 3. Map of Sudbury Basin showing locations of initial Lithoprobe 2-D profiles and subsequent vertical seismic profiles (VSP) and the Trill 3-D survey.

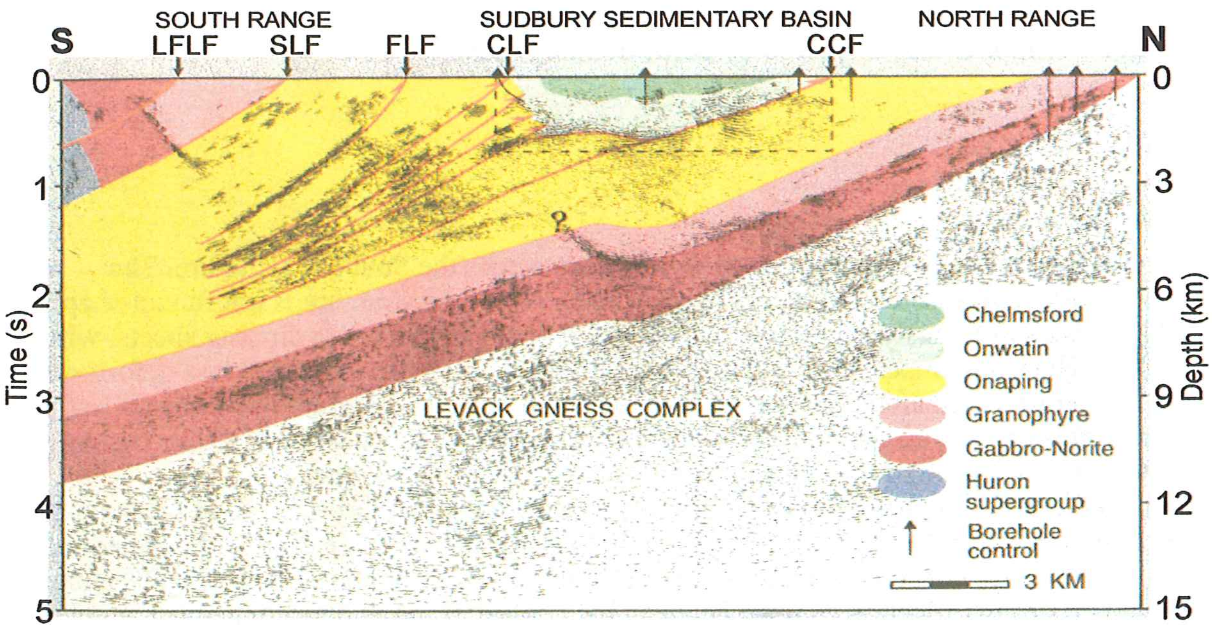

Figure 2. Composite N-S cross section of the Sudbury basin based on the Lithoprobe Lines 40 & 41.

biggest Ni area
DS – impact melt
– Ni assoc to base of melt

Cobalt Plate, Southern Province

Commonly overlooked is the existence of a seismic reflection survey over the large aeromagnetic anomaly that underlies the Cobalt Plate east-northeast of Lake Wanapitei (Milkereit and Wu, 1996). Interpretation of the data suggests the presence of a graben beneath the metasedimentary rocks of the Cobalt Group exposed on surface, a graben interpreted to be filled with metavolcanic and metasedimentary rocks equivalent to the Elliot Lake Group of the Huronian Supergroup. The orientation of this graben has implications with respect to modelling the orientation of rifting along edge of the Superior Province just prior to Hurnian Group sedmentation.

Grenville Front

In Ontario, only 2 seismic lines image the full width of the Grenville Front and the Grenville Front tectonic zone, these are the Glimpce line in Lake Huron (Green *et al.*, 1988) and Lithoprobe Line 15 (Kellett *et al.*, 1994), which crosses the Grenville Front east of the Ottawa River in the Temiskaming area. Lithoprobe Lines 41 and 43 across the Sudbury Structure do not extent far enough south to image the Grenville Front, however, the southern end of these lines likely exhibit some effects related to Grenville deformation. Unpublished oil industry seismic reflection lines in Essex County, east of Windsor, also partially images the hanging wall of the Grenville Front, again displaying prominent east-southeast dipping reflectors (Carter *et al.*, 1996).

The two profiles are very different in character, with the Glimpce line imaging a wide, prominent zone of southeast-dipping reflectors, an image so striking that it was featured on the cover of *Geology*. The preliminary interpretation of this data by Green *et al.* (1988) suggested that the southeast-dipping reflectors extended through the entire crust to the Moho. White *et al.* (2000) and Carr *et al.* (2000), re-examined the data, and now suggest that the reflectors bottom out in a decollement about 20 km depth at the Grenville Front, which reaches depths of about 25 km further south in the Parry Sound area. In contrast, on Line 15, the Grenville Front is not clearly visible. The difference between the Glimpce Line and Line 15 probably reflects the age of the crust. On Line 15, Archean rocks form the crust on either side of the Grenville Front, and apart from some metamorphic reworking, are not significantly different in physical properties north and south of the Front. In contrast, the Glimpce line images the Grenville Front in an area where there has been protracted activity along the southern edge of the Superior Province, and the strong reflectors imaged long this line may represent a composite of Proterozoic deformational activity, not simply the Grenvillian event. Detailed interpretation of the Glimpce line is also hampered by the fact that it is located well into Lake Huron, far from any surface exposure of rocks of the orogen.

Parry Sound Domain, Grenville Province

Lithoprobe lines 30 and 31 (White *et al.*, 1994, 2000) in the Parry Sound area were designed to test geophysical models of the Parry Sound domain which suggested that it was a relatively flat-lying sheet, roughly 6 km thick at its maximum, of denser, mafic rocks that likely had been thrust onto more quartzofeldspathic gneisses of the Central Gneiss Belt. The seismic images were consistent with this interpretation as well as gravity models of the structure (Lindia *et al.*, 1983). On the northern end of the domain, the Parry Sound shear zone is imaged as a series of prominent, southeast-dipping reflectors, consistent with the lithological heterogenity observed within the shear zone, and the moderate dips observed within the zone.

Day 2:
The Sudbury Structure and related rocks

Drive from Timmins along Highway 144 toward Sudbury.

Stop for lunch at the A.Y. Jackson Lookout at Onaping Falls, which is located in the northern part of the Sudbury Structure. The lookout and the falls are underlain by rocks of the Onaping Formation, which has been interpreted to be a fall back breccia related to the meteorite impact which is thought to have formed the Sudbury Structure. A display area near the parking lot contains large blocks of most major rock types present in the Sudbury Structure, and is part of a local effort to increase public awareness of geoscience. Following lunch, we will examine several outcrops that illustrate key features of the Sudbury structure, which is believed by many workers to result from a meteorite impact event that occurred at 1850±1 Ma. In the interest of time, not all of the stops described below will be visited. Stop descriptions have been adapted from guidebooks by Young (1991), Dressler *et al.* (1992) and Cosec (1999).

Exit lookout, turn north on Highway 144 and proceed northwest for about 11.5 km. Park in the pullout provided and examine rock exposures.

Caution: Highway 144 is a busy road with narrow shoulders. Stay on exposures away from the road using existing paths.

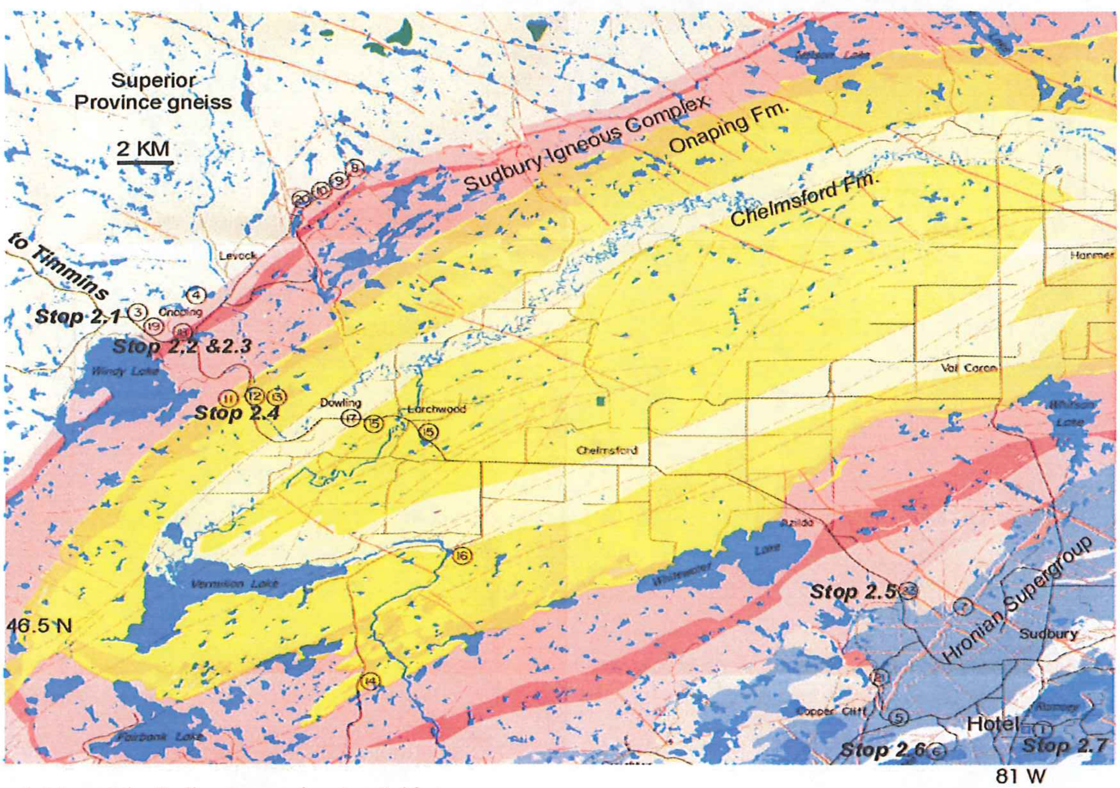

Figure 4. Map of the Sudbury area showing field stops.

Chixalub - see ... domes ... shatter event retrograde to greenschist

Stop 2.1:
Sudbury breccia in Archean basement

Archean rocks exposed immediately on the north side of the Sudbury Structure consist of migmatitic gneisses with mafic layers that constitute the Levack gneiss. The Levack gneiss consists of rocks ranging in age from 2711 to 2660 Ma, all affected by granulite facies metamorphism dated at 2647±4 Ma (Krogh *et al.* 1984; Wodicka and Card 1995). To the north, the Levack gneiss is intruded by monzogranite of the Cartier granite, which has been dated at 2642±1 Ma (Meldrum *et al.* 1997).

mapped out to 100 km from Sudbury deposits ... lost

At this locality, the gneisses are cut by a series of veins of Sudbury Breccia. Sudbury breccias are pesudotachylites believed to have formed during fracturing, comminution and melting related to the explosive event that produced the Sudbury structure at 1850±1 Ma (Dressler 1984). Most breccias consist of an aphanitic, siliceous, medium to dark grey matrix, which may be massive, flow banded or vesicular. Fragments are commonly rounded, range in size from less than one centimetre up to several metres, and typically are dominated by fragments of the host rock. Exotic fragments do exist, however, especially in the larger breccia bodies. Major breccia zones commonly follow formational contacts, or may be associated with major faults that developed prior to, or during the Sudbury event.

Return to vehicle, turn around and head southeast along Highway 144. At roughly 3.8 km, the road cuts through the first outcrops of the North Range of the Sudbury Igneous Complex. Turn right onto Elks Club Road, turn around and park. Examine outcrops on the north side of the road between the Highway and the railroad.

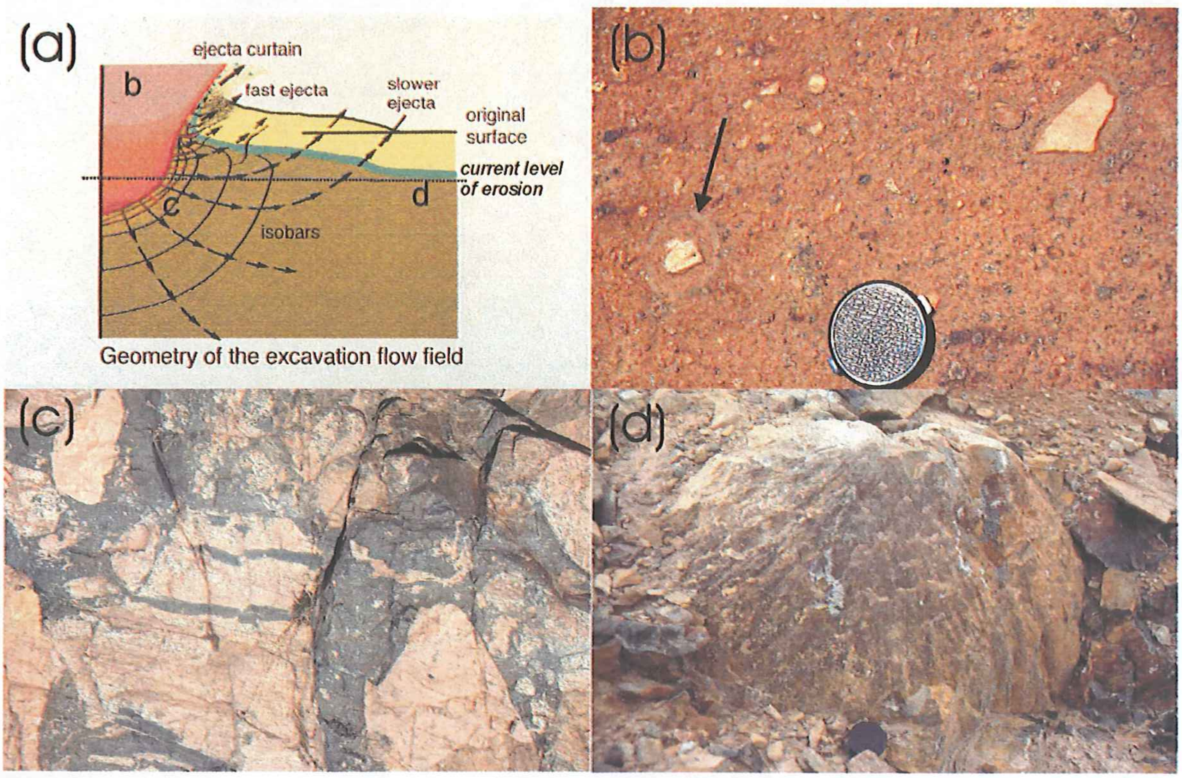

Figure 4. Composite showing three key outcrops and where the proto-lith was formed during the Sudbury impact event. (a) Geometry of the excavation flow field (modified from Melosh, 1989, Impact cratering: a geological process; Oxford Monographs on Geology & Geophysics 11); (b) Accretion Lapilli in the Onaping Formation; (c) Sudbury Breccia as exposed on the North Range-black rock is "pseudotachyite", a micro-crystalline glass; (d) shatter cone.

Stop 2.2:
Felsic norite

The Sudbury Igneous Complex is composed of the Sublayer (contact sublayer and offset dikes) and the Main Mass. The Main Mass is further subdivided into 4 units, from base to top, these are the mafic norite, felsic norite, quartz gabbro and granophyre units. The felsic norite is a medium-grained to coarse-grained rock with hypidiomorphic equigranular texture. It is composed of 50 to 70% plagioclase (An_{50-65}), 3 to 10% potassium feldspar, 10 to 30% orthopyroxene, 5 to 15% granophyric quartz and 2 to 5% biotite, with mnor amounts of aoatite, zircon, ilmenite and sulphide minerals (Coats and Snajdr 1984).

Return to vehicle, continue southeast along Highway 144 for roughly 1.4 km to the junction of Regional Road 8. Turn left and park on the east side of the road. First, walk to the easternmost end of the high rock cut on the north side of Regional Road 8 to examine the quartz gabbro unit. Then cross the road to examine rocks of the granophyre.

Caution: Regional Road 8 is a busy road with narrow shoulders. Please be careful crossing the road.

Stop 2.3:
Quartz gabbro and granophyre

The quartz gabbro of the Sudbury Igneous Complex, also called the Transition Zone, is an oxide-rich gabbro containing up to 8% oxides. Major minerals are plagioclase, clinopyroxene, amphibole, quartz, and a micrographic intergrowth of quartz and potassium feldspar. Apatite and pyrite are also present. The unit is roughly 225 m thick in the North Range.

The granophyre is approximately 900 m thick and is pink. It consists of roughly three parts micrographic intergrowth of quartz and potassium feldspar to one part plagioclase. Quartz, biotite, amphibole and opaque minerals are also present.

Return to vehicle, return to Highway 144, and continue southeast for 4.7 km and park in a large gravelled area on the right.

Caution: Please be aware of traffic!

Stop 2.4:
Onaping Formation

The Onaping Formation is one of the most controversial units within the Sudbury Structure. Proponents of the meteorite impact theory regard it as the fall back breccia (e.g. Peredery and Morrison 1984), whereas those who favour an endogenic origin regard it as a complex unit of ash flow tuffs (aqueous and subareial) and related volcanic rocks (e.g. Muir 1984). It is intruded by granophyre of the Sudbury Igneous Complex at its base, and the top of the Onaping Formation grades into the overlying Onwatin Formation slates.

The lowermost, Basal Member is discontinuous and comprises a variety of monolithic and heterolithic breccias consisting of material derived from the Archean basement and the Huronian Supergroup, and rages from 150 to 300 m thick. The breccias may be either matrix- or clast-supported, with fragments varyng in size and shape, some of which are up to 100 m in diameter.

The Grey Member, which we see here, is a major breccia unit ranging from 200 to 700 m thick that is stratified on a gross scale. The contact with the underlying Basal Member or granophyre is sharp, whereas the contact with the Black Member can be sharp or gradational.

Fragments are composed of country rocks, microbreccia fragments, divitrified glass fragments, crustals, crystal fragments, sulphides and sulphide fragments. Archean and Huronian rocks make up to 30% of the rock fragments. Shock metamorphic features are observed in many of the country rck fragments, and zircons from the Onaping Formation suffer shock-related lead loss at circa 1850 Ma (Krogh *et al.* 1996). The matrix of the Onaping Formation varies compositionally from dacitic to andesitic in the Basal Member to tholeiitic basalt in the Black Member.

The Black Member is marked by an abrupt increase in carbon content, however, Muir (1984) suggested that a zone of chloritized glass shards was a more reliable indicator of the Grey Member-Black Member contact.

Melt bodies constitute a fourth unit within the Onaping Formation. They are inclusion-bearing igneous textured rocks with chilled margins, and occur in clusters, mainly near the Basal Member-Grey Member contact. Country rock inclusions can amount to 80% of the rock by volume, and are generally well sorted; ranging from rounded to angular, and from 1 cm to 1 m in diameter.

Return to vehicle. Continue southeast of Highway 144, passing through the town of Dowling.

As we pass through Dowling, we drop off of the more rugged topography of the Onaping Formation into a plain developed on the metasedimentary rocks of the Onwatin and Chelmsford Formations. At 5.5 km from the last stop, there are some low-lying outcrops on the south side of the road. These outcrops expose the contact between the Onwatin and Chelmsford formations. The Onwatin Formation consists of cleaved, black, finely laminated, carbonaceous mudstone, and regionally is 600 m thick, along at this locality, it is much thinner. The Chelmsford Formation consists of dark grey, graded sandstone beds almost 2 m thick, exhibiting coarse Bouma cycles. Carbonate concretions are common in the sandstone beds. Paleocurrent data indicate transport to the southwest, along the length of the Sudbury Basin. The Chelmsford Formation is roughly 850 m thick at its present erosional level.

Continue southeast on Highway 144 for roughly 20 km to the Sudbury Discovery Site, which is marked by a blue historical plaque. Park in the paved area on the east side of the Highway. INCO's Clarabelle No. 2 Open Pit Mine is visible to the east.

Stop 2.5:
Discovery Site

Nickel was first discovered in the Sudbury area in 1856 by A. Murray of the Geological Survey of Canada, 200 m west of what is now the Creighton Mine (Giblin 1984). It was not until 1883, however, during construction of the Canadian Pacific Railway that a nickel boom started. A railway rock cut exposed copper and nickel mineralization that was subsequently acquired by T. Murray (unrelated) and became the Murray Mine. The discovery gossan formed over high-grade ore composed of rounded inclusions of exotic mafic and untramafic rocks in a matrix (sublayer) of nearly massive pyrrhotite, chalcopyrite and pentlandite. Within a few years afterward, many deposits had been located around the margin of the Sudbury Igneous Complex. The Sudbury ores were first mined for copper, with the Canadian Copper Company producing the first ore from the Copper Cliff Mine in 1866 (Giblin 1984). It was not until the first load of ore was sent for smelting, that furnace assays revealed the presence of nickel. Since then, over one billion tonnes of ore have been extracted from the Sudbury camp.

The actual discovery outcrop was located approximately in the middle of the open pit on the west side of the road. In mid-1970, the road and railway were relocated to permit mining in the Clarabelle No. 2 open pit. At this site, weakly mineralized Sublayer is exposed on the west side of the tracks. The Sublayer dips north beneath the norite at roughly 60°.

Quartz-rich norite is exposed on the east side of the tracks. The norite contrasts sharply with the Sublayer, being homogeneous, free of inclusions, and has only a minor sulphide content.

Return to vehicle. Drive 2 km south to Godfrey Drive. Turn right and continue south through the town of Copper Cliff to the junction of Regional Road 55 and Balsam Street. Drive south on Regional Road 55 to Kelly Lake Road. Turn right. Follow Kelly Lake Road to Southview Drive. Turn right on Southview Drive and continue for 1.1 km and park.

Stop 2.6:
Sudbury Breccia in rocks of the Huronian Supergroup

At this locality, the road roughly follows the strike of the Huronian rocks. On the south side of the road, polymictic matrix-supported conglomerate of the Ramsey Lake Formation and wackes and feldspathic quartz arenites of the Mississagi Formation. Within the Mississagi Formation is a well developed zone of Sudbury Breccia. Here, well-rounded fragments of quartz arenite occur in a very fine-grained, black weathering matrix.

Turnaround, and follow Southview Drive east to Bouchard Street at the intersection with Regent Street (traffic lights). Turn left onto Regent Street, then turn right onto Walford Road at the next lights. At Walford Road and Paris Street (lights), turn left. Follow Paris Street to Ramsey Lake Road (second set of lights), and turn right onto Ramsey Lake Road. Science North is located to the north, the Travelway Inn and the Sudbury Regional Hospital (Laurentian site) to the south. Continue on Ramsey Lake Road for 1.4 km, and park in pullout by electrical transformer. The Willet Green Miller Centre (Ontario Geological Survey headquarters) and the campus of Laurentian University are located immediately to the south. Walk back along the south side of the road to examine high road cuts.

Caution: Traffic can be heavy on Ramsey Lake Road, especially on week days. Exercise extreme care while on the side of the road.

2.6.A > 3.1

Stop 2.7:
Shatter Cones (Please, No Hammering)

The outcrops of Mississagi Formation quartz arenite on the south side of Ramsey Lake Road are best viewed in late afternoon. Cones are approximately 50 cm long. Shatter cones form by the passage of shock waves through rocks. The Sudbury Structure is surrounded by a zone, over 16 km wide, or rocks containing shatter cones. Shatter cones are most commonly observed in quartz arenite of the Mississagi Formation. The apices of the cones should point upward and inward to the Sudbury Basin, however, subsequent tectonism in the southern part of the Sudbury structure and its country rocks means that this is not always the case.
Return to vehicle. End of Sudbury Structure portion of trip.

Day 3.
The Grenville Province

The descriptions provided in this section have been adapted from Davidson *et al.* (2002) and Davidson and Easton (1998).

Grenville Front

Start at the intersection of Ramsey Lake Road and Paris Street. One of 2 routes is possible, depending on the state of road construction.

Route 1. Turn left and proceed south on Paris Street to the junction with Regent Street (3rd set of traffic lights). Turn left and follow Regent Street south to the Highway 17 Bypass. At the 3rd set of traffic lights, turn right onto the bypass and head east toward North Bay. Take the ramp right onto Highway 17 and proceed east on Highway 17 for 5.6 km. Just past the water treatment plant to the north, turn right onto a gravel road that leads to the Coniston Hydro dam. Follow the road to the entrance to the powerhouse (1.6 km) and park. Walk back to the intersection and continue westward across bridge; leave the road just beyond the railway crossing and examine outcrop at the base of the hill to the north.

Route 2. Turn right and proceed north on Paris Street to Brady Street (4th set of traffic lights), turn right and follow the Brady Street extension to the Kingsway and continue eastward for roughly 15 km. Just past the water treatment plant to the north, turn right onto a gravel road that leads to the Coniston Hydro dam. Follow the road to the entrance to the powerhouse (1.6 km) and park. Walk back to the intersection and continue westward across bridge; leave the road just beyond the railway crossing and examine outcrop at the base of the hill to the north.

Stop 3-1:
Mississagi Formation and migmatitic gneiss and mylonite on opposite sides of the Wanapitei fault (Grenville Front) east of Coniston

The Wanapitei fault is the boundary between the Southern and Grenville provinces at this locality, and thus by definition marks the Grenville Front. Its trace to the north-northeast lies in a shallow valley (along which is a power line); at several places along this draw, low-grade ultramlyonite to ultracataclastite is exposed at the base of the ridge on the south side. To the west-southwest, the fault passes just south of the black slagheap of the defunct smelter at Coniston, beyond which it diverges from the Grenville Front and is referred to, in the Southern Province, as the Murray fault.

At this stop, Mississagi Formation quartz arenite on the north side dips steeply northwst, and faces in the same dierction, as indicated by festoon-type crossbedding. An east-northeast-striking cleavage is evident in the darker, silty interbeds; this dips more steeply than bedding, in accord with the fact that these rocks lie on the southeast limb of the Coniston syncline. Metamorphic grade appears to be low; small flakes of green-brown biotite are present, along with muscovite and chlorite, so the grade is likely no higher than middle- to upper greenschist facies.

Return to the road and re-cross the railway; examine first roadcut south of the tracks.

The outcrop lies on the south side of a covered interval, only a few tens of metres wide, which conceals the Wanapitei fault. The outcrop ridge is underlain by migmatitic quartzofeldspathic and mafic gneiss, garnet amphibolite and minor pelitic schist rich in muscovite and containing minor amounts of garnet, kyanite and sillimanite. Sheath folds may be observed on the outcrop surfaces above the road cut. Narrow mylonite zones that diverge southwestward from the

Wanapitei fault contain highly stretched quartz and large potassium feldspar augen whose sense of rotation implies south-side-up displacement of the higher grade rocks. The marked difference in metamorphic grade across the Wanapitei fault attests to considerable upward displacement on the Grenville side.

Retrace route to Highway 17 bypass, and head southwest. At the junction with Regent Street, follow the ramp to the right, then turn left at the lights onto Regent Street/Highway 69. Continue south on Highway 69. The highway crosses the Murray fault near the bottom of the first hill, and then passes outcrops on the right that expose the contact between the Pecors and Mississagi formations. The first large roadcut is thinly bedded Mississagi Formation that faces south and is tectonically attenuated.

Just after the roadcut on the right, park on the right (paved area) by the swampy valley.

Caution: This is very busy highway with fast-moving traffic, with 2 lanes of traffic merging after coming around a blind corner. Exercise extreme caution; keep an eye on traffic and do not step back onto the roadway when taking photographs.

Stop 3.2: (optional)

Grenville Front and deformed Mississagi Formation arenites

Looking to the west from Highway 69 at this stop, mylonite and ultramylonite marking the Grenville Front are well exposed along the south side of the broad swamp. The front passes through the low point in the skyline to the left of the island in the swamp. The island and the high part of the hill behind it are underlain by coarse grained granite of the Chief Lake intrusion, which at a locality 7 km to the southwest yielded a U/Pb zircon age of 1465 Ma (Davidson and van Breemen 1994). The Mississagi Formation is in contact with the granite on the right (north) flank of this hill. The contact dips steeply to the south and is he locus of shearing with south-side-up sense, but the displacement is probably not great, as dikes of granite are present in the Mississagi Formation north of the contact. The ridge to the right of the swamp is Nipissing gabbro, and Mississagi Formation rocks are exposed sporadically along the lower slope.

Walk north on the west side of Highway 69 to examine a small, duplex-like structure in Mississagi Formation quartz arenite exposed in the roadcut beneath Nipissing metagabbro. Metamorphic grade here is greenschist facies.

Davidson *et al.* (2002) provide a more extensive description of rocks south of the Grenville Front on the east side of the Highway, however, these localities are on private property and are not readily accessible.

Return to vehicle and continue south on Highway 69.

At first, the route passes numerous roadcuts in which gneissc layering dips consistently to the southeast at moderate angle, a characteristic feature of the Grenville Front tectonic zone. Outcrops for the first few kilometres expose a variety of metasedimentary gneisses, quartzite and amphibolite, but further south, outcrops of relatively uniform pink and grey granitoid orthogneisses (Archean?) are more common. After the junction with Highway 637 to Killarney (26 km south of Stop 3.2), the route passes into northern Britt domain of the Central Gneiss Belt, marked by a change in strike of gneissosity to the northwest, with varied dip. Within Britt domain, migmatitic gneiss ad sheets of granitoid gneiss are interlayered and form large open folds with shallowly plunging, southeast-trending axes.

Ductile Shear Zones in the Central Gneiss Belt, Grenville Province

Between Dumont Road and the Shawanaga River, roughly 91 km south of Stop 3.2, roadcuts expose highly strained metasedimentary gneisses of the Shawanga shear zone, a major tectonic break within the Central Gneiss Belt, which marks the trace of the Allochthon Boundary Thrust in this part of the Grenville orogen. North of the Allochthon Boundary Thrust, rocks of Britt domain preserve evidence locally for a 1450 Ma granulite facies metamorphic event (Ketchum *et al.* 1994), which has not been recognized anywhere to the south, despite the presence of metaplutonic rocks of similar age. In addition, as detailed by Ketchum and Davidson (2000), remnants of mafic dikes found within gneisses north of the Allochthon Boundary Thrust have the chemical hallmarks of 1240 Ma Sudbury swarm diabase dikes (e.g., high Ba, P, Ti). In contrast, mafic enclaves in the gneisses south of the Allochthon Boundary Thrust have different chemistry and age (~1160 Ma, Davidson and van Breemen 1990), and in their cores, preserve coronitic textures. The Shawanaga shear zone is also the locus of several widely distributed, relatively small tectonic remnants of mafic rock bodies that retain evidence of former ecologite-facies metamorphism (Davidson 1991). This outcrop was described in detail by Davidson *et al.* (2002), but since then has been cut-back considerably in order to accommodate construction of passing lanes.

Roughly 3 km south of this outcrop, a large body of black weathering, coronitic metagabbro is exposed on the west side of the Highway. From here to Nobel, the route passes though Shawanaga domain. Many low roadcuts expose a variety of shallowly inclined, mainly grey orthogneiss, much of it strongly migmatitic. Beyond Nobel, Highway 69 passes through the Parry Sound shear zone, characterized by dark, shallowly southeast-dipping highly strained gneisses with intervals of stretched anorthositic gneiss. South of the bridge over Mill Lake, where the stretch of divided highway begins, the route passes into granulite-facies gneisses of Parry Sound domain.

Gas station and restaurant on the left in Nobel.

The route now passes into granite orthogneiss that has been dated at ~1350 Ma (van Breemen *et al.* 1986). This pluton locally has a sharp upper contact with metasedimentary rocks, including quartzite, pelitic gneiss and minor conglomerate, preserved as a thin lens beneath the Parry Island thrust sheet. The contact may represent either a deformed unconformity or a flattened intrusive contact, preserved where the sole thrust lies at higher structural level. After 600 m, turn right on Bayside Drive (formerly Highway 69), after another 600 m, park on the right shoulder.

Stop 3-3:

Ductile gneiss in the lower part of the Parry Sound shear zone (optional)

This stop illustrates aspects of the Parry Sound shear zone (PSSZ), a major zone of highly strained rocks that separates typically pink and grey migmatitic rocks of the Shawanaga (formerly southern Britt) domain from overlying, dark-hued mafic and felsic granulites of the Parry Sound allochthon. The PSSZ along the west and north sides of the allochthon dips moderately to shallowly southeastward. High-strain rocks can be traced around the north end of Parry Sound domain, where they turn southward and dips steeply westward. Gravity modeling suggests that at its widest part, the Parry Sound allochthon "bottoms out" at a maximum depth of 14 km (Lindia *et al.* 1983). One in particular of two seismic reflections surveys carried out recently across the allochthon (Abitibi-Grenville Lithoprobe lines 30 and 31) confirms this configuration, the southeast-dipping boundary being particularly well imaged (White *et al.* 1994).

At this stop we examine a short section in the lower part of the PSSZ. The base of the PSSZ is not seen at this locality, but to the southwest on the Georgian Bay coast it is marked by straight gneisses which enclose isolated, fist- to house-size pods of meta-anorthosite. Here we are within a microdomain of metasedimentary rocks (Lighthouse association of Culshaw *et al.* 2004) sandwiched between the basal straight gneisses at the structural top of Shawanaga domain and mylonites in the immediate footwall of the Parry Sound granulite allochthon. Thinly layered quartzofeldspathic gneisses low in the section given way up section to biotite gneiss with garnet and kyanite; muscovite in these rocks has formed at the expense of kyanite, which probably involved consumption of pre-existing K-feldspar, no longer present. This thin unit is succeeded by garnet amphibolite (this is a good place to observe the difference between the red colour of garnet associated with hornblende and the violet-pink colour of garnet in the kyanite-bearing rocks; the latter is often a good indicator for the presence of aluminosilicate). Next appear more biotitic gneisses with a distinctly disrupted appearance, characterized by isolated mafic boudins and scattered, rounded, white feldspar augen (gneissic tectonite). Mafic layers in feldspathic gneiss in the upper part of the roadcut are disrupted, with misoriented internal foliation. These rocks are cut by boudined pegmatites, which are more somewhat more steeply inclined than the main layering, indicating rotation toward the shear plane during thrust-sense displacement. Similar pegmatite higher in the PSSZ has been dated at 1160 Ma (van Breemen *et al.* 1986). This assemblage is structurally overlain by a thin layer of disrupted anorthosite laced with pegmatite, which forms a mappable unit that has been traced in both directions along strike for a total in excess of 27 km.

END OF FIELD TRIP.

Return to vehicle. Continue south on Bayside Drive.

Structurally above the deformed anorthosite seen at the last stop is a unit of amphibolitic gneiss with a decidedly podded structure, in which competent lenses lie within an anastomosing shear mesh.

Junction with Highway 124; turn right. Shopping centre to the right; turn left onto former Highway 69.

Podded nature of amphibolite is well displayed in the outcrop face on the left. Beyond the small valley ahead, metasedimentary gneiss that includes thin layers of quartzite is structurally overlain by anorthositic gabbro. In the large roadcut, note the gradual reduction in grain size from coarse-grained anorthositic gabbro with prominent white plagioclase grains to fine-grained layered rock, which is a mylonite. A boudined pegmatite preserved in the middle part of this roadcut has been dated (U-Pb zircon) at 1160 Ma (van Breemen *et al.* 1986).

Junction with Highway 69; turn right. Continue south on Highway 69 and 400 to Toronto. Alternatively, to go to Ottawa, take either Highways 518 or 141 east to Highway 11, then follow Highway 11 to Huntsville, joining with Highway 60, which runs east through Algonquin Provincial Park. Highway 60 links with Highway 17/417 which then continues to Ottawa.

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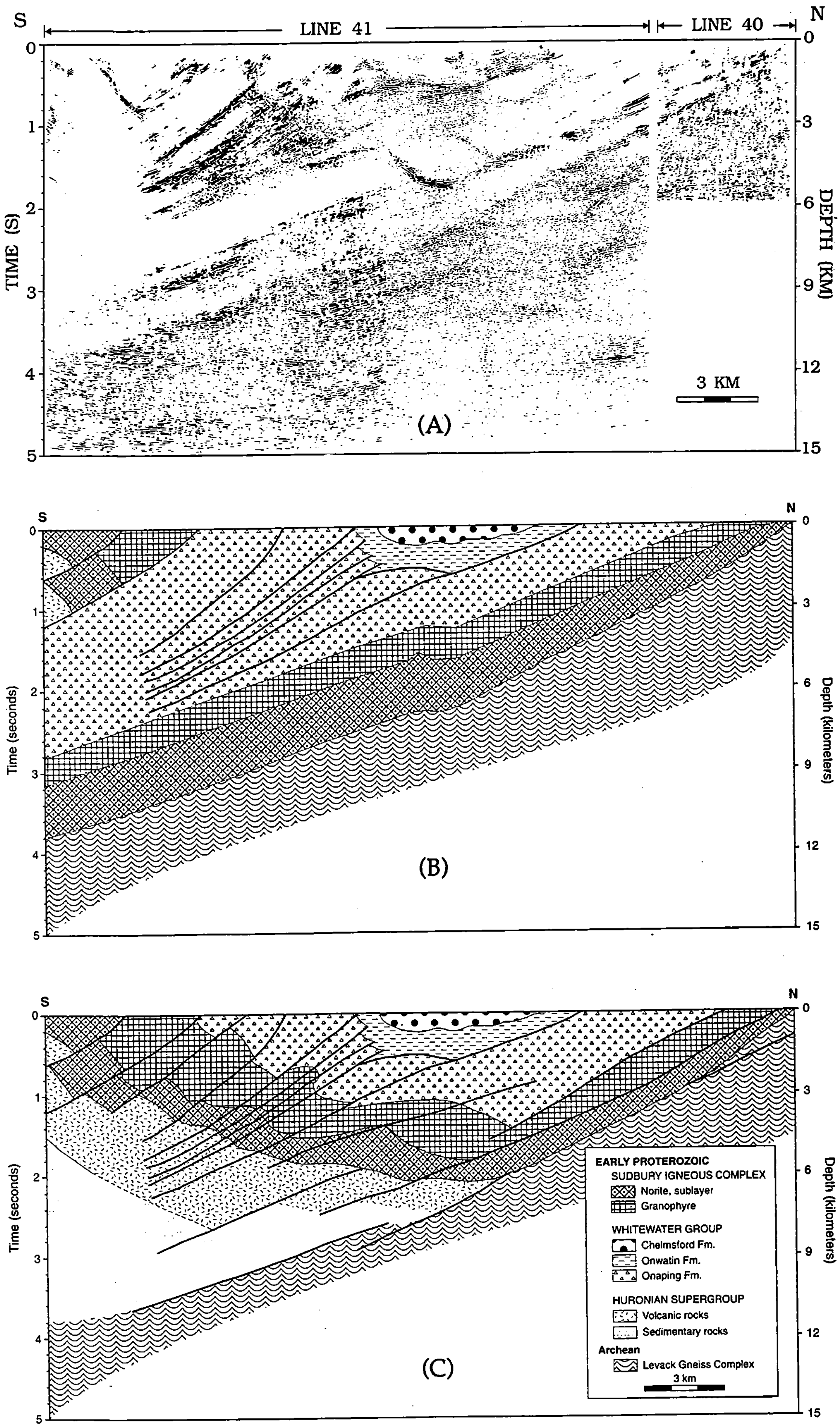

Figure 13: Interpretations of the deep geometry of the Sudbury structure from seismic reflection data (A), by Milkereit et al. (1992) (B), and the authors (C)