The Kapuskasing uplift: a geological and geophysical synthesis[1]

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Over the past decade, the Kapuskasing uplift has been the subject of intense geological and geophysical investigation as Lithoprobe's window on the deep-crustal structure of the Archean Superior Province. Enigmatic since its recognition as a positive gravity anomaly in 1950, the structure has been variably interpreted as a suture, rift, transcurrent shear zone, or intracratonic thrust. Diverse studies, including geochronology, geothermobarometry, and various geophysical probes, provide a comprehensive three-dimensional image of Archean (2.75–2.50 Ga) crustal evolution and Proterozoic (2.5–1.1 Ga) cooling and uplift. The data favour an interpretation of the structure as an intracratonic uplift related to Hudsonian collision.

Eastward across the southern Kapuskasing uplift, erosion levels increase from <10 km in the Michipicoten greenstone belt, through the Wawa gneiss domain (10–20 km), into granulites (20–30 km) of the Kapuskasing structural zone, juxtaposed against the low-grade Swayze greenstone belt along the Ivanhoe Lake fault zone. Most volcanic rocks in the greenstone belts erupted in the interval 2750–2700 Ma and were thrust, folded, and cut by late plutons and transcurrent faults before 2670 Ma. Wawa gneisses include major 2750–2660 and minor 2920 Ma tonalitic components, deformed in several events including prominent late subhorizontal extensional shear zones prior to 2645 Ma. Supracrustal rocks of the Kapuskasing zone have model Nd ages of 2750–2700 Ma, metamorphic zircon ages of 2696–2584 Ma, and titanite ages of 2600–2493 Ma, reflecting deposition, intrusion, complex deformation, recrystallization, and cooling during prolonged deep-crustal residence. Postorogenic unroofing rapidly cooled shallow (10–20 km) parts of the Superior Province, but metamorphism and local deformation continued in the ductile deep crust, overlapping the time of late gold deposition in shear zones in the shallow brittle regime.

Elevation of granulites, expressed geophysically as positive gravity anomalies and a west-dipping zone of high refraction velocities, dates from a major episode of transpressive faulting. Analysis of deformation effects in Matachewan (2454 Ma), Biscotasing (2167 Ma), and Kapuskasing (2040 Ma) dykes, as well as the brittle nature of fault rocks and cooling patterns of granulites, constrains the time of uplift to ca. 1.9 Ga. Approximately 27 km of shortening was accommodated through brittle upper crustal thrusting and ductile growth of an 8 km thick root in the lower crust that has been maintained by relatively cool, strong mantle lithosphere. The present configuration of the uplift results from overall dextral displacement in which the block was broken and deformed by dextral, normal, and sinistral faults, and modified by later isostatic adjustment. Seismic reflection profiles display prominent northwest-dipping reflectors believed to image lithological contacts and ductile strain zones of Archean age; the indistinct reflection character of the Ivanhoe Lake fault is probably related to its brittle nature formed through brecciation and cataclasis at temperatures <300°C. The style and orientation of Proterozoic structures may have been influenced by the Archean crustal configuration.

Au cours de la dernière décennie, le soulèvement de Kapuskasing a fait l'objet de recherches géologiques et géophysiques intensives, tel le transect Lithoprobe de la structure crustale archéenne profonde, dans la province du lac Supérieur. Reconnue en 1950, comme anomalie gravimétrique positive, cette structure a été interprétée différemment, soit une suture, un rift, une zone de cisaillement coulissant ou un chevauchement intracratonique. Diverses études, incluant la géochronologie et la géothermobarométrie, et plusieurs sondages géophysiques ont fourni une image tri-dimensionnelle détaillée de l'évolution crustale à l'Archéen (2,75–2,50 Ga), du refroidissement et du soulèvement au Protérozoïque (2,5–1,1 Ga). Les résultats plaident en faveur de l'interprétation de cette structure comme étant un soulèvement intracratonique relié à la collision hudsonienne.

En traversant, de l'ouest vers l'est, la région méridionale du soulèvement de Kapuskasing, on observe un accroissement des niveaux de l'érosion de <10 km dans la zone de roches vertes de Michipicoten, à 10–20 km dans le domaine des gneiss de Wawa, à 20–30 km dans les granulites de la zone structurale de Kapuskasing, qui est juxtaposée à la zone de roches vertes de Swayze de faible degré métamorphique qui suit la zone de failles du lac Ivanhoe. La majorité des roches volcaniques dans les zones de roches vertes sont le produit d'éruptions survenues entre 2750 et 2700 Ma, elles furent chevauchées, plissées et recoupées par des plutons tardifs et des failles coulissantes antérieurement à 2670 Ma. Le domaine des gneiss de Wawa comporte des composantes tonalitiques majeurs datées de 2750 à 2660 Ma, en outre des composantes tonalitiques datées de 2920 Ma, qui ont été déformées à plusieurs reprises antérieurement à 2645 Ma, incluant les zones de cisaillement extensif majeures d'attitude subhorizontale. Les roches supracrustales de la zone de Kapuskasing fournissent des âges modèles Nd de 2750–2700 Ma, des âges métamorphiques sur zircon de 2696–2584 Ma, et des âges sur titanite de 2600–2493 Ma, reflétant le dépôt, l'intrusion, la déformation complexe, la recristallisation et le refroidissement qui fut étalé sur une longue période de résidence au sein de la croûte profonde. L'effondrement postorogénique de la voûte a refroidi rapidement les matériaux des zones peu profondes (10–20 km) de la province du lac Supérieur, mais à plus grande profondeur, dans la croûte ductile, l'activité métamorphique et le tectonisme persévéraient localement, en même temps que précipitait tardivement l'or dans les zones de cisaillement en régime fragile à faible profondeur.

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Le soulèvement des granulites, correspondant aux anomalies gravimétriques positives et à la zone de vitesses de réfraction élevées inclinée vers l'ouest, coïncide chronologiquement avec le développement des failles en transpression majeures. L'analyse des effets de déformation dans les dykes de Matachewan (2454 Ma), de Biscotasing (2167 Ma) et de Kapuskasing (2040 Ma), ainsi que la fracturation des roches faillées et les modalités de refroidissement des granulites, contraignent l'âge du soulèvement à ~1,9 Ga. Un raccourcissement d'environ 27 km a pu être réalisé grâce aux chevauchements dans la croûte supérieure fragile, et à la croissance ductile d'un racine de 8 km d'épaisseur dans la croûte inférieure, supportée par le manteau supérieur rigide et relativement froid de la lithosphère. La morphologie structurale actuelle du soulèvement résulte de l'ensemble du coulissage dextre, dans lequel le bloc a été fragmenté par des failles dextre, normale et senestre, et modifié ultérieurement par un ajustement isostatique. Les profils de sismique réflexion montrent des réflecteurs dominants, inclinés vers le nord-ouest, interprétés comme l'image des contacts lithologiques et des zones de déformation ductile d'âge archéen; l'absence d'un réflecteur distinctif pour la faille du lac Ivanhoe est probablement due à la déformation fragile sous forme de bréchification et cataclase, à des températures <300°C. Le style et l'orientation des structures protérozoïques ont pu être influencés par la configuration de la croûte à l'Archéen.

[Traduit par la rédaction]

Introduction

The Superior Province, Earth's largest exposed Archean craton, records at the close of Archean time a dramatic transformation from a primitive, primarily oceanic environment to stable granitoid-dominated continental crust. In what may be one of the oldest examples of large-scale plate interaction, prominent lithotectonic subprovinces appear to have been produced by accretionary processes like those active at modern margins. Uplift and erosion that followed, or were part of the cratonization process, have removed several kilometres of material from the province's surface, but erosion levels rarely exceed 10 km.

For a comprehensive view of orogenic processes that formed the Superior Province, information from the third dimension must be integrated into the regional framework (e.g., Fountain and Salisbury 1981). This goal is feasible in the south-central part, where rocks from crustal levels down to 30 km paleodepth are exposed in oblique cross section in the Kapuskasing uplift (Percival and Card 1983, 1985; Percival and McGrath 1986; Percival et al. 1989). Using the natural advantage of variable exposure levels, contemporary problems of crustal construction and deformation can be addressed. Through a combined study of the present three-dimensional crustal geometry and the structure, petrology, and geochronology of the exposed rocks, the history of the crust at various levels may be unravelled to reveal a three-dimensional image of crustal evolution (Percival 1990*a*). For these reasons, Project Lithoprobe (see Clowes et al. 1992) chose to direct one of its first transects toward the Kapuskasing region.

Recent interpretations of the evolution of the southern Superior Province (e.g., Card 1990; Williams 1990) involve successive lateral accretion of volcanic–plutonic terranes, analogous to island arcs (Langford and Morin 1976), and metasedimentary belts, representing accretionary prisms (Percival and Williams 1989), onto a northern cratonic nucleus (Thurston and Chivers 1990; Percival et al. 1992). The late Archean period between 2750 and 2650 Ma witnessed rapid growth in the amount of preserved continental crust, both in the Superior Province and on a global scale.

Nowhere are the inner workings of Archean granite–greenstone terranes exposed in such complete cross section, nor studied to such an extent, as in the Abitibi–Kapuskasing–Wawa region. Here, exceptionally well-preserved supracrustal features, commonly at subgreenschist facies, are juxtaposed along strike with rocks exhumed from granulite-facies conditions. The insights provided by more than a decade of research on the terranes exposed through this unique combination of geological circumstances are summarized in this paper.

Regional geology

The Superior Province was stabilized by ca. 2.5 Ga. Proterozoic tectonism and intrusion by mafic dyke swarms have only qualitatively modified its predominantly cratonic nature. Syntheses of the Superior Province in Ontario and adjacent Quebec have long recognized distinct granite–greenstone and metasedimentary gneiss subprovinces organized in east-striking belts up to 200 km wide (Card and Ciesielski 1986, and references therein) (Fig. 1).

Granite–greenstone terranes, dominated by granitoid and gneissic rocks, contain supracrustal assemblages of mainly volcanic origin, generally metamorphosed to greenschist–amphibolite facies. The metasedimentary belts consist predominantly of greywackes and derived migmatite and granite. Subprovince boundaries are generally marked by late, steep transcurrent faults, with only local evidence of original stratigraphic and structural continuity. In regions where geological mapping is incomplete, or where the craton is beneath thin cover (such as in the Hudson Bay Lowland), the regional structure can be deciphered from aeromagnetic maps.

The Superior Province, a remnant of an even larger craton, is bounded by orogens of Early and Mid-Proterozoic age (Fig. 1). To the northwest, the Trans-Hudson orogen contains Early Proterozoic (2.0–1.8 Ga) oceanic sequences that were accreted to the Superior margin at ca. 1.82 Ga (Weber 1990). To the east lies the New Quebec orogen, deformed at 1.87–1.81 Ga (Skulski et al. 1993). To the south, the Huronian basin formed on the rifted Superior margin ~2.5–2.45 Ga ago (Krogh et al. 1984), and was deformed during Penokean and Hudsonian events at 1.9–1.7 Ga. To the southwest, the Penokean orogen underwent tectonism at 1.86 and 1.84 Ga (Sims et al. 1989). To the southeast, the Grenville Province underwent severe deformation at ca. 1.1 Ga.

Deformation of Early Proterozoic age is recognized near the margins of the craton, as well as in a few internal structures such as the Kapuskasing uplift. Rock units of Proterozoic age within the Superior Province include at least 10 swarms of mafic dykes and at least two sets of carbonatite – alkalic rock complexes. Most are prominent features in aeromagnetic maps.

The Kapuskasing uplift

In the south-central Superior Province (Fig. 2), east-trending Archean belts are interrupted over a distance of 500 km by the northeast-trending Kapuskasing structural zone, made up of high-grade metamorphic rocks and marked by associated strong positive gravity and aeromagnetic anomalies (Fig. 3). Ideas on the origin of the structure include "thinning of the granitic upper crustal layer" (Garland 1950), Mid- to Late Proterozoic

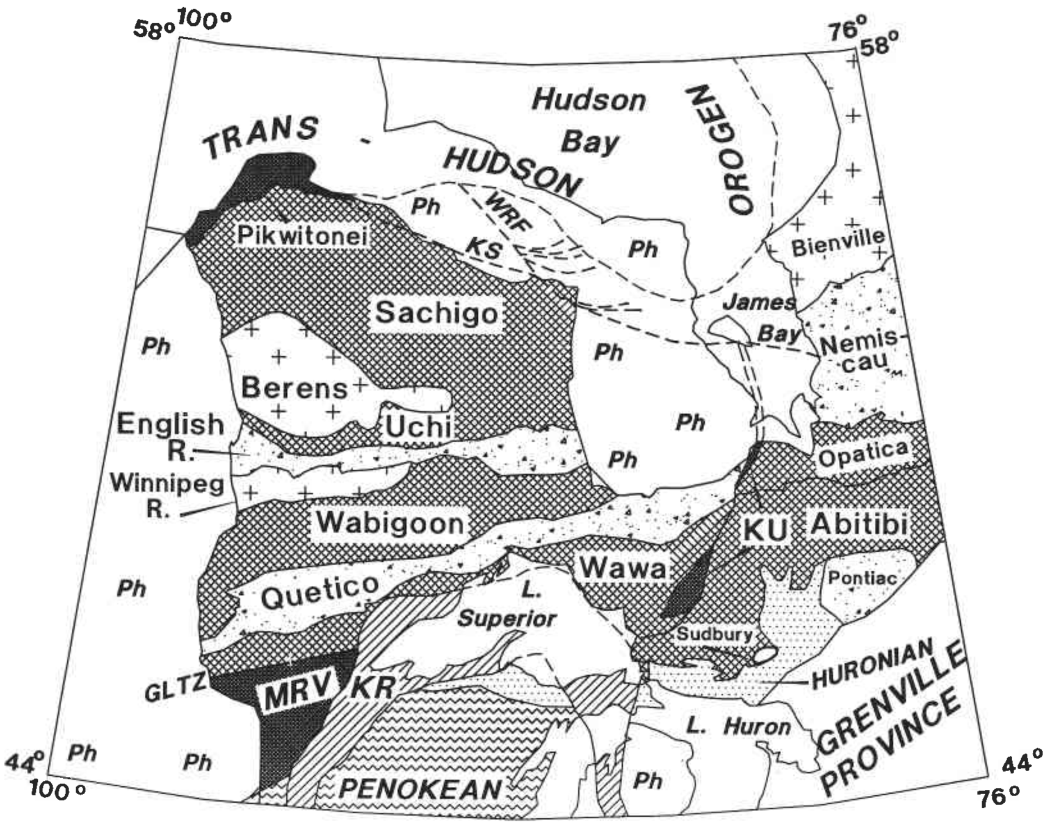


FIG. 1. Regional geology of southwestern Superior Province and surrounding Proterozoic belts. GLTZ, Great Lakes tectonic zone; KR, Keweenawan rift; KS, Kenyon structure; KU, Kapuskasing uplift; MRV, Minnesota River valley; Ph, Phanerozoic cover; WRF, Winisk River fault. Superior Province subprovinces after Card and Ciesielski (1986).

rifting (MacLaren et al. 1968; Bennett et al. 1967; Burke and Dewey 1973; Thurston et al. 1977), a suture between the eastern and western Superior Province (Wilson 1968), sinistral transcurrent faulting (Watson 1980), dextral transcurrent faulting (Goodings and Brookfield 1992), and an east-verging thrust exposing an oblique crustal cross section (Percival and Card 1983; Percival 1986). The Kapuskasing Lithoprobe transect was designed to test the various hypotheses through multidisciplinary studies in the relatively well-exposed southern and central parts of the Kapuskasing uplift.

Well-preserved supracrustal sequences occur in the weakly metamorphosed volcanic belts of the Abitibi and Wawa subprovinces, whereas high-grade metamorphic rocks make up the Wawa gneiss domain and Kapuskasing structural zone. The geological character and geophysical signature of the high-grade areas form the basis for subdivision of the uplift into several lithotectonic elements (Percival and McGrath 1986). From west to east across the southern Kapuskasing uplift they are the low-grade Michipicoten greenstone belt, the Wawa gneiss domain including the Val Rita block, and the high-grade Kapuskasing structural zone (Fig. 2). The Abitibi greenstone belt lies to the east. Gravity and aeromagnetic anomalies are associated with the Kapuskasing zone from James Bay to Chapleau. They attenuate rapidly to the southwest, although a zone of amphibolite-facies rocks does extend toward Lake Superior (Card 1979; Corfu 1987).

Regional geophysics

Geological structures of both local and subprovincial scale are visible in regional aeromagnetic compilations. In particular, the Quetico metasedimentary belt forms a linear, relatively quiet, regional magnetic low separating the curvilinear patterns of variable intensity that characterize the abutting granite–greenstone subprovinces. High-grade domains including both metamorphosed supracrustal and some granitoid rocks more commonly form positive anomalies than do low-grade terranes. Where exposed at the surface, high-grade terranes form numerous intense local anomalies, whereas at depth and where warped towards the surface, they cause broad regional highs.

Bouguer gravity anomalies are commonly negative over granitic regions and extensive, low-grade metasedimentary belts, and are positive over mafic metavolcanic belts, high-grade metasediments, and gneisses. A general trend towards higher background gravity values in the northern part of Fig. 3 is part of a continent-scale free-air gravity low centred on Hudson Bay (Peltier et al. 1992), possibly due to a deep, cold lithospheric and asthenospheric root.

Seismic studies of the Superior Province (Halls 1984, and references therein) indicate a crust averaging about 40–45 km thick, generally including an upper, relatively low velocity (5.9–6.3 km/s) zone separated by a more or less distinct mid-crustal discontinuity from higher velocity (~6.8 km/s) lower crust. Towards the base of the crust, P-wave velocity (Vp) generally rises, often in a complex manner, reaching about 8.1 km/s in the uppermost mantle. Where determinations exist, there is evidence of a further rise to about 8.5 km/s within about 50 km below the Moho. S-wave crustal structure in the crust is less well determined, but appears generally to mimic the P-wave structure. Tomographic teleseismic studies have detected S-wave anisotropy in the mantle lithosphere. The fast

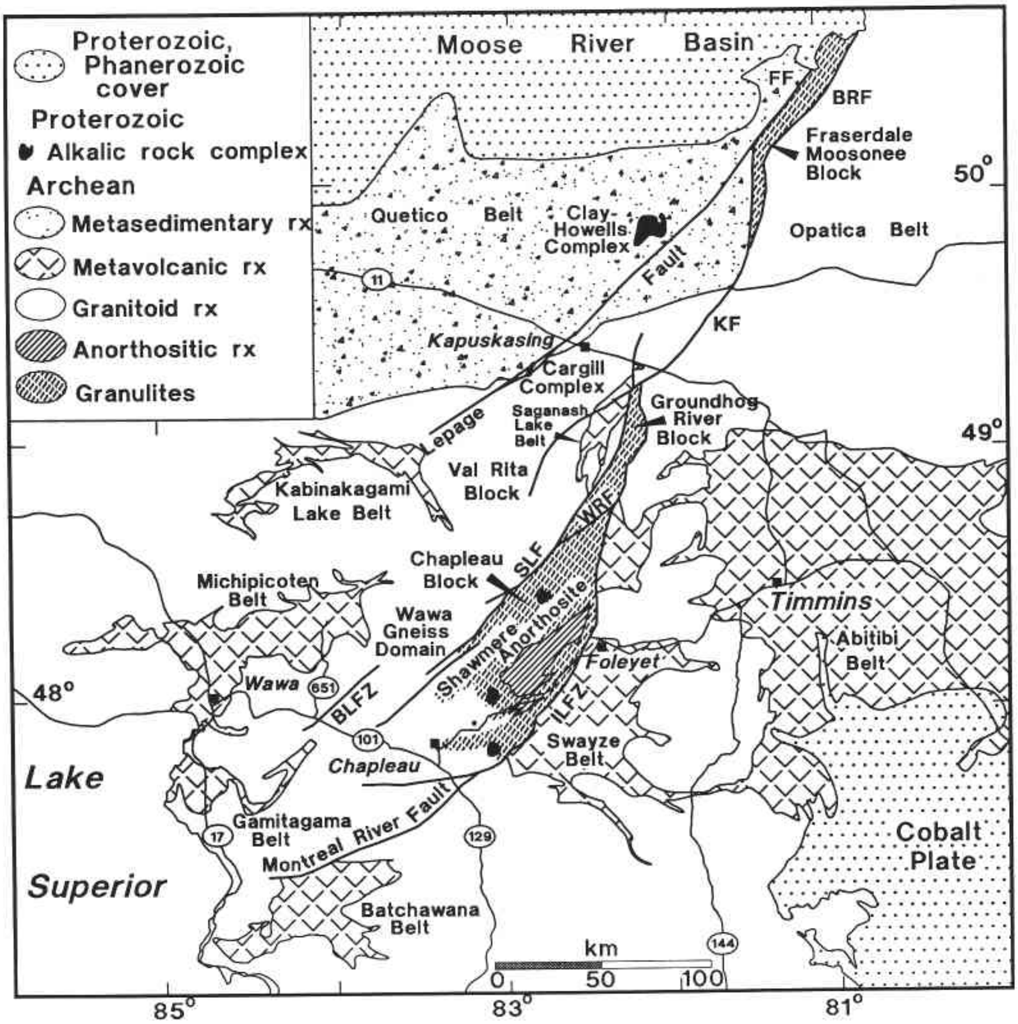


FIG. 2. Geology of the Kapuskasing uplift and surrounding parts of the central Superior Province, showing major geological and geographic features referred to in text. BLFZ, Budd Lake fault zone; BRF, Bad River fault; FF, Foxville fault; ILFZ, Ivanhoe Lake fault zone; KF, Kineras fault; SLF, Saganash Lake fault; WRF, Wakusimi River fault; rx, rock types.

axis is parallel to the east–west crustal fabric, which suggests that the present lithospheric root has been in place since the late Archean (Silver and Chan 1988).

Geology and geophysics of the central Superior Province

Wawa Subprovince

The Michipicoten greenstone belt (Williams et al. 1991) in the Lake Superior area is a stratigraphically and structurally complex assemblage of volcanic, sedimentary, and plutonic rocks, metamorphosed to greenschist and amphibolite facies. On the east, it is intruded by plutonic rocks of the Wawa gneiss domain composed mainly of tonalitic gneiss with abundant granitic and pegmatitic intrusions.

The oldest rocks recognized in the Wawa Subprovince are tonalitic gneissic components of the Wawa gneiss domain, dated by U–Pb on zircon at 2920 Ma (Moser et al. 1991). These rocks are indistinguishable petrographically from younger gneisses and therefore their extent is not known. They resemble in age the Hawk Lake sequence (Turek et al. 1992) on the eastern edge of the Michipicoten belt, composed of mafic volcanic and tonalitic rocks of 2881–2889 Ma. Material of this age may represent basement to the younger (2750 to <2698 Ma) volcanic successions of the Michipicoten belt (Jackson and Sutcliffe 1990). Evidence in the younger volcanics for interaction with sialic crust includes ancient lead isotope values (Thorpe et al. 1987), and evolved major and trace element character (Sylvester et al. 1987).

The volcanic–sedimentary rocks comprise two cycles of mafic-to-felsic character. The age of each is constrained by U–Pb zircon ages of the felsic tops and plutons of synvolcanic character (Turek et al. 1984; 1990). The cycles have lower mafic sections of tholeiitic basalt with some andesite and are capped by calc-alkalic dacite and rhyolite associated with high-level tonalitic intrusions. The Doré conglomerate, containing tonalitic clasts of 2698 Ma, lies in depositional contact on the upper cycle and may be correlative with younger (<2680 Ma) sedimentary rocks (Corfu and Sage 1992). Various facies of iron formation, including the producing sideritic Helen Formation, occur within the volcanic section.

The youngest supracrustal rocks of the Wawa Subprovince occur in the Borden Lake belt of the eastern Wawa gneiss domain. This package of deformed and metamorphosed mafic volcanic rocks, felsic porphyry, conglomerate, and wacke includes conglomerate clasts younger than 2667 Ma (Krogh 1993).

Several suites of plutonic rocks occur within and around the Michipicoten belt. Plutons within the belt, generally tonalite and granodiorite, have ages similar to those of felsic extrusive rocks and are presumed to be synvolcanic intrusions. Most of the Wawa gneiss domain consists of variably foliated gneissic and xenolithic tonalitic rocks whose compositions, including steep rare earth element (REE) patterns, are consistent with derivation from garnet-bearing basaltic sources (Rudnick and Taylor 1986; Truscott and Shaw 1990; Shaw et al. 1994). Tonalites of the Wawa gneiss domain were initially considered to be broadly synvolcanic in age (e.g., Percival and Card 1985). However, recent geochronology shows a wide spectrum of ages for rocks with similar appearance, from 2920 Ma, older than any dated volcanic rocks of the Michipicoten belt, to 2660 Ma (Moser et al. 1991). Near Chapleau, tonalitic units extend from the eastern Wawa gneiss domain northeast into the Kapuskasing zone. Bodies of homogeneous plutonic rock in the Wawa gneiss domain are mainly granitic and granodioritic in composition, and have ages from 2690 to 2633 Ma (Percival et al. 1981; Frarey and Krogh 1986; Moser et al. 1991).

The structural history of the Michipicoten belt is known through detailed structural–stratigraphic studies (Attoh 1981; Arias and Helmstaedt 1990; Williams et al. 1991; McGill 1992). Early structures include thrusts, recumbent folds, and cleavage. Later, superimposed upright folds are accompanied by steep cleavages (McGill and Shrady 1986). Steep, northwest–southeast-trending shear zones, among the latest structures, are host to numerous gold showings (Heather 1989; Fyon et al. 1992), and are associated locally with coarse-clastic deposits that are possibly correlative with the Timiskaming Group (Shegelski 1980; Leclair et al. 1993). Supracrustal and associated plutonic rocks were metamorphosed to greenschist and local amphibolite facies, at pressures of 0.2–0.3 GPa (2–3 kbar) (Studemeister 1983).

To the east, in the Wawa gneiss domain, composite deformation fabrics trend generally east–west, with northerly dips varying from steep in the west to gentle farther east (Moser 1994). Detailed structural studies in seven subareas have recognized at least five deformational events characterized by discrete deformation styles and timing. The earliest structures, including upright cleavage in metavolcanic rocks and gneissosity in tonalites, predate granodiorite of 2677 Ma age. Early structures are reoriented by third-phase tight folds that are also recognized in the Borden Lake belt and hence are younger than 2667 Ma. Fourth-phase structures include ductile subhorizontal shear zones and ductile extensional faults (Moser 1988, 1989, 1994), bracketed by geochronology of deformed plutons and crosscutting leucosome at 2660–2645 Ma (Moser et al. 1991; Moser 1994). Late, north-trending folds constitute a weak, fifth, deformation phase.

The northeastern Wawa Subprovince consists mainly of plutonic rocks of the Wawa gneiss domain but includes the Kabinakagami and Saganash Lake greenstone belts (Leclair and Nagerl 1988; Leclair and Poirier 1989). The structural history is similar to that recognized to the south, including upright folds of early gneissosity, and third-phase ductile subhorizontal extensional shear zones (Leclair 1992; Bursnall et al. 1994). Several major faults transect the area, including the southeast-striking Puskuta Lake shear zone, with movement bracketed between 2682 and 2665 Ma (Sullivan and Leclair 1992*a*, 1992*b*). The north-northeast-trending Lepage and Saganash Lake faults, forming the boundaries of the Val Rita block, are geologically and aeromagnetically mapped faults of probable Early Proterozoic age with inferred northwesterly dip that accommodated ~10 km of normal displacement (Percival and McGrath 1986; Leclair 1990; Leclair et al. 1994; Atekwana et al. 1994).

Geobarometric studies of a suite of tonalites from across the southern Wawa gneiss domain suggest crystallization pressures of 0.5 GPa (5 kbar) in the west, increasing to 0.65 GPa (6.5 kbar) in the east (Fig. 4) (Percival 1990*b*; Mäder et al. 1994). Titanite ages in tonalites vary from 2685 Ma in the west to ca. 2600 Ma in the east (Fig. 5), reflecting complex metamorphic processes and prolonged high temperatures at deep structural levels.

Abitibi Subprovince

The Abitibi Subprovince hosts one of the world's largest and best-preserved Archean greenstone belts. Large parts of the Abitibi belt are metamorphosed only to subgreenschist facies (Jolly 1978). Although many theories on the tectonic setting of Abitibi volcanism have been advanced, those of the past decade have converged on analogues with modern island arcs, back arcs (e.g., Dimroth et al. 1982; Jackson and Fyon 1991), rifted arcs (Ludden et al. 1986), and oceanic plateaus (Desrochers et al. 1993; Kimura et al. 1993).

In a recent review of the southern Abitibi belt, Jackson and Fyon (1991) recognized distinct stratigraphic assemblages to the north and south of the Porcupine–Destor fault zone. The northern region contains komatiitic and tholeiitic volcanic rocks mainly 2730–2700 Ma old (Corfu et al. 1989; Barrie and Davis 1990; Mortensen 1993), whereas the southern part is characterized by assemblages of tholeiitic and calc-alkalic volcanic rocks mainly <2705 Ma, with local units as old as 2747 Ma (Jensen and Langford 1985; Mortensen 1993). Volcanic assemblages of similar age occur in the northern Abitibi belt (Chown et al. 1992).

Volcanic and associated sedimentary rocks in both areas were folded, thrusted, and intruded by granitoid batholiths prior to deposition of an unconformable sequence of fluviatile sedimentary and alkaline volcanic rocks, the Timiskaming Group. Associated spatially and temporally with large deformation zones such as the Porcupine–Destor fault, the Timiskaming Group was deposited ~2680–2677 Ma ago and underwent subsequent folding and faulting prior to intrusion by minor porphyry dykes of 2677 Ma age (Corfu et al. 1991). The steep structures resulting from these several stages of deformation are not directly imaged by seismic reflection surveys, but below a depth of ~10 km a variety of extensive subhorizontal reflectors are observed (Jackson et al. 1990, in press; Ludden et al. 1993). Green et al. (1990*a*) suggested that the truncations associated with the major steep strike-slip faults can be traced to 15 km depth.

The Abitibi belt is bounded to the north by high-grade rocks of the Opatica gneiss belt (Benn et al. 1992), which appear to structurally underlie the lower grade volcanic assemblages (Sawyer and Benn 1993). To the south, metasedimentary rocks of the Pontiac belt apparently extend beneath the southern margin of the Abitibi belt (Dimroth et al. 1982; Jackson et al. 1990; Green et al. 1990*a*; Ludden et al. 1993).

West of the Abitibi belt proper, supracrustal rocks are less abundant. The Swayze belt adjacent to the Kapuskasing zone contains many of the stratigraphic assemblages and structures typical of the southern Abitibi belt (Heather 1993; Heather and van Breemen 1994). Although similar in age to Abitibi rocks (2717 Ma), exact stratigraphic correlation has not been made.

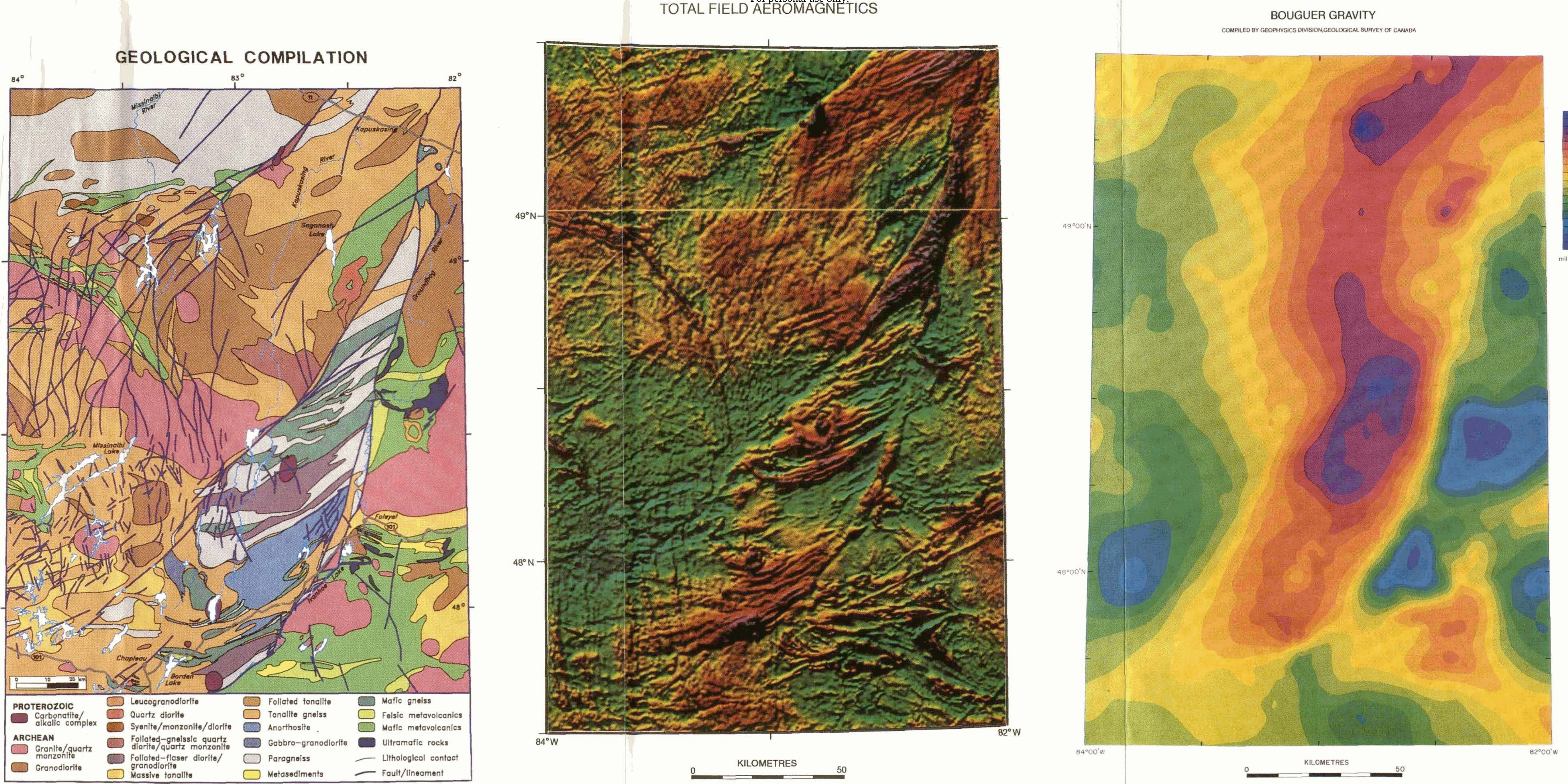


FIG. 3. Regional maps of the Kapuskasing uplift: geological compilation (after Bursnall et al. 1994); shaded relief aeromagnetic map with sun angle set parallel to the Matachewan dyke swarm to highlight Archean units (colours (red: high; blue: low) indicate intensity of the total field anomaly (after West and Ernst 1991)); Bouguer gravity anomaly map. 1 mGal = 10^{-3} cm/s^2.

To the south, the Batchawana belt (Grunsky 1981) contains low-grade volcanic and plutonic rocks ranging in age from 2716 Ma for synvolcanic tonalites to 2673 for posttectonic granite and granodiorite (Corfu and Grunsky 1987). The Batchawana belt is bounded against amphibolite-facies gneisses to the northwest (Corfu 1987) by the Montreal River fault, the probable southwestern extension of the Ivanhoe Lake fault.

Correlation of subprovinces across the Kapuskasing zone is important in distinguishing between models of the Kapuskasing zone as an Archean suture (e.g., Wilson 1968), or an intracratonic structure. Percival and Card (1983, 1985) argued for correlation between the Abitibi and Wawa subprovinces on the basis of similarity in the age of felsic volcanic rocks (2750–2700 Ma; cf. Corfu et al. 1989 and Turek et al. 1984), as well as the character and age of syn- and postvolcanic plutonic rocks (2730–2670 Ma; Frarey and Krogh 1986; Corfu et al. 1991). Subsequent determination of the age of late sedimentary units (ca. 2680 Ma; Corfu et al. 1991; Corfu and Sage 1992) and the nature of major faults (Leclair et al. 1993) further support the correlation. Differences in the prevolcanic history of the two regions were inferred by Jackson and Sutcliffe (1990), who concluded that the Michipicoten belt had been deposited on sialic basement, whereas the southern Abitibi belt formed in an oceanic setting.

Kapuskasing structural zone

The Kapuskasing structural zone can be defined spatially on the basis of its geophysical attributes, or by its geological character, principally its high-grade metamorphic rocks. Although generally associated, the two sets of attributes are not strictly coincident (Fig. 3). The 60 mGal (1 mGal = 10^{-3} cm/s^2) positive gravity anomaly extends sinuously from the Chapleau–Foleyet region in the south to northern James Bay in the north, whereas the complex pattern of associated positive aeromagnetic anomalies from Chapleau to southern James Bay is discontinuous and varies in intensity.

Especially clear in the aeromagnetic expression are numerous sharp truncations and offsets that indicate fault boundaries. The outcrop pattern of granulite-grade rocks coincides well with groups of strong aeromagnetic anomalies over most of the Kapuskasing structural zone. Combined geological and geophysical attributes were used by Percival and McGrath (1986) to define several domains within the Kapuskasing structure (Fig. 2). The Chapleau block in the south has positive gravity and aeromagnetic anomalies and is bounded to the southeast and northwest by brittle fault zones. Generally metamorphosed to granulite facies, rock units including tonalite gneiss and the Borden Lake belt extend downgrade to the west into the Wawa gneiss domain. High-grade rocks also characterize the central Groundhog River block, which has an intense positive aeromagnetic anomaly and negligible gravity expression (Fig. 3). To the east and west it is bound by brittle faults. To the west the Val Rita block grades from granulite facies in the northwest adjacent to the Lepage fault to amphibolite facies in the Saganash Lake belt. A broad positive gravity anomaly occurs in the west. A 65 km gap without granulites separates the Groundhog River block from the northernmost Fraserdale–Moosonee block, characterized by a positive aeromagnetic anomaly and high-grade paragneiss.

Crustal structure

Crustal refraction experiments centred on Lake Superior in 1963 and 1965 (Halls 1984) were the first to show an anomalous seismic velocity structure associated with the Kapuskasing structure (at that time known mainly from its gravity anomalies). In 1984, as an initial task, Lithoprobe conducted a regional refraction survey of a roughly 300 km × 300 km region centred on the northern Chapleau block (Northey and West 1985). Although detailed for its time, this survey was carried out with only 30 recording units, in comparison with the several hundreds employed in more recent, higher resolution experiments.

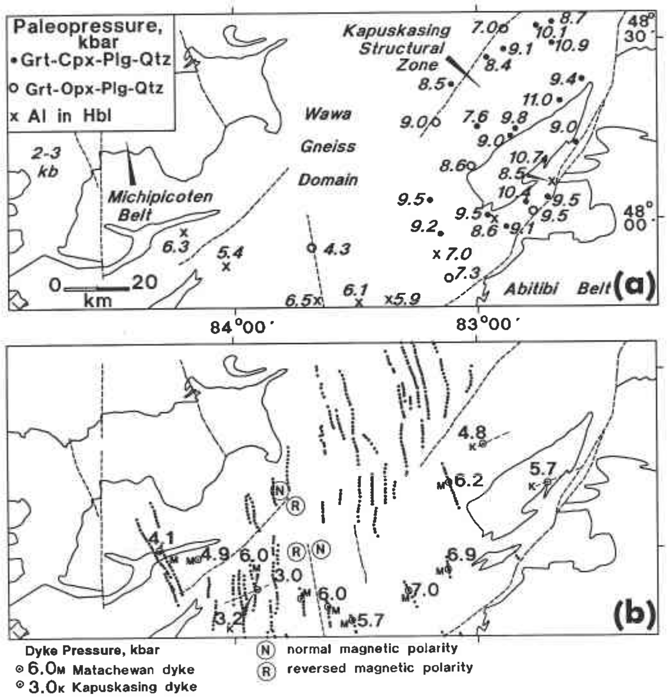


FIG. 4. Summary of geobarometric data for the Kapuskasing uplift. (*a*) Paleopressure from Archean rock units (after Mäder et al. 1994), yielding the magnitude of erosion since high-grade metamorphism (2696–2625 Ma) and magmatism (ca. 2700–2660 Ma). (*b*) Estimates of emplacement pressure for Matachewan (2454 Ma) and Kapuskasing (ca. 2040 Ma) mafic dykes (after Percival et al. 1994). The difference between the metamorphic and Matachewan pressures gives uplift magnitude in the interval 2650–2450 Ma; between Matachewan and Kapuskasing, the uplift between 2450 and 2040 Ma, and the Kapuskasing pressures, uplift since 2040 Ma.

Structural interpretations of the data were presented by Boland (1989) and Wu and Mereu (1990). Both substantiated the presence of thick crust beneath the Kapuskasing uplift, but differed in detail. Strong amplitudes were found in the seismograms at traveltimes and distances appropriate to the wide-angle reflections, but discrete reflections from the base of the crust generally were indistinct, leading to significant ambiguity in determining velocities and structure in the lower crust and uppermost mantle (White and Boland 1992). Both interpretations revealed anomalously high velocities in the upper crust beneath the Kapuskasing structural zone. In a west–east profile, the depth at which high velocity (6.5–6.8 km/s) is reached rises from a typical regional level of ~20 km beneath the Michipicoten belt to within ~5 km of the surface in the Kapuskasing zone.

Late-arriving energy is strong in the data, implying a complex crust–mantle boundary (Wu and Mereu 1990). Preferring an interpretation in which velocity heterogeneity or P-wave anisoptropy in the mantle under the region was minimal, Boland et al. (1988) and Boland and Ellis (1989, 1991) concluded that

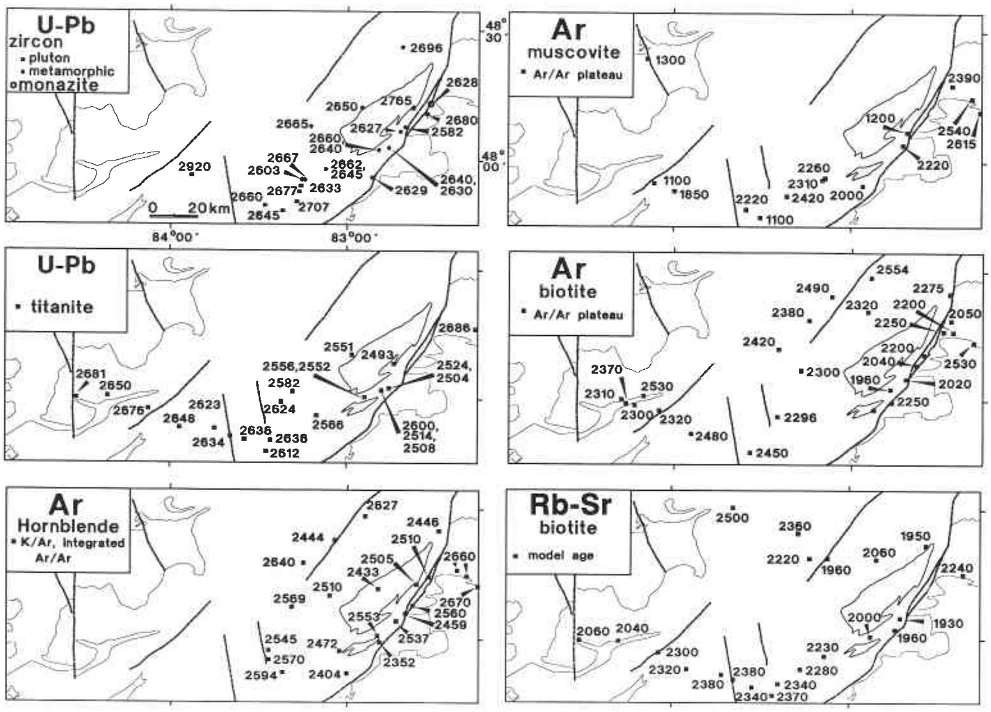


FIG. 5. Summary of geochronological information for the Chapleau block and surrounding Abitibi and Wawa subprovinces. U–Pb zircon and monazite data after Krogh and Moser (1994); U–Pb titanite from Krogh (1993) and T.E. Krogh and J.A. Percival (unpublished data); Ar hornblende, muscovite, and biotite data compiled from Hanes et al. (1994); Rb–Sr biotite data from Percival and Peterman (1994).

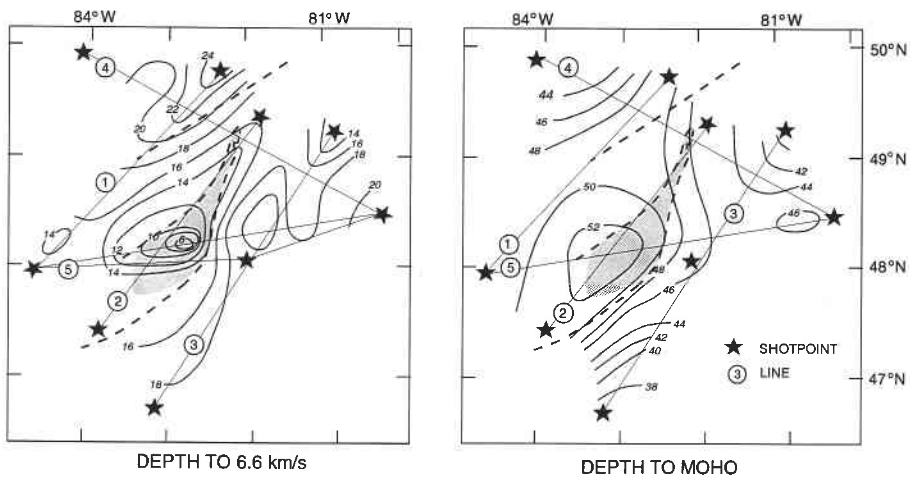


FIG. 6. Compilation map of seismic refraction results. Left: depth to 6.6 km/s velocity contour showing the presence of high velocities at unusually shallow depth beneath the Chapleau block. Right: crustal thickness in the central Superior Province (after Boland and Ellis 1989; Wu and Mereu 1990).

the mantle velocity is 8.1 km/s at the Moho, in close accord with results from earlier surveys in other parts of the Superior Province, and that the crust is thicker by about 8 km than background levels of ~45 km in a broad zone centred beneath the Chapleau block. Contour maps and sections of the Vp structure based on their interpretation are shown in Figs. 6 and 7. 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), analyzing individual profiles, inferred a simpler structure for the upper crust and projected the high-velocity zone and region of thick crust farther northeast. They also found that S arrivals generally mimic P arrivals (Poisson's ratio = 0.25). Laboratory measurement of Vp and Vs (shear-wave velocity) on samples link the seismic characteristics observed at depth to rock types exposed in the Kapuskasing structure (Fountain et al. 1990; Salisbury and Fountain 1994).

A variety of regional and detailed electromagnetic studies have been carried out in the Kapuskasing region, especially in the Chapleau block (Mareschal 1990; Mareschal 1994). A reconnaissance investigation using an array of long-period magnetometers showed that there is no significant regional-scale conductivity anomaly associated with the Kapuskasing zone (Woods and Allard 1986). Subsequent reconnaissance magnetotelluric measurements (Kurtz et al. 1989) detected a zone of enhanced conductivity in the mid- to lower crust, as

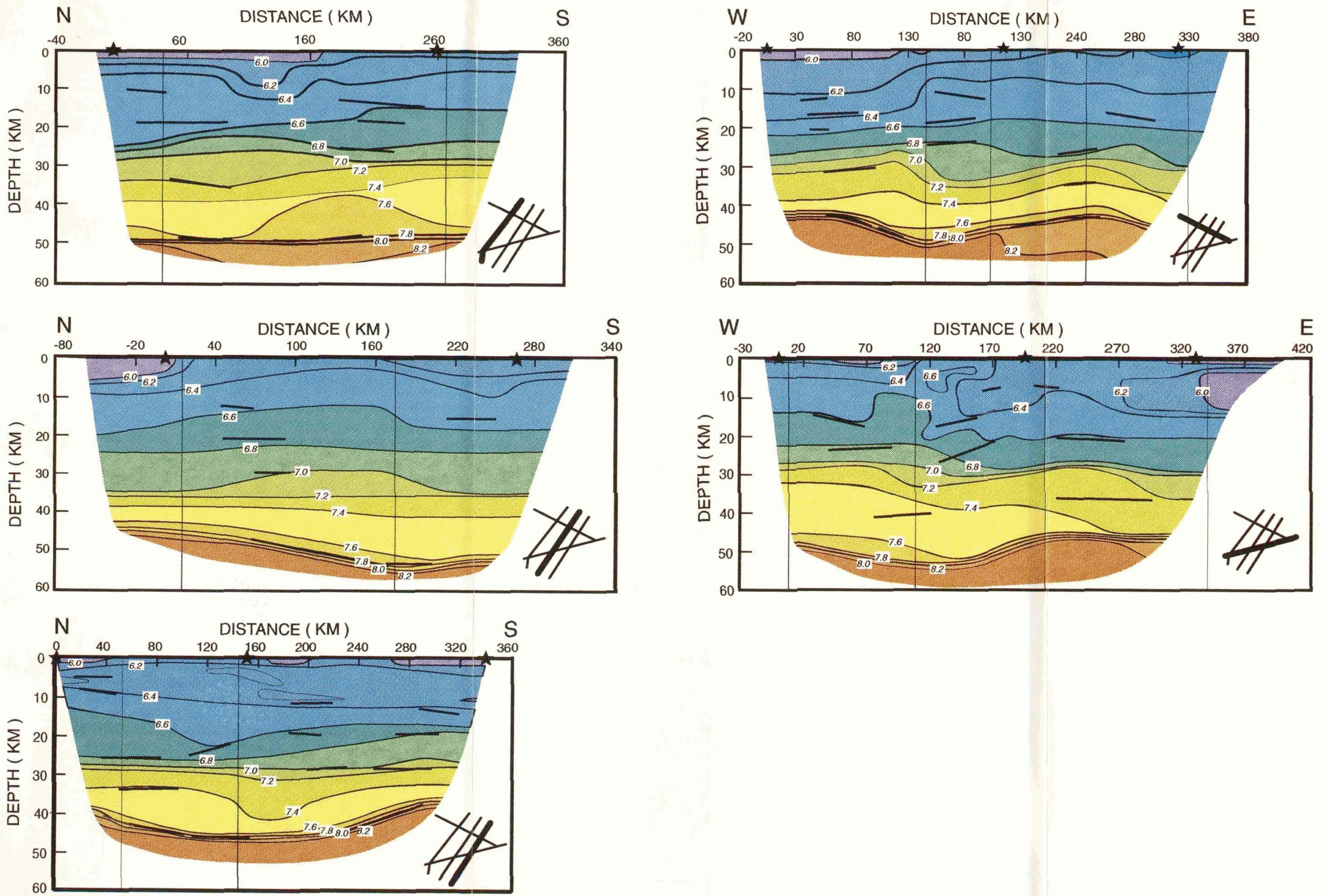


FIG. 7. Crustal sections for the Kapuskasing region showing Vp based on Lithoprobe refraction data (after Boland and Ellis 1989). Line locations are plotted in Fig. 6.

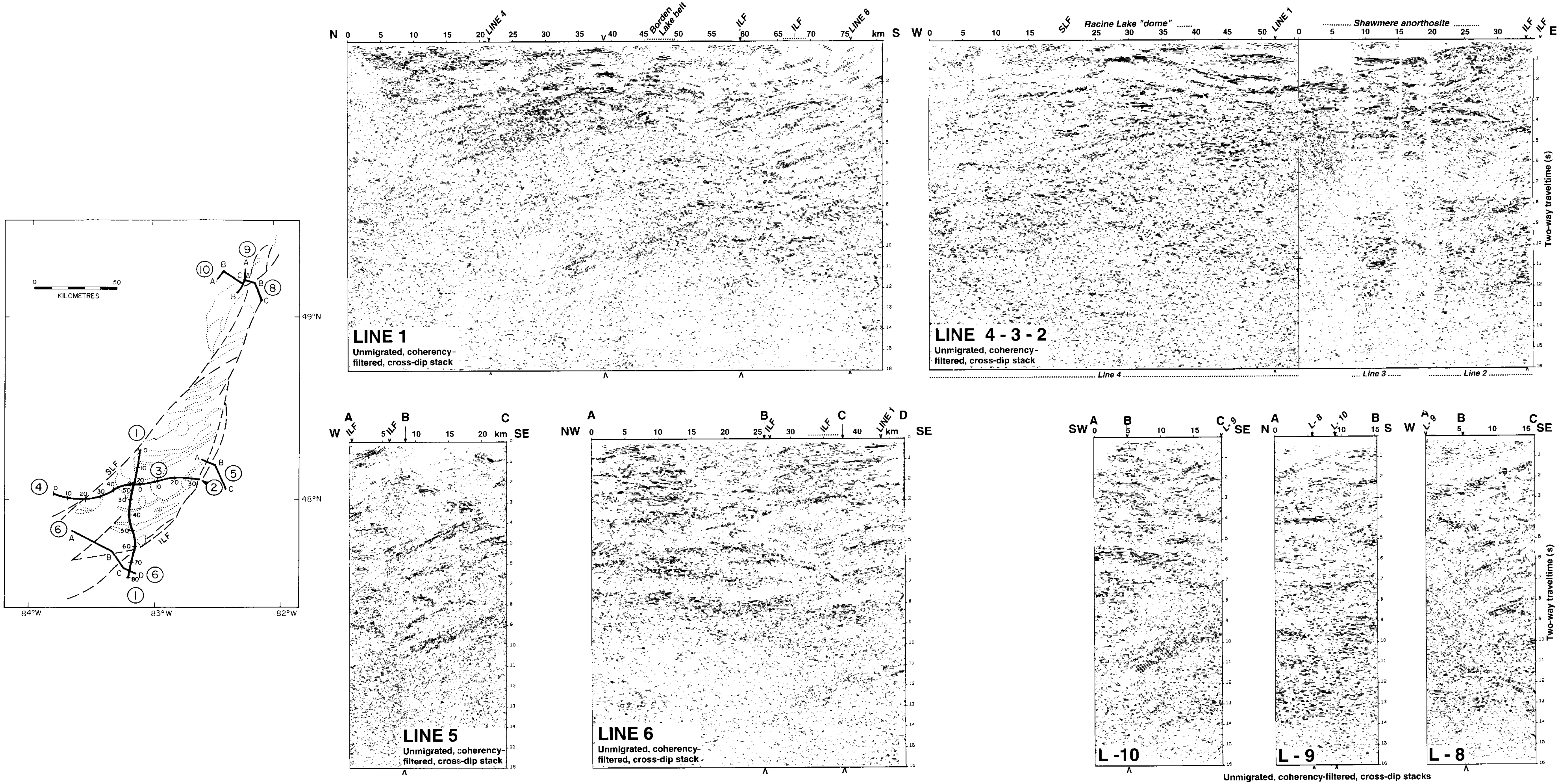


FIG. 8. Reprocessed profiles of Lithoprobe regional reflection seismic data. Line locations are identified in the map. ILF, Ivanhoe Lake fault zone; SLF, Saganash Lake fault.

is typical of many areas. Many authors have suggested that lower crustal conductivity is due to trapped, overpressured, saline water in dilated fractures. Based on observations of graphite grain-boundary films in Kapuskasing rocks (Mareschal et al. 1992), Mareschal et al. (1994) relate it to the presence of graphite, the conductive effects of which were presumably destroyed when the films cracked during exhumation.

Chapleau block

The Chapleau block is the widest, most accessible, and most thoroughly studied part of the Kapuskasing structure. Although also the best exposed, outcrop rarely exceeds a few percent of the total area. The block is bounded against the Abitibi belt to the southeast by the Ivanhoe Lake fault zone and against the Wawa gneiss domain to the northwest by the Saganash Lake fault. South of Chapleau, the Ivanhoe Lake fault appears to splay into a west-trending strand and a southwest-trending extension that may connect with the Montreal River fault. Units including tonalite gneiss, mafic gneiss, and metaconglomerate extend southwestward from the Chapleau block into the Wawa gneiss domain (Bursnall et al. 1994; Moser 1994).

Rock types of the Chapleau block include mafic gneiss and paragneiss, thought to be of supracrustal origin (Percival 1983). Their age is probably similar to that of the supracrustal component of the Abitibi and Wawa subprovinces, based on model Nd ages (Shaw et al. 1994). Although the paragneisses are geochemically similar to low-grade greywackes, the mafic rocks have oxygen isotopic signatures indicating a lack of low-temperature exchange with seawater (Li et al. 1991; Shaw et al. 1994).

Supracrustal rocks of the Borden Lake – Hellyer belt (Bursnall et al. 1994) include distinctive conglomerate and pillowed mafic volcanic rocks (Moser 1988) with oxygen isotopic ratios typical of submarine deposits (Li et al. 1991). Based on U–Pb zircon ages of conglomerate cobbles of 2690, 2680, and 2667 Ma (Krogh 1993), the unit is considered to be one of the youngest in the region. Intrusive rocks of the Chapleau block include anorthosite of the Shawmere complex (Simmons et al. 1980; Riccio 1981), as well as abundant sheets of tonalite and diorite (Percival 1981*a*). Primary ages are available for few units of the Kapuskasing zone, owing to the effects of high-grade metamorphism. Tonalite adjacent to the Shawmere complex contains zircon with a minimum age of 2765 Ma (Percival and Krogh 1983) and tonalite from near the Ivanhoe Lake fault zone has zircons of 2826 and 2717 Ma age (Krogh and Moser 1994).

At least four sets of structures are recognized within the Chapleau block. Layering in mafic gneiss and paragneiss predates intrusion of regional tonalite bodies. Small, second-generation folds trend northwest and are overprinted by pervasive third-phase structures, including large-scale south-verging folds, probably associated with fourth-phase ductile shears (Bursnall et al. 1994). Constraints on the timing of the deformation include minimum ages from little-deformed pegmatites of 2582 Ma age that cut D_4 shears (Krogh 1993).

Ductile deformation (phases 1–4) was broadly synchronous with high-grade metamorphism, which varies from upper amphibolite to granulite facies within the block. High-grade mafic rocks commonly contain the assemblage garnet–clinopyroxene–hornblende–plagioclase–quartz, whereas associated paragneisses have garnet–biotite–plagioclase–quartz ± orthopyroxene. Regional thermobarometry on these assemblages yields temperature estimates generally in the range 720–780°C, at pressures of 0.8–1.1 GPa (8–11 kbar) (Fig. 4*a*) (Hartel 1993, Mäder et al. 1994). A regional pattern of high (> 1 GPa (10 kbar)) pressures in a north-trending arch is at a high angle to the north-northeast–south-southwest structural grain. Assemblages in D_4 ductile shears record pressures lower, by ~50 kPa (0.5 kbar), than those in associated mafic gneiss (Hartel 1993), suggesting that 1–2 km of uplift may have accompanied this deformation.

Fluids associated with the metamorphism were depleted in H_2O (a_{H_2O} ~0.1–0.5; Mäder et al. 1994) and probably contained CO_2, as deduced through studies of "primary" fluid inclusions and observations of graphite grain-boundary films (Mareschal et al. 1992). Rock-to-fluid ratios were on the order of 1:1 and fluid flow may have been predominantly parallel to lithological contacts (Puris and Wickham 1994). Fluids preserved in inclusions in alteration assemblages have a range of compositions, generally involving CO_2. Some of these may be primary, representing metamorphic fluids (Rudnick et al. 1984), but most probably formed in response to uplift and late alkalic rock magmatism (Lamb and Morrison 1992; Channer and Spooner 1994).

Metamorphic zircons from mafic gneisses are inferred to date granulite-facies metamorphism because rocks of equivalent composition at lower grade do not contain zircon (Percival and Krogh 1983). However, the exact conditions of zircon crystallization and recrystallization are probably a complex function of temperature, strain, bulk, and fluid composition. A range of ages from 2696 to 2582 Ma (Fig. 5) (Krogh 1993) has been documented, although most ages from the Chapleau block are in the range 2663–2630 Ma. The oldest age, 2696 Ma, on a mafic gneiss with 1.1 GPa (11 kbar) paleopressure (Mäder et al. 1994), indicates that the crust was at least 30 km thick by that time.

In the southern Chapleau block, zircon ages appear to decrease with increasing paleodepth, from 2663 Ma near the western limit of granulite facies to 2636, 2616, and 2582 Ma for components of a complex population (Krogh 1993). For example, single outcrops showing primary zircons in the range 2660–2640 Ma are cut by pegmatites and retrogressively altered to amphibolite at 2640, 2630, and 2582 Ma (Krogh 1993). Regional geobarometry (Mäder et al. 1994) shows that granulites of different apparent age formed at similar structural level. Either the metamorphism was polychronous or factors other than temperature strongly influenced the zircon age pattern.

An asymmetric +60 mGal gravity anomaly coincides with the Chapleau block, its sharper eastern edge corresponding to the Ivanhoe Lake fault zone (Fig. 3). Based on weighted average rock densities of 2820 kg/m^3 for the Chapleau block and 2700 kg/m^3 for the surrounding granitoid regions, a west-dipping slab geometry matches the observed anomaly (Percival and Card 1983). The intensity of the positive aeromagnetic anomaly (Fig. 3) can be modeled using measured susceptibilities (Shive and Fountain 1988), although no simple correlation of high susceptibility with map units is apparent.

Seismic reflection profiling

A major component of the Lithoprobe program was the acquisition of 358 km of crustal seismic reflection profiling carried out in 1987–1988, largely in and near the Chapleau block. Although limited by the sparse road network, surveys were laid out so as to cross contacts between high- and low-grade rocks. Preliminary interpretations have been given by Cook (1985), Percival et al. (1989), and Geis et al. (1990). Composite line 4–3–2 and its nearby extension (line 5) cross the southern Chapleau block in a generally west to east direc-

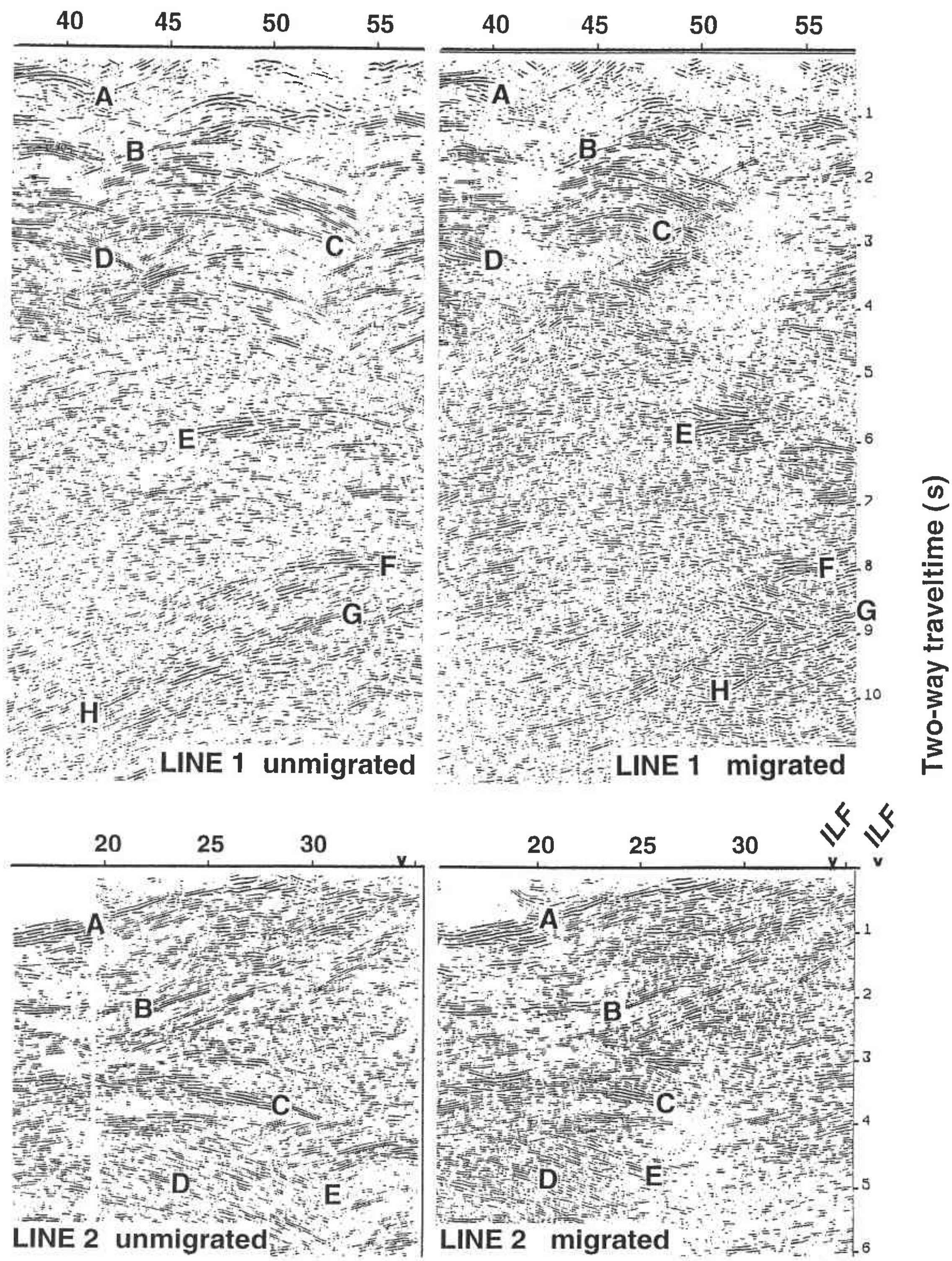


FIG. 9. Comparison of unmigrated and migrated seismic data sections (lines 1 and 2; see Fig. 8 for line locations), showing how two-dimensional migration repositions reflection events (strong reflections are labelled). Successful two-dimensional migration places reflections close to the reflector's true structural position along the profile, but the reflector may (depending on cross dip) lie on or to either side of the profile section anywhere compatible with the reflection time. However, events in the migrated sections may not be accurately repositioned because of data imperfections and complex, three-dimensional geology. ILF, Ivanhoe Lake fault zone.

tion from the eastern Wawa gneiss domain into the western Abitibi belt, traversing many of the structures in the Chapleau block approximately along strike (Fig. 8). Line 1 runs approximately north–south, and crosses line 4–3–2 almost perpendicularly. Line 6, which comes close to linking lines 1 and 4–3–2, is in the southernmost Chapleau block.

Initially, an elementary processing sequence was applied to the seismic data, and the quality of the resulting sections varied greatly. Consequently, all data were reprocessed. A compilation of the unmigrated sections is shown in Fig. 8 and a com-

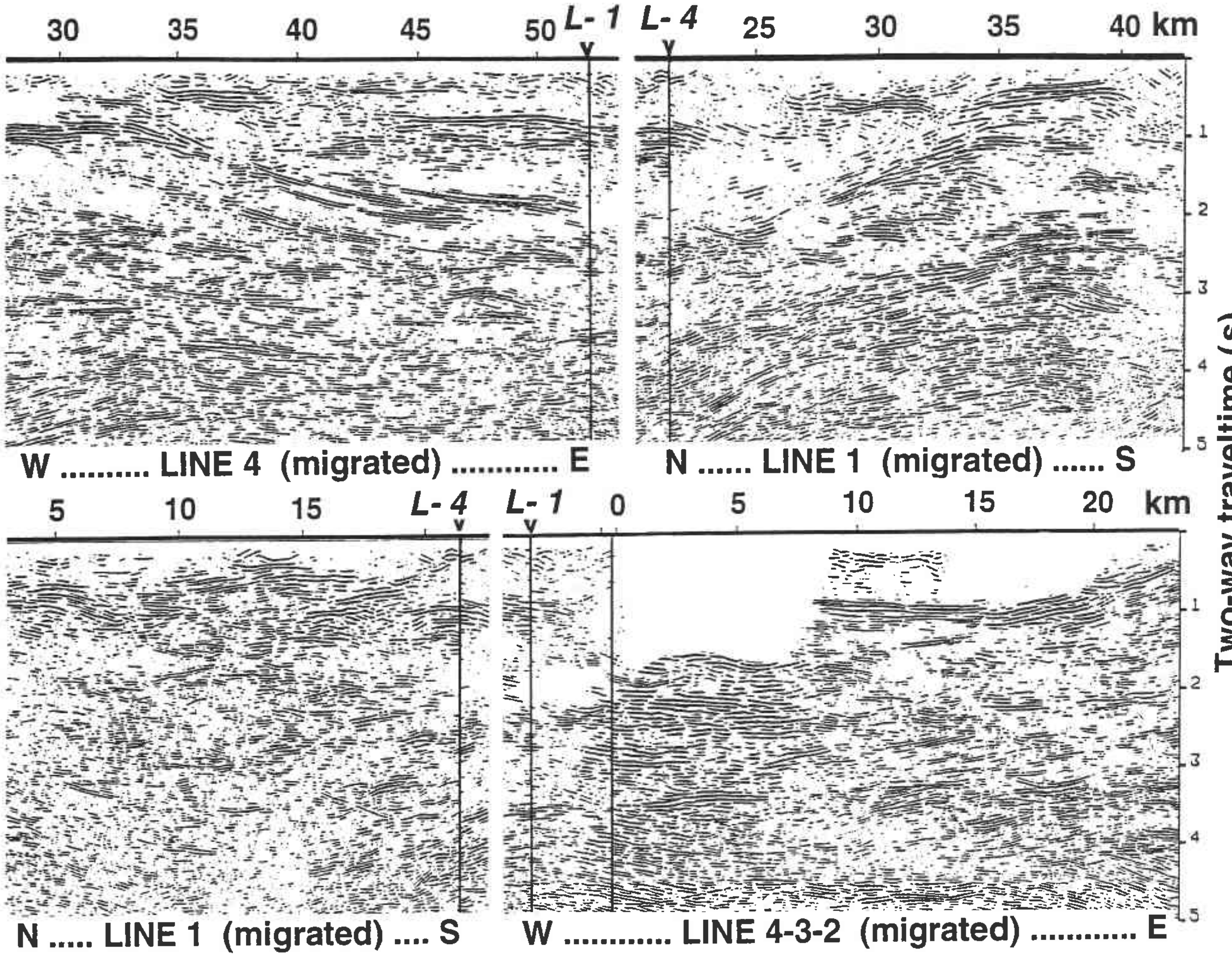


FIG. 10. Comparison of reflection character near the intersection of lines 1 and 4 (see Fig. 8 for line locations). Reflections are much smoother and more horizontal in the east–west than north–south direction. The sections along the perpendicular lines have been split about the crossing point and repositioned to illustrate the correspondence of events.

parison of migrated and unmigrated profiles in Fig. 9. In both the initial and the final processed data, strong reflectivity is widely observed in the upper crust. At greater depth, the strength and continuity of reflectors generally fade. No events associated with the Moho are recognizable. A striking characteristic of the Chapleau block profiles, most evident after migration, is the very different appearance of near-surface reflectivity on east–west and north–south profiles (Fig. 10). Migrated reflectors in the upper 4 s (~12 km) of line 4–3–2 image as strong, highly continuous, subhorizontal features, whereas they generally appear as short, discontinuous, dipping, convex-up events on line 1 (Fig. 10). The asymmetric pattern suggests that most strong reflections arise from anticlinal tops of moderately corrugated surfaces striking roughly east–west.

Overall (as is clear in the unmigrated sections of Fig. 8), there is a marked change in the intensity and character of the reflectivity between shallower and deeper parts of the sections that generally occurs in the depth range 6–16 km (2–5 s). The base of the shallow layer dips north at about 15°, and is about 12 km (4 s) deep at the intersection of lines 1 and 4 (Figs. 8, 10). Reflections in the deeper parts of the sections are more prominent beneath the east and south sides of the Chapleau block and under the Abitibi belt than in the west, and they mostly dip 20–45° to the northwest. The deep regime may come to the surface at the south end of line 1.

Near the surface, the western end of line 4 within the Chapleau block is characterized by a prominent arch whose culmination coincides with the Racine Lake "dome" (line 4, Fig. 8) (Percival 1981*a*; Moser 1994). West of the culmination, reflectors dip gently into the Wawa gneiss domain. On the preliminary sections, the westward transition appears gradational, but individual reflectors are more distinct in the final sections, and patterns of truncation consistent with fault displacement can be identified (Fig. 8).

In a special effort to image the Ivanhoe Lake fault zone in the upper crust, line 2 was also recorded to high-resolution (HR) specifications (Fig. 11). The general pattern of shallowly dipping reflectivity seen on line 4–3–2 and line 2 HR 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 (Fig. 12).

In the vicinity of lines 2 and 5, the Ivanhoe Lake fault zone includes two subparallel branches several kilometres apart. The intervening region contains amphibolite-grade rocks that appear to be a hybrid of Kapuskasing gneisses and rocks typical of the Swayze belt (Percival 1981*a*). Although it was

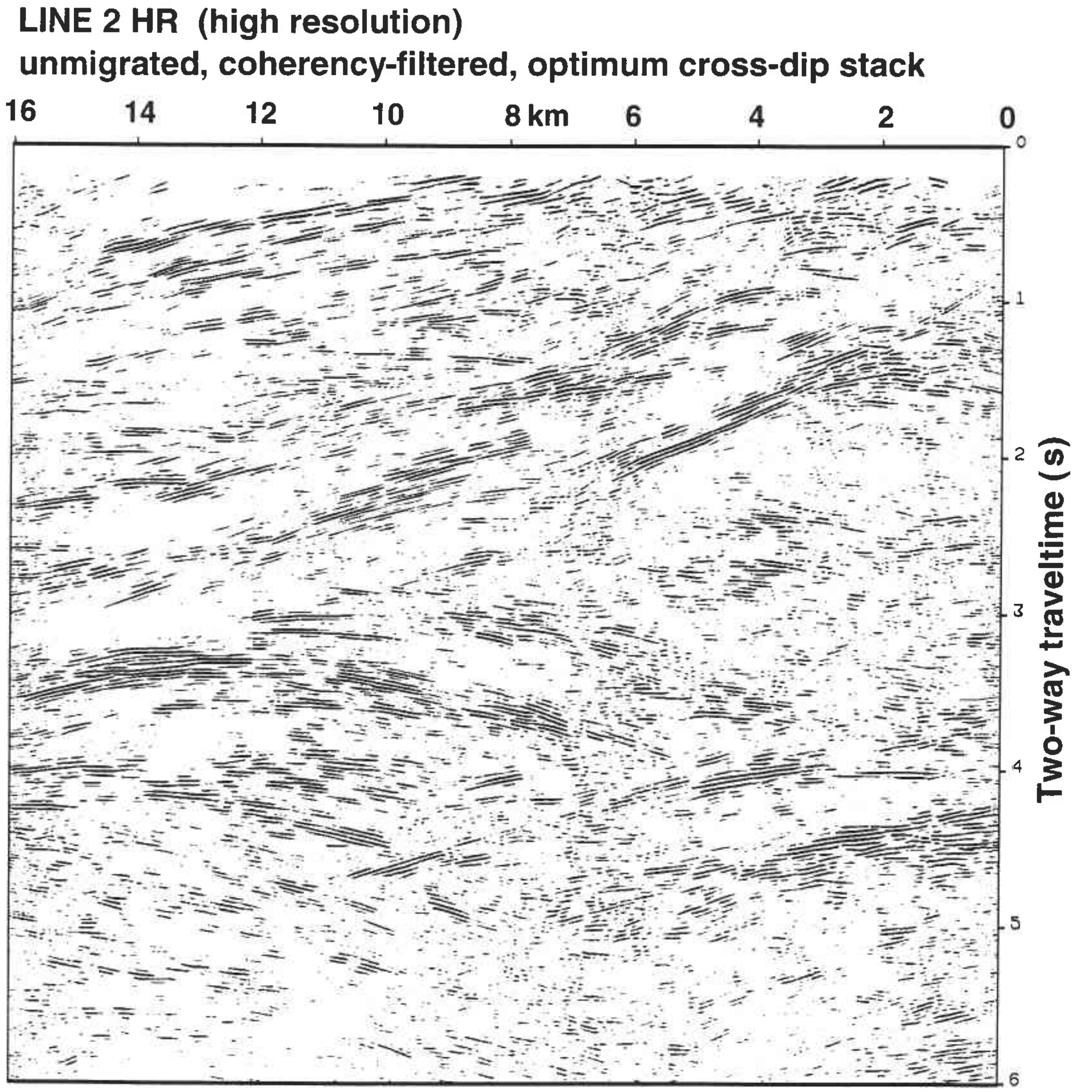


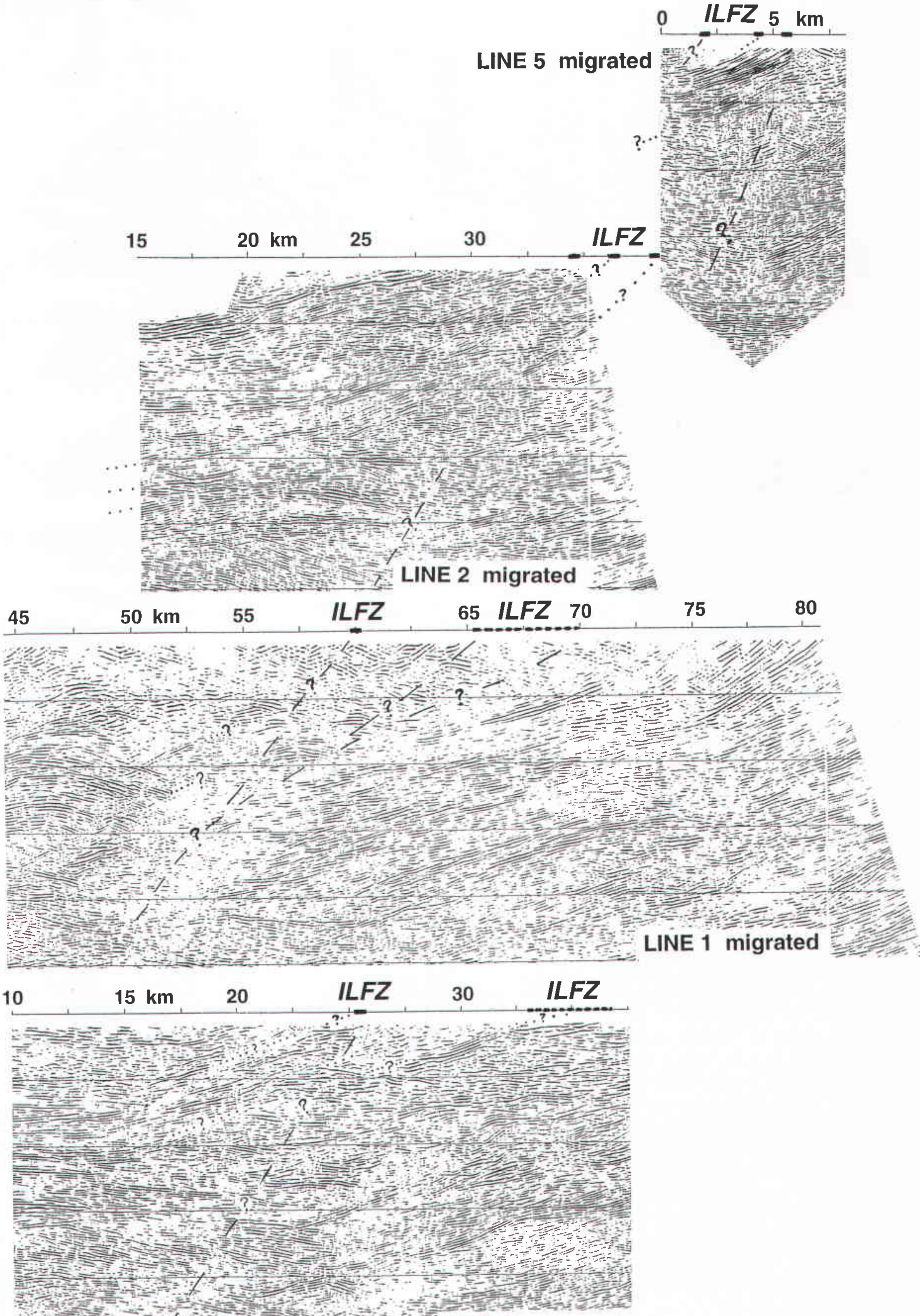
FIG. 11. Line 2 HR (high resolution) shown in unmigrated, coherency-filtered form to emphasize the continuity of events. The near-surface event from 8 to 14 km is believed to correspond to the base of the Shawmere anorthosite complex. A very strong event (dipping approximately 40°W) at about 2 s (~6.2 km depth) is likely associated with the Ivanhoe Lake fault zone. An extensive hyperbolic event with apex at about 3.3 s and 13 km collapses greatly under migration.

impossible to cross the Ivanhoe Lake fault completely on line 2, line 5 crossed the structure in a region of very sparse outcrop. After careful reprocessing (Wu et al. 1992) (Fig. 12), this section displays a remarkable sequence of near-surface reflections dipping about 20–30° west-northwest throughout the 5 km wide zone.

Line 2 (regional and HR; Figs. 8, 11, 12) contains a prominent, near-horizontal, laminar band of reflections at very shallow depth immediately east of the Shawmere anorthosite. Traceable to within a kilometre of the surface and therefore within reach of an ordinary diamond-drill hole, these events have attracted much interest (Green et al. 1990*b*; Percival et al. 1991; Milkereit et al. 1991) and have been the subject of special studies. Their true dip of 17°W (Kim et al. 1992) projects to surface outcrops of interlayered tonalitic and mafic gneiss, which underlie the anorthosite and have appropriate scale and velocity contrasts to produce the observed reflection response (Percival et al. 1991; Milkereit et al. 1991; White et al. 1992). However, these units are folded about gently (15–20°) west-plunging axes, yielding layer dips in the range ~10–80°N (Bursnall and Moser 1989; Percival et al. 1991), whereas a small-scale reflection-seismic survey on a pair of crossed lines sees the structures as nearly planar.

Controlled-source electromagnetic surveys show that a layer

FIG. 12. Comparison of migrated data profiles that cross the Ivanhoe Lake fault zone (ILFZ; see Fig. 8 for line locations). All sections show reflections with north or west dips of less than 45° that project toward the surface trace of the faults. However, there is no clear one-to-one correspondence of reflections with mapped faults. The shallow-dipping, fault-correlated events seem to blend west and northwards into a diffuse pattern of reflectors (seen inside and outside the Kapuskasing structural zone) that are likely related to Archean structures that may have been deformed during the Kapuskasing uplift event. The reflections are discontinuous in many places, and could easily have been cut by steeply dipping brittle faults that exhibit no direct reflections. Dotted lines on the sections indicate some of the interpreted Archean structures; broken lines outline some possible brittle faults.


0
ILFZ
5 km
LINE 5 migrated
15
20 km
25
30
ILFZ
LINE 2 migrated
45
50 km
55
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65
ILFZ
70
75
80
LINE 1 migrated
10
15 km
20
ILFZ
30
ILFZ
LINE 6 migrated

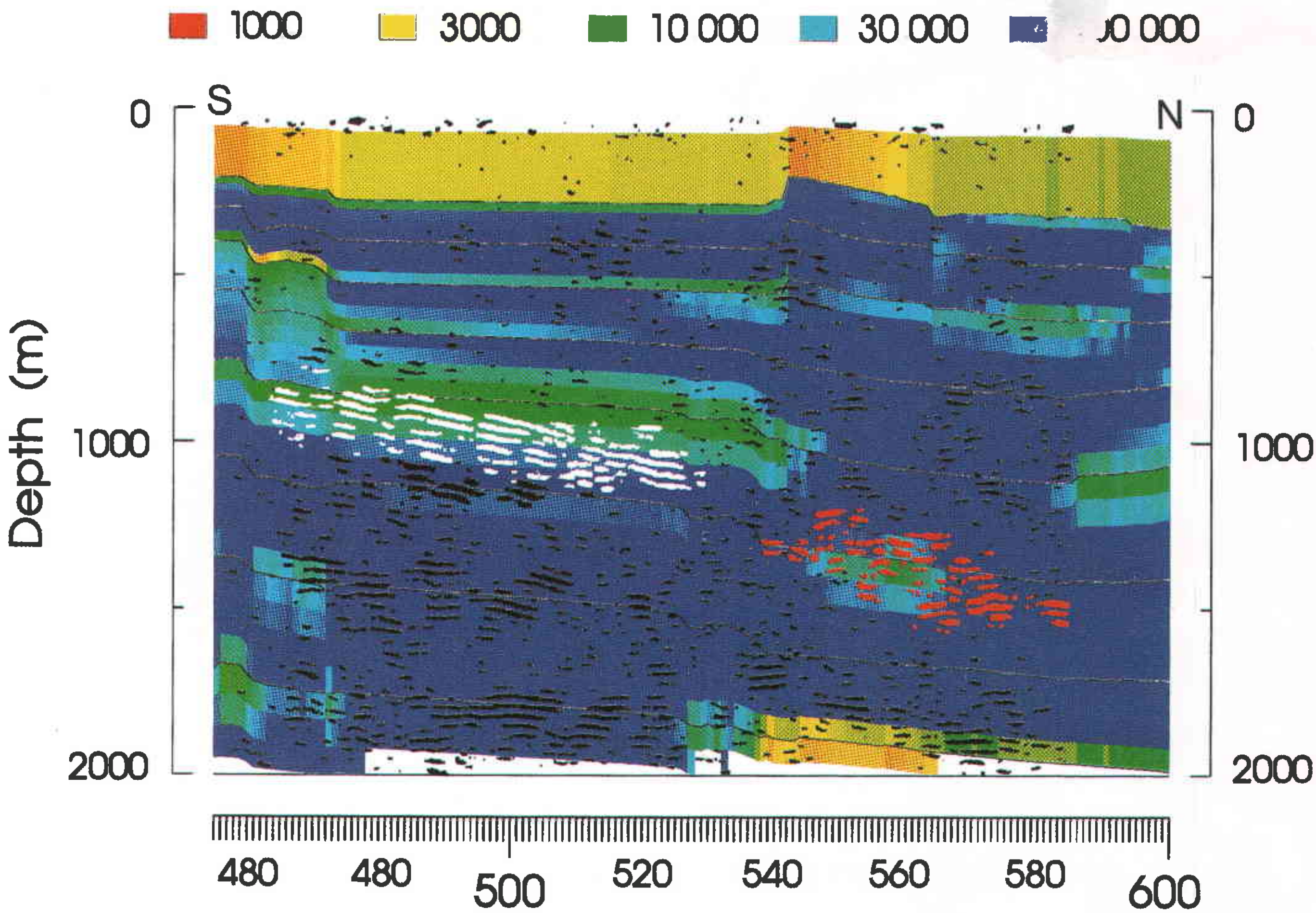

FIG. 13. High-resolution UTEM electromagnetic image of resistivity (Ω · m) showing enhanced conductivity (green and light blue relative to deep blue) in the vicinity of strong seismic reflections. The electrical results for the north – south profile crossing line 2 (Fig. 8) are superimposed on reflections from a coincident high-resolution dynamite-source seismic reflection survey from Milkereit et al. (1991). Coherent groups of reflections are highlighted in black, white, and red. The gently north-dipping dark lines are various time channels, with the earliest channel (0.0144 ms) giving the shallowest information (modified after Jones et al. 1994).

of slightly enhanced electrical conductivity (in highly resistive host rock) nearly coincides with the seismic reflectors (Fig. 13), leading to the possibility that both are caused by an extensive zone of near-horizontal open fracturing (Bailey et al. 1989; Jones et al. 1994). However, in general, the configuration of crustal seismic reflectors and their correlation with bedrock features make it likely that most (if not all) are imaging Archean or Early Proterozoic tectonic or lithologic boundaries.

Lines 8, 9, and 10 in and around the Groundhog River block show strong reflectivity throughout the crust, dipping in a variety of directions (Fig. 8; see also Leclair et al. 1994). Correlations with the geology can be made, but the shortness and crookedness of the lines make meaningful migration (and thus positioning of the deeper reflectors) very difficult and identification of regional patterns impossible. Interpretations of deep structure are thus speculative.

The geological cause of ubiquitous deep-crustal reflections is still debateable. However, the reflectors' large extent, lateral continuity, general smoothness, and relatively low dip in comparison with the complexity and variable dips of lithological contacts typical of high-grade domains suggests that the majority arise from relatively young tectonic surfaces such as faults and shear and mylonite zones that have suffered relatively little deformation subsequent to their formation.

Groundhog River block

The poorly exposed Groundhog River block forms a prominent positive magnetic anomaly (Fig. 3; MacLaren et al. 1968), but lacks significant gravity expression (Atekwana et al. 1994). Made up predominantly of granulite-facies mafic gneiss and tonalite (Leclair 1992) with 0.7 – 0.9 GPa (7 – 9 kbar) metamorphic pressures (Leclair 1990), the block is bounded by brittle faults to the east (Ivanhoe Lake) and west (Saganash Lake) and the ductile Wakusimi River fault to the southeast (Leclair et al. 1993). Ages of metamorphic zircon (2657, 2648 Ma) and titanite (2603 Ma) (Leclair and Sullivan 1991) are similar to those of the Chapleau block (Fig. 5), and suggest a similar sequence of metamorphic processes.

The crustal structure of the Groundhog River block has been examined through a variety of geophysical methods. Based on studies of rock properties and observed potential fields, the prominent magnetic high in the absence of a gravity anomaly can be modeled as a thin (2 – 4 km thick) wedge of moderately dense granulite with high magnetic susceptibility (Atekwana et al. 1994). Geometry involving a gently west-dipping Ivanhoe Lake fault and steeply west-dipping Saganash Lake fault (Percival and McGrath 1986) is consistent with seismic reflection profiles (Leclair et al. 1994) as well as with regional geobarometry (Leclair 1990). Magnetotelluric and University of Toronto electromagnetic system (UTEM) studies in the region indicate crust of normal (Abitibi-like) resistivity (Mareschal et al. 1994), supporting thin-sheet geometry for the granulites.

Val Rita block

West of the Saganash Lake fault, the Val Rita block contains gneisses typical of the Wawa gneiss domain. It is characterized by a positive gravity anomaly, apparently continuous with the Chapleau block anomaly to the south, located 40 km west of the Groundhog River block. Most interpretations of the geology and geophysics involve a crustal slab, dipping ~10°SE, with minor granulites exposed in the west and present at depths of ~10 km in the east, rotated on normal faults (Percival and McGrath 1986; Leclair 1990; Atekwana et al. 1994).

Fraserdale–Moosonee block

The northern part of the Kapuskasing structural zone, the Fraserdale–Moosonee block, is a <15 km wide region of granulite-facies paragneiss and mafic gneiss with 0.8–1.0 GPa (8–10 kbar) paleopressures (Percival and McGrath 1986), a coincident positive aeromagnetic anomaly, and a broad positive gravity anomaly. Across the brittle Foxville fault to the west, the Quetico belt consists of 0.5–0.6 GPa (5–6 kbar) granulite-facies paragneiss. The crustal structure is geometrically like a pop-up (Percival and McGrath 1986). Goodings and Brookfield (1992) interpreted the geometry as a flower structure in a transcurrent regime.

Post-Archean units

At least 10 swarms of postmetamorphic dykes cut the southern Superior Province (Halls et al. 1994) and two sets of alkalic rock – carbonatite complexes occur in spatial association with the Kapuskasing structural zone. Three of the swarms provide constraints on the uplift history of the Kapuskasing zone: the Matachewan, Biscotasing, and Kapuskasing dykes.

The large fan-shaped Matachewan swarm (Fig. 14) extends some 500 km northwest from a focal point near Lake Huron. South and east of the Kapuskasing zone it forms three subswarms (West and Ernst 1991), whereas to the northwest, two subswarms and a prominent, broad Z bend are evident. The dykes are absent from the Chapleau block, but occur within the Groundhog River block (Bates and Halls 1991*a*). Two antipodal paleomagnetic populations occur in dykes with identical (2454 ± 2 Ma, Heaman 1988) U–Pb ages, recording the earliest magnetic field reversal to date (Fig. 4*b*) (Halls 1991). Regional polarity domains (Halls and Bates 1990; Bates and Halls 1991*b*) can be attributed to later acquisition of normal polarity in dykes intruded into deep-crustal levels west of the Kapuskasing zone with respect to the reversed poles that characterize high-level dykes.

Characteristics such as cloudy feldspar and high-Al amphibole (Halls and Palmer 1990; Palmer and Barnett 1992) and high emplacement pressures (Percival et al. 1994) (Fig. 4*b*) indicate deep (12–15 km) injection levels for the normal-polarity dykes. A negative contact test (Halls et al. 1994) indicates elevated country rock temperatures at 2454 Ma, or pervasive alteration. Subsequent deformation involving both vertical and horizontal displacement is recorded by the swarm. Strain models include dextral transcurrent movement on northeast-trending faults, accompanied by minor southeast-directed thrusting (West and Ernst 1991) and east-southeast-directed shortening with associated east-northeast-trending dextral and west-northwest-trending sinistral faults (Halls et al. 1994).

The small Kapuskasing swarm of east-northeast-trending dykes occurs within and west of the Kapuskasing structural zone. The best estimate of its age is 2040 Ma, based on $^{40}Ar/^{39}Ar$ data from baked contacts (Hanes et al. 1988; Lee et al. 1990). Dykes with both olivine and quartz are considered part of the swarm, although it is possible that they represent different ages, based on their variable paleomagnetic character (Halls et al. 1994; Symons et al. 1994). Several features of the Kapuskasing dykes indicate emplacement at deep structural levels followed by uplift: (1) cloudy feldspar and high-Al amphibole (Palmer and Barnett 1992); (2) equilibration pressures as high as 0.57 GPa (5.7 kbar) (Percival et al. 1994); (3) negative contact tests, suggesting elevated country rock temperatures (Symons et al. 1994); (4) consistent southerly dips and brittle offsets of dykes near the Ivanhoe Lake fault zone (Percival 1981*b*). The Ar and paleomagnetic data appear to be inconsistent: if country rock temperatures were >400°C, then biotite in the country rock should not be reset adjacent to the dykes. The discrepancy might be explained if the Kapuskasing dykes studied in the independent investigations have different ages and emplacement depths, or if the temperature regime in which dykes with cloudy feldspar and high-Al amphiboles were produced is only ~300°C.

Biscotasing dykes (formerly known as Preissac) occur east of the Kapuskasing zone and trend east-northeast (Buchan et al. 1993). North-trending Onaping faults with some effect on the 1850 Ma Sudbury structure offset these 2167 Ma dykes (Buchan and Ernst 1994). The faults, which have sinistral horizontal displacements, extend far to the north, where they are themselves cut by northeast-trending dextral faults of the northern Kapuskasing zone.

In addition to constraints from the dykes, alkalic rock – carbonatite complexes provide information on the faulting history in the Kapuskasing region. The Cargill complex of 1885 Ma age (L.M. Heaman, personal communication, 1988) is cut and dextrally offset by the Lepage fault (Sandvik and Erdosh 1977), which probably originated as a normal fault (Percival and McGrath 1986) and was reactivated as a transcurrent structure (Buchan and Ernst 1994). Younger alkalic rock complexes of ~1.1 Ga age record only minor (<6 km) uplift and paleomagnetically undetectable horizontal translations (Symons et al. 1994). Associated lamprophyres (1144 Ma; Hanes et al. 1988) have minor offsets in the Ivanhoe Lake fault zone (Bursnall 1990; Bursnall et al. 1994), possibly resulting from differential isostatic adjustments to continued erosion.

Cooling and uplift history

Constraints on the cooling and uplift history of the Kapuskasing zone are provided by a variety of isotopic systems with a range of closure temperatures. However, interpretation of results is complicated. Radiometric ages resulting from static passage through a closure temperature must be distinguished from those reset, possibly below the closure temperature, in response to local thermal or alteration events.

Zircon retains radiogenic Pb to very high temperatures (~1000°C?, Heaman and Parrish 1991), making it the most accurate record of crystallization age; however, it does not behave well as a thermochronometer in most metamorphic terranes. Conversely, new zircon may grow at $T < 900°C$ and thus provide an estimate of the timing of metamorphism. Like zircon, titanite may grow below its closure temperature (~600°C, Heaman and Parrish 1991), and regional titanite age patterns can provide information on temperature history in

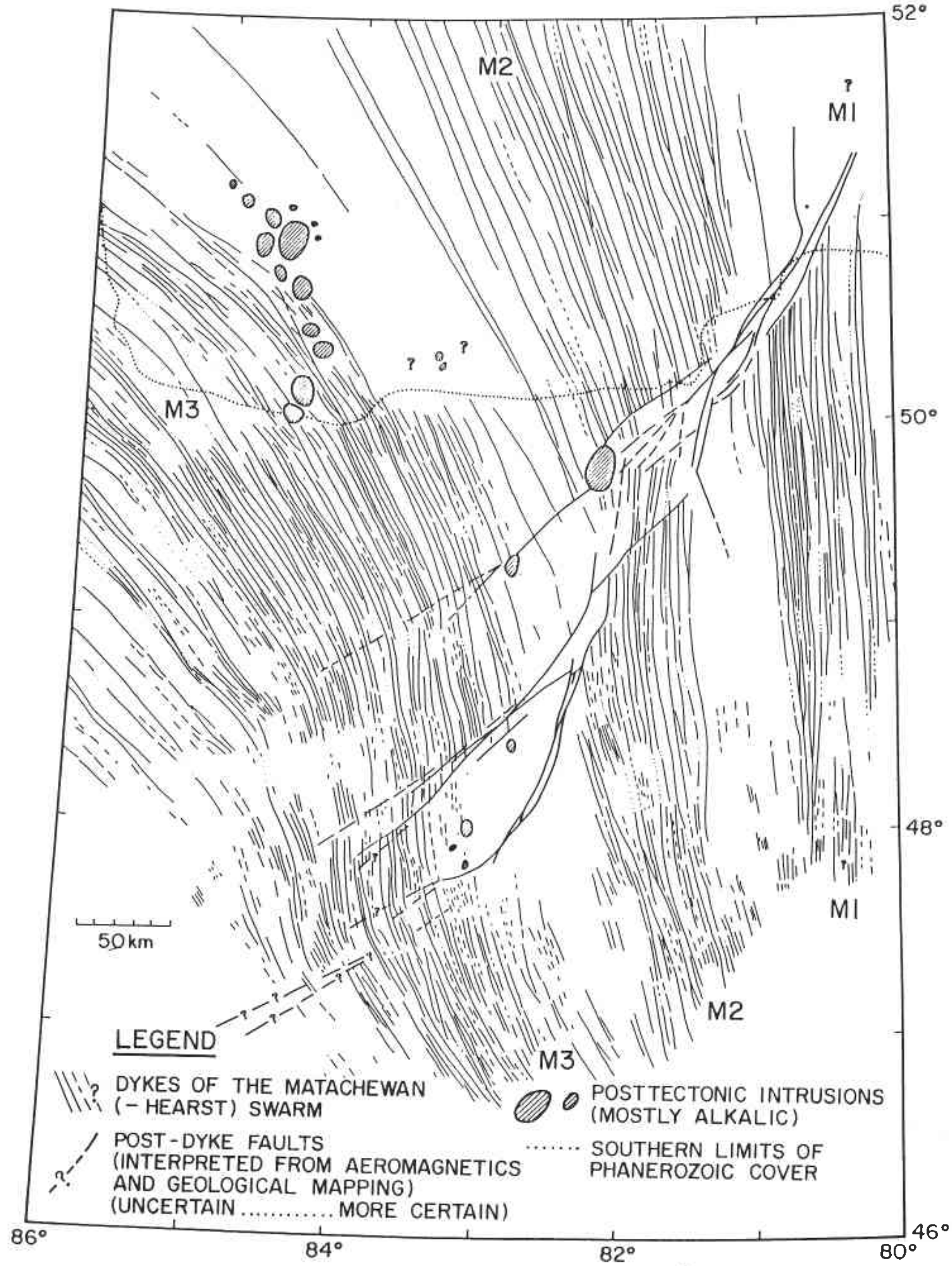

FIG. 14. Distribution of the Matachewan dyke swarm from geological maps and aeromagnetic anomaly patterns (after West and Ernst 1991).

the 600°C range. With closure temperatures of ~550°C, hornblende K–Ar and $^{40}Ar/^{39}Ar$ systems in granulite-facies rocks record cooling within the amphibolite facies. The timing of cooling through greenschist facies is recorded by biotite, with a closure temperature of ~280°C for Rb–Sr and 250°C for Ar (Heaman and Parrish 1991). A few fault rocks (cataclasite, pseudotachylite) were dated by $^{40}Ar/^{39}Ar$ with the hope that fusion might have reset the Ar clock. However, they gave complex spectra (Hanes et al. 1994). A brief summary of the data sets is provided here and followed by a discussion of the deduced complex history.

U–Pb titanite ages from the Kapuskasing zone and its surroundings range from 2685 to 2493 Ma, with systematic regional variation (Fig. 5). At high structural levels within and adjacent to the Michipicoten and Swayze greenstone belts, they are close to crystallization ages (2685–2650 Ma zircon dates, Fig. 5), reflecting rapid cooling (T.E. Krogh and J.A. Percival, unpublished data). Eastward through the Wawa gneiss domain ages decrease from 2650 to 2600 Ma, decreasing further to 2550 Ma through the amphibolite–granulite transition to the Kapuskasing structural zone. The youngest ages (2547–2493 Ma) are from the Chapleau block of the Kapuskasing zone, with a concentration of dates in the range 2514–2493 Ma at the deepest structural levels within a few kilometres of the Ivanhoe Lake fault zone. Ages within the ~200 Ma age spectrum correlate closely with the paleopressure determinations, which range from 0.3 to 1.1 GPa (3 to 11 kbar) (Mäder et al. 1994) (Fig. 4).

K–Ar and $^{40}Ar/^{39}Ar$ hornblende dates range from 2670 to 2400 Ma (Hanes et al. 1994) (Fig. 5). Within the Abitibi belt, hornblende ages are >2650 Ma, resembling titanite dates. In the eastern Wawa gneiss domain they are a little younger

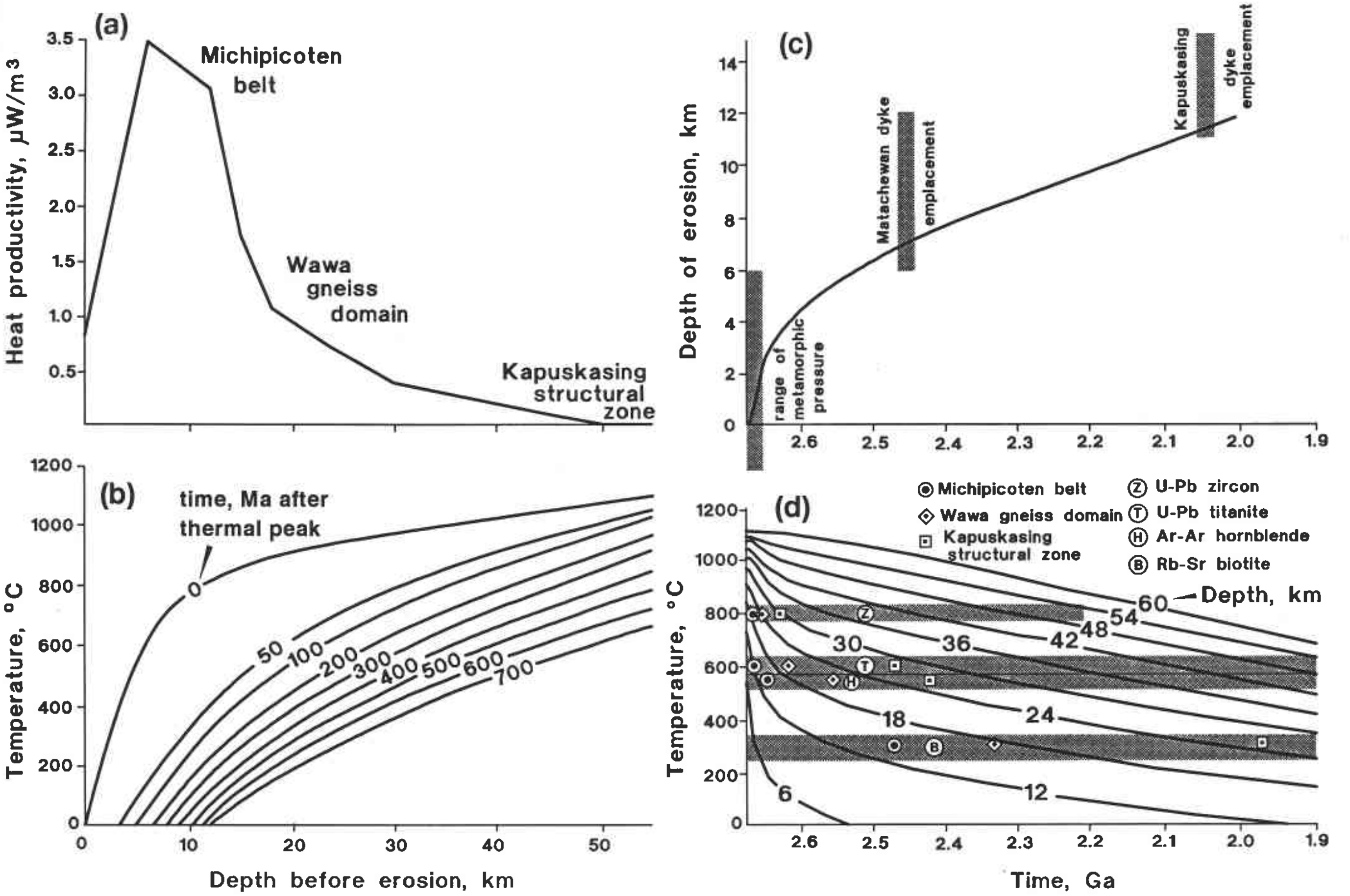

FIG. 15. One-dimensional thermal model for cooling in the Kapuskasing uplift prior to 1.9 Ga fitted to age data. Three crustal levels are represented in outcrop: (1) the Michipicoten belt (0–10 km depth), (2) Wawa gneiss domain (11–20 km), and (3) Kapuskasing structural zone (>21 km). (*a*) Assumed heat production values versus depth (present day) based on Ashwal et al. (1987). Values at 2670 Ma are approximately double. (*b*) Thermal evolution, showing decay of thermal anomaly following late Archean (2670 Ma) metamorphism and magmatism. (*c*) Assumed timing and magnitude of erosion for Kapuskasing granulites. Range of dyke emplacement pressures from Percival et al. (1994). (*d*) Model showing thermal evolution of Kapuskasing crust. Temperature–time data points (averages from the geochronological compilation of Fig. 5) can be fit to Superior Province-wide erosion of ~10 km betweem 2.67 and 2.1 Ga with no Kapuskasing-specific uplift event in this interval. (Note that the low-temperature ends of the cooling curves are represented by Rb/Sr biotite data rather than Ar–Ar results.) Present heat flow in the Kapuskasing area is 33 mW/m^2 (Cermak and Jessop 1971). Depths in (*d*) are depths before erosion.

(2640–2540 Ma). Ages in the Kapuskasing zone are in the range 2570–2370 Ma, except for a conventional K–Ar age of 2630 Ma from the northern Chapleau block. Dates in the range 2570–2500 Ma from the southeastern Chapleau block appear to violate the closure temperature heirarchy, in that they are older than titanite from the same area. However, ages are not from the same samples.

Biotite ages have been determined by both the Ar (Hanes et al. 1994) and Rb–Sr (Percival and Peterman 1994) methods, although not at the same sample sites. Closure temperatures for the two systems are similar, but the regional age patterns differ markedly. Argon ages range from 2554 to 1960 Ma, with complex regional distribution. The oldest ages (>2400 Ma) generally occur in the Wawa gneiss domain, but are also present, along with the youngest dates, adjacent to the Ivanhoe Lake fault zone. Dates in the 2300–2250 Ma range constitute age domains partly associated with fault zones. Hanes et al. (1994) interpreted the pattern to be the result of cooling of the entire region to below 250°C by 2450 Ma, with subsequent local resetting of biotite ages by fluid flow at 2300–2200 Ma.

A similar range of dates (2500–1930 Ma) is derived from the Rb–Sr biotite system, but the regional pattern differs. Ages decrease systematically southeastward toward the Ivanhoe Lake fault zone (Fig. 5), with the exception of ~2200 Ma ages west of the Budd Lake fault zone (Fig. 2) and a 1950 Ma age near the Saganash Lake fault (Percival and Peterman 1994). In contrast to the Ar-based interpretation, the Rb–Sr biotite pattern suggests late (post-1930 Ma) uplift of the crustal profile. A one-dimensional thermal model for the regional cooling history can be constructed using zircon, titanite, hornblende, and Rb–Sr biotite ages, representing the temperature range 750 to ~300°C at different crustal levels (Fig. 15). The data can be fit to a set of cooling curves that are partly a function of erosion magnitude over the interval 2.67–2 Ga. Cooling rates fit a model using a steady decrease of erosion rate to total ~10 km during the Early Proterozoic. The model is consistent with prolonged residence and slow cooling of granulites at >20 km depths prior to post-2-Ga uplift.

Interpretation of the thermochronometric data can best be addressed in terms of the geological history outlined below. Events proposed by various workers are weighed in light of complementary evidence.

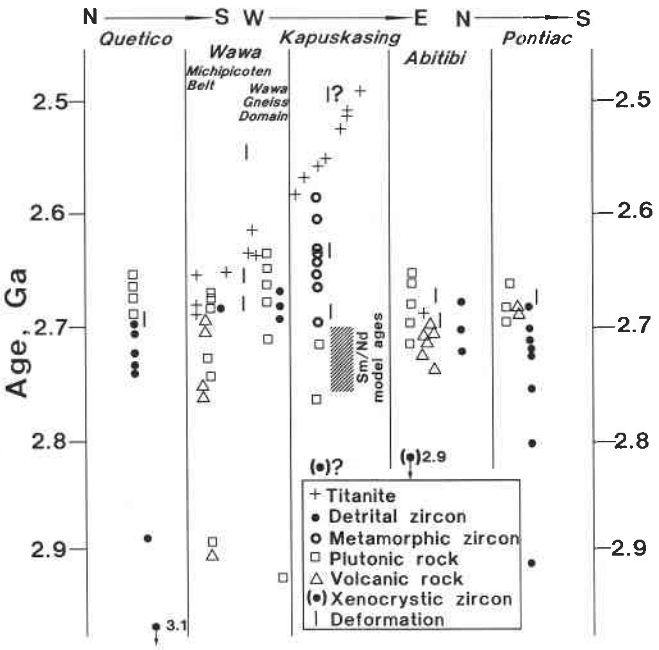

FIG. 16. Summary of U–Pb data for central Superior Province. Sources of information: Quetico: Percival and Sullivan (1988), Davis et al. (1990); Michipicoten: Turek et al. (1984, 1990, 1992), Krogh and Turek (1982), Corfu and Stott (1986), Corfu and Muir (1989), Jemielita et al. (1990), Corfu and Sage (1992); Wawa gneiss domain: Percival et al. (1981), Percival and Krogh (1983), Moser et al. (1991), Leclair and Sullivan (1991), Sullivan and Leclair (1992*a*, 1992*b*), Krogh (1993), Moser (1994); Kapuskasing zone: Percival and Krogh (1983), Corfu (1987), Krogh (1993), T.E. Krogh and J.A. Percival (unpublished titanite data); Abitibi: Frarey and Krogh (1986), Corfu and Grunsky (1987), Mortensen (1993), Heather and Van Breemen (1994), Corfu et al. (1992), Barrie and Davis (1990); Pontiac: Mortensen and Card (1993).

Geological history

The timing of major events in the geological evolution of the central Superior Province has been well defined by U–Pb geochronology (Fig. 16). Some are widely represented regionally and at a variety of crustal levels, whereas others are restricted spatially or occur only at specific levels. In the following section we describe depositional, plutonic, metamorphic, and structural events within the various lithotectonic elements of the central Superior Province.

Arc I (2920–2880 Ma)

This time period is represented by scattered occurrences of volcanic and plutonic rocks of the Wawa Subprovince in and near the Michipicoten belt. The Hawk Junction assemblage of basal komatiite overlain by dacite and iron formation, as well as associated 2888 Ma tonalite, is similar to 2920 Ma tonalite of the Wawa gneiss domain. Old structural elements have not been recognized in rocks of this vintage. Although poorly represented in outcrop, rocks in the 2.9 Ga age range may underlie significant areas of the geochemically evolved Michipicoten belt. The extent of the older rocks is not known. Jackson and Sutcliffe (1990) speculated that sialic rocks of this age extend east as far as the Kapuskasing zone; however, model Nd ages from the Kapuskasing zone (2.75–2.70 Ga) do not indicate the presence of older crust.

Arc II (2770–2696 Ma)

This wide time window encompasses much of the volcanism and plutonism of the Abitibi and Wawa subprovinces. It is likely that several tectonic settings and discrete periods are represented in the volcano-plutonic assemblages (Jackson and Fyon 1991).

In the Michipicoten belt two cycles of supracrustal rocks accompanied by synvolcanic intrusions were deposited in this time interval (2750–2696 Ma). Volcanic rocks from the Hemlo belt, in the 2770 Ma age range, appear to represent early components of this volcanic cycle, as do tonalitic rocks associated with the Shawmere anorthosite in the Kapuskasing uplift (2765 Ma). Geochemical studies of the anorthosite indicate an origin through fractionation of tholeiitic magmas (Simmons et al. 1980). Components of the Wawa gneiss domain, including tonalitic plutons and gneisses, belong to this group.

Volcanic and sedimentary rocks of the Abitibi belt have been grouped into lithological assemblages corresponding to a variety of tectonic settings ranging from ocean floor and oceanic plateau through arc and back-arc environments. Most volcanism occurred in the interval 2750–2700 Ma. Clastic sedimentary rocks, which overlie volcanic sequences conformably or unconformably, have been attributed to collisional processes (Williams et al. 1992).

Compressional tectonism; Quetico sediment deposition and deformation (2700–2689 Ma)

This period marks a significant regional deformational event in the Abitibi–Wawa belts and may have witnessed the accumulation of Quetico sediments to the north. In the greenstone–granite regions, this time span encompasses the termination of arc volcanism, a period of deformation related to north–south compression, and intrusion of the earliest posttectonic plutons. The same time period includes deposition and initial deformation of turbidites in the Quetico belt (Percival 1989; Williams 1991).

The oldest deformation recognized in the Michipicoten belt is expressed as folds and thrusts related to large-scale stratigraphic inversion. It occurred after 2680 Ma, the age of detrital zircons in the youngest sedimentary rocks involved, and before emplacement of posttectonic plutons of 2670 Ma age (Corfu and Sage 1992).

At the deeper structural levels represented by amphibolite-facies gneisses of the Wawa gneiss domain, both a pervasive gneissosity and F_2 folds predate emplacement of plutons at 2691 Ma (Bursnall et al. 1994). Metamorphism and migmatite production at this time are indicated by 2690 Ma leucosome. Metamorphic zircons of 2696 Ma age from 1.1 GPa (11 kbar) mafic gneiss in the northern Chapleau block indicate the presence of Kapuskasing gneisses at high temperatures and >30 km depths by this time.

At least two early structural events affected rocks of the Abitibi belt. Precleavage folds may have been associated with early thrust faults. The deformation may have predated volcanic rock sequences of 2.70 Ga age (Desrochers et al. 1993; Kimura et al. 1993). A later, widespread steep cleavage forms the major fabric element of the belt and is truncated by massive plutons of <2.69 Ga age. The depositional age of Quetico sedimentary rocks is bracketed by detrital zircons as young as 2691 Ma and truncating plutons of 2689 Ma. Deposition and deformation of the Quetico prism may be related to collision and uplift in the Wawa–Abitibi belt (Percival and Williams 1989; Davis et al. 1990).

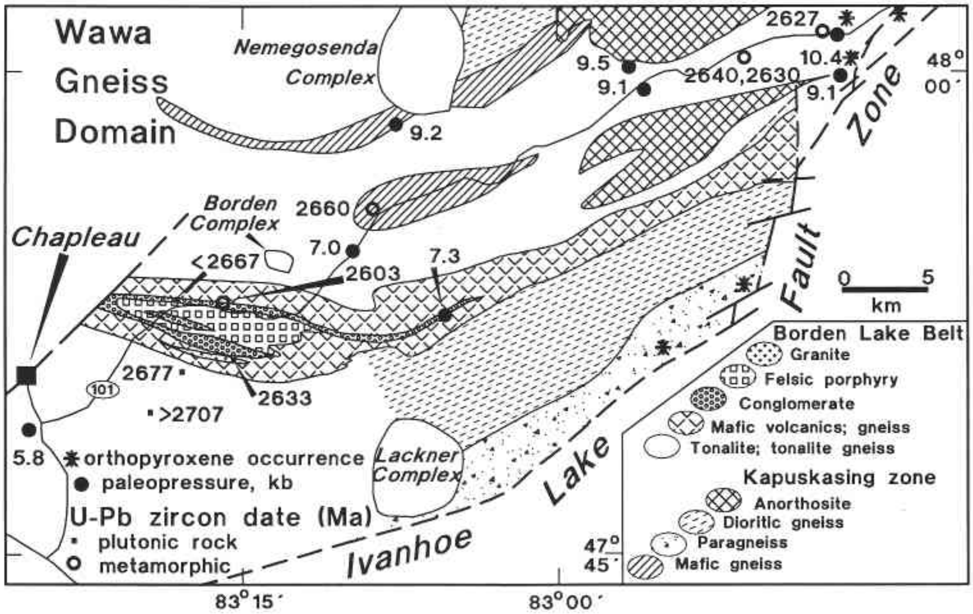

FIG. 17. Geology of the Borden Lake belt, with geochronological constraints (modified after Moser 1994; Bursnall et al. 1994).

Dextral transpression (2689–2684 Ma)

At the Wawa–Quetico subprovince boundary, a dextral transpressive (D_2) deformation event is bracketed between 2689 and 2684 Ma (Corfu and Stott 1986). Similar structures occur throughout the western Wawa Subprovince (Hudleston et al. 1988) and western Quetico belt (Borradaile et al. 1988; Bauer et al. 1992). Transpressional deformational effects have not been recognized in rocks from mid- to deep-crustal levels. Some plutons of the "posttectonic" suite in the Michipicoten and Abitibi belts occupy this time interval.

"Posttectonic" plutonism (2684–2665 Ma)

Throughout much of the Abitibi and Wawa belts, plutons and batholiths of massive granodiorite and granite are considered to postdate regional deformation. However, detailed studies indicate that tectonism younger than 2680 Ma, followed by 2670 Ma plutonism, marked the end of the deformation (Corfu 1993). This time period also includes major granitoid plutonism in the Quetico belt that postdated regional compressional deformation.

Timiskaming volcanism and sedimentation (2680–2677 Ma)

Lying unconformably on deformed, metamorphosed, and intruded supracrustal rocks of >2.7 Ga age are coarse clastic sedimentary and associated alkalic volcanic rocks of the Timiskaming association. Originally defined in the Abitibi belt, the association has also been recognized in parts of the Wawa Subprovince (Williams 1991). Rocks of Timiskaming type occur in elongate basins associated with faults and are commonly deformed and metamorphosed to greenschist facies. The depositional age in the type area is bracketed at 2680–2677 Ma. Consequently, "posttectonic" granitoid plutonism was apparently coincident with and outlasted Timiskaming activity. The extent of Timiskaming-age deformation at deep-crustal levels is not known.

Pontiac sediment deposition and deformation (2683–2675 Ma)

Bordering volcanic rocks of the Abitibi belt on the south are metasedimentary rocks of the Pontiac belt. Immature clastic sedimentary rocks and minor associated mafic volcanics constitute the Pontiac's supracrustal assemblage, which is cut by numerous suites of plutonic rock. The depositional age is bracketed between 2683 and 2675 Ma (Mortensen and Card 1993), close to that for sedimentary rocks of the Michipicoten belt (2680–2670 Ma), the Timiskaming association (2680–2677 Ma), and possibly the undated Timiskaming-style rocks of the Swayze belt.

Borden Lake conglomerate deposition (2667–2650 Ma)

The Borden Lake belt (Fig. 17), extending about 30 km from the Wawa gneiss domain into the Kapuskasing zone, may be the youngest supracrustal unit in the region. It contains metabasalt, conglomerate, and felsic porphyry, possibly of more than one age. The mafic rocks are cut by 2677 Ma granodiorite, whereas a depositional age of <2667 Ma has been inferred for conglomerate. The significance of the event that deposited, buried, and metamorphosed conglomerate after 2667 Ma is discussed further below.

D_{3-4} extensional deformation (2660–2637 Ma)

In the Wawa gneiss domain, subhorizontal D_4 shear zones associated with gently east- or west-plunging lineations overprint gneisses ranging in age from 2925 to 2660 Ma. Crosscutting pegmatites are 2645–2637 Ma, providing a bracket on the deformation of 2660–2645 Ma. Structures with similar geometry and relative age in the Kapuskasing structural zone appear to be younger still, in the 2640–2630 Ma range. These events are closely temporally related to some late plutonism at high structural levels and growth of zircon marking high-grade metamorphism at deep-crustal levels and thus may be part of a crustal-scale process. Convective removal of mantle lithosphere thickened by earlier compressive tectonism has both extensional strain and heating effects on the overlying crust (Platt and England 1994) and could have driven activity in this age range.

Deep-crustal metamorphism, late deformation, and gold mineralization (2665–2625 Ma)

Metamorphic zircons from high-grade mafic gneiss and migmatite leucosome within the Kapuskasing zone record a range of ages mainly between 2665 and 2625 Ma. Similar ages (2660–2637 Ma) are represented in leucosome and pegmatites in the Wawa gneiss domain. It is apparent that metamorphism at deep structural levels occurred through a protracted series of events, beginning as early as 2696 Ma and continuing at least until ca. 2580 Ma. Krogh (1993) related these events to ductile tectonic emplacement of oceanic slabs from a distant subduction zone. In this interpretation, the supracrustal rocks of the Kapuskasing zone are allochthonous and younger than the overlying Wawa gneiss domain. This theory seems to contradict the observation that the supracrustal rocks occur as inclusions in tonalite, which apparently extends into the Wawa gneiss domain. Also, although tectonic burial may be rapid, heating of large thick slabs of cold supracrustal rocks to granulite temperatures is a slow process, requiring several tens of million years (England and Richardson 1977).

Various authors have drawn attention to the fact that the young ages for metamorphic zircons (2660–2580 Ma) of the Kapuskasing structural zone overlap estimates of the time of gold mineralization (2630–2580 Ma) at high structural levels in the Abitibi belt (Fig. 18) (Colvine et al. 1988; Jemielita et al. 1990; Wong et al. 1991; Hanes et al. 1992; Krogh 1993). Models involve extraction of gold and lithophile elements from deep-crustal rocks during granulite-facies metamorphism, and subsequent upward transport along major conduits in the form of transcurrent fault zones such as the Porcupine–Destor and Kirkland – Larder Lake breaks.

The timing and magnitude of motion on these major gold-associated fault zones is not well defined. The latest major motion appears to have been dextral transcurrent (Robert 1989; Jackson and Fyon 1991), with only a minor dip-slip component (Powell et al. 1993). Faults affect Timiskaming sequences and therefore have a substantial displacement increment younger than 2677 Ma. Dates of hydrothermal minerals in shear-zone-hosted gold deposits of 2630–2580 Ma provide lower limits on the age of major movement.

Rotational uplift and burial in the Kapuskasing zone (2630 ± 5 Ma)

Combined U–Pb and K–Ar geochronology suggest the possibility of an event involving some uplift in parts of the Kapuskasing structural zone at ~2630 Ma. Mylonitic monzonite with a strong, southwest-plunging mineral lineation from the Abitibi belt near the eastern margin of the Kapuskasing zone (Fig. 5) contains concordant monazite of 2628 ± 2 Ma age, possibly dating the mylonite-forming event. Pegmatites cutting high-strain gneisses of the Kapuskasing zone have an identical zircon age (2629 Ma). Granitic sheets near Borden Lake have an approximate age of 2633 Ma, and orthopyroxene-bearing leucosome in paragneiss from the southeastern Chapleau block formed at 2627 Ma. These observations indicate the effects of deformation, metamorphism, and magmatism, but are not themselves diagnostic of an uplift event. However, in the northern Chapleau block, a K–Ar hornblende age of 2630 Ma suggests that in this area, cooling to below the hornblende closure temperature (~550°C) had occurred by this time. Regional deformation involving tilting at ~2630 Ma could have elevated the northern Chapleau block as the south was descending.

The oldest metamorphic zircon age (2696 Ma) also comes from the northern Chapleau block. The old argon age suggests that the cycle of granulite-facies metamorphism and cooling occurred earlier in rocks of the northern Chapleau block than elsewhere. Metamorphic zircon dates in the 2630 Ma range from farther south indicate active high-grade metamorphism at the same time that granulites in the north were cooling.

An event at 2630 Ma could account for burial and deformation of the Borden Lake belt. Conglomerate appears to have been incorporated late into the deep-crustal levels of the eastern Wawa gneiss domain and Kapuskasing structural zone. Structural analysis indicates two sets of structures in the Borden lake belt that are overgrown by garnet, inferred to be 2662 Ma in age. The age bracket 2667–2662 Ma would therefore include crystallization of the plutonic source of the youngest cobble, erosion of the pluton, deposition of the conglomerate, burial, as well as D_2 and D_3 structural events. Granulites with 0.9 GPa paleopressures (~30 km burial depths) in adjacent areas were undergoing high-grade metamorphism at 2663–2662 Ma (Fig. 17). It is unlikely that the protoliths of the granulites are the same age as the Borden Lake conglomerate, because even if thrust very rapidly to 30 km levels, they are too extensive to have had time (<4 Ma) to be heated to temperatures >700°C. Accomplishment of these events and heating to upper amphibolite facies in the allotted (5 Ma) time is equally difficult to envision for the Borden Lake conglomerate. The eastern end of the belt contains regional D_4 (extensional) structures with a probable age of ~2630 Ma. Therefore, the amphibolite-facies metamorphism in the Borden Lake area could have postdated the 2663 Ma granulite-forming event. As regional high-grade conditions persisted at least until 2625 Ma, a slightly younger metamorphic age for them is possible.

Several models have been proposed to explain the presence of the Borden Lake conglomerate, including an origin as part of a tectonic underplate (Krogh 1993), a tectonically buried Timiskaming-type conglomerate of the Swayze belt (Moser 1994), and a tectonic sliver of Timiskaming-type conglomerate ingested along an east–west transcurrent fault zone (Leclair et al. 1993). Given the <2667 Ma age of the unit and evidence for continued high-grade metamorphism in the Kapuskasing zone during the interval 2665–2625 Ma, the most likely explanation is probably the last.

High-temperature cooling (2680–2493 Ma)

Titanite U–Pb ages have traditionally been used to determine cooling to below the blocking temperature in the 600°C range. In high-level rocks of the western Wawa gneiss domain and western Abitibi belt, titanite ages in the 2680 Ma range indicate early cooling to this level. Ages decrease systematically, as morphological complexity and age variability increase in rocks at deeper structural levels, to a minimum of 2493 Ma adjacent to the Ivanhoe Lake fault zone. The spatial distribution of titanite ages illustrates the later effects of metamorphism and maintenance of comparatively high temperatures at the deepest exposure levels until at least 2493 Ma.

The nature of late deep-crustal tectonism: downward-migrating ductility front

At high structural levels represented by the Michipicoten and Abitibi greenstone belts, regional deformation appears to have ceased by 2670 Ma, the age of posttectonic plutons. Conversely, deformation and high-grade metamorphism continued for some time at deeper crustal levels. Regional ductile east–west extensional structures, bracketed between 2660 and 2645 Ma, occur in the Wawa gneiss domain and geometrically similar structures with younger (2640–2630 Ma) ages characterize the Kapuskasing structural zone. The structural-

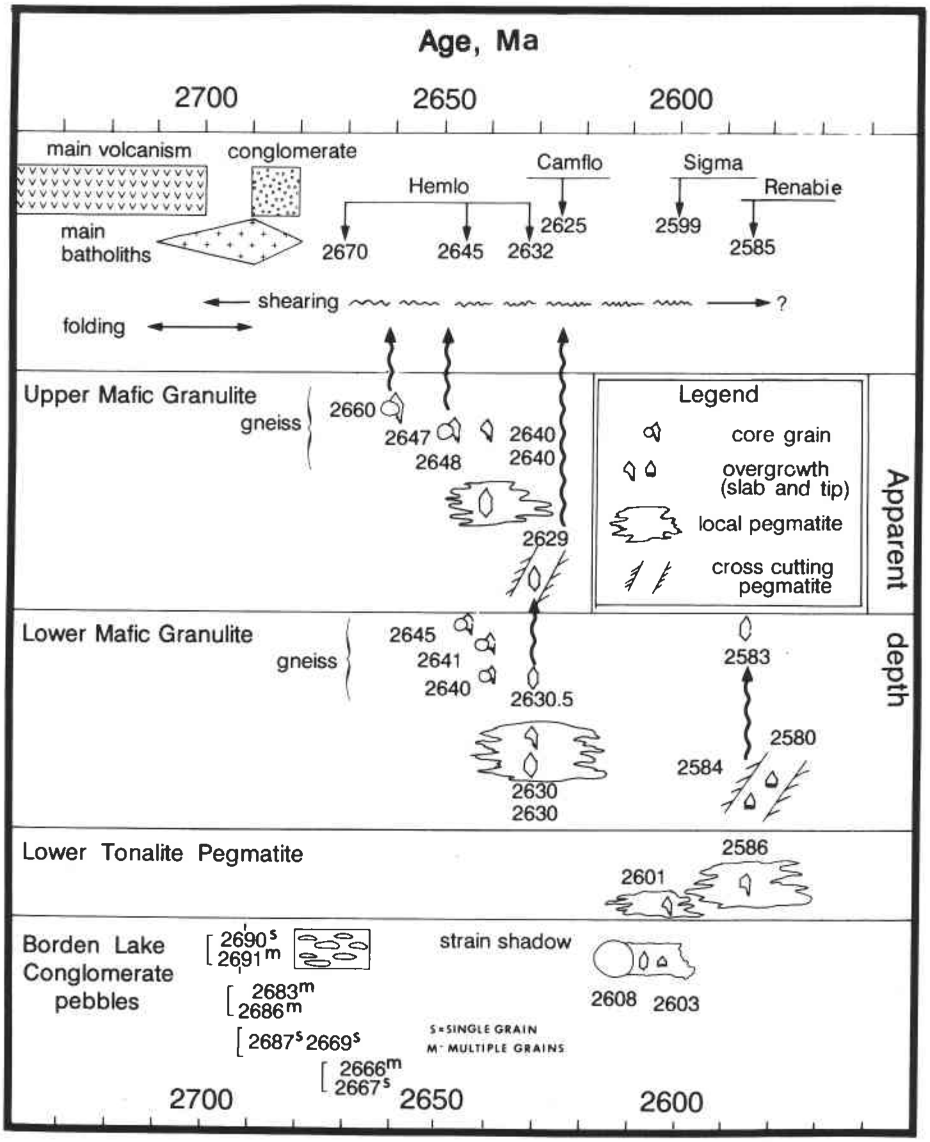

FIG. 18. Summary diagram showing the distribution of U–Pb zircon ages with apparent depth in the Kapuskasing structural zone and their relationship to the age of high-level gold deposits and the Borden Lake conglomerate (after Krogh 1993).

age observations may be reconciled by considering the mechanical behaviour of a collapsing orogen (e.g., Dewey 1988; Sandiford 1989). As compressional deformation wanes, new penetrative structures will form in ductile (deep) structural levels, as the upper crust adjusts through brittle faulting. Penetrative deformation is accomplished through grain-scale recrystallization, promoting, among other things, the growth of new zircon. Although such structures are imposed on rocks throughout the ductile zone, the deeper regions are susceptible to overprinting by subsequent strain. Only rocks in a "window of preservation" immediately below the brittle–ductile transition have potential for formation and retention of structures of a particular age. Therefore, deeper crustal levels may have geometrically similar structures that were preserved at progressively younger times. Ductile conditions imply that high temperatures (granulite-facies conditions) were maintained in the deep crust for extended periods. Application of this phenomenon to the Kapuskasing uplift provides an alternative to the underplating model (Krogh 1993) for explaining the observed "younging-with-depth" relationship. This mechanism may also account for the predominance of subhorizontal reflective zones in the deep crust, formed and retained progressively during late orogenic stages.

Postmetamorphic uplift occurred along a family of large northeast-striking structures including the Miniss and Gravel River faults, which Williams et al. (1992) related to the Kapuskasing zone. The faults were inferred to be an effect of tectonic indentation late during subprovince accretion history and therefore of Archean age. Other authors have suggested an Early Proterozoic age for this family of structures (West and Ernst 1991; Zhang and Halls 1992; Halls et al. 1994).

Uplift 1 (2500–2454 Ma)

Constraints on the later Archean thermal history are provided by U–Pb dates on a variety of accessory minerals. The growth or resetting of titanite as young as 2493 Ma suggests that the Kapuskasing rocks (except in the northern Chapleau block) remained at relatively high temperature (>600°C?) and, by inference, at significant depth, at least until this time.

Paleomagnetic data from Archean rocks in the region provide constraints on the uplift and displacement history in the interval ~2665–2450 Ma, during cooling of the rocks to their Curie point. For the oldest units analyzed, in the 2665 Ma range, there is no statistical difference between paleomagnetic poles for the high-level Abitibi and Wawa subprovinces, indicating no significant (>10°) relative movement since ca. 2650 Ma (Symons and Vandall 1990; Symons et al. 1994). High-grade rocks from the eastern Wawa gneiss domain and Kapuskasing zone are believed to have acquired their magnetization at ca. 2550 and 2430 Ma respectively, based on $^{40}Ar/^{39}Ar$ hornblende ages (Lopez Martinez and York 1990). Costanzo-Alvarez and Dunlop (1988) reported a systematic difference between the pole position for the eastern Wawa gneiss domain and the central Kapuskasing zone, which they attributed to differential tilting. More recently Symons and Vandall (1990) and Symons et al. (1994) assessed no statistical difference between the two areas. Initial cooling and remanence acquisition occurred while the gneisses were still deep in the crust; subsequent tilting is inferred to have taken place prior to emplacement of the Matachewan dyke swarm. This conclusion appears to conflict with the Matachewan failed contact test (Halls et al. 1994), which suggests that magnetization was not retained in country rocks until some time after intrusion of the Matachewan dykes. Also, with the exception of the Matachewan dykes, pole positions on the 2600–2450 Ma apparent polar wander path are not well dated. Therefore considerable interpretational flexibility exists.

Combining the evidence from the U–Pb geochronology for slow cooling until ~2500 Ma with that from paleomagnetics for tilting prior to Matachewan dyke emplacement at 2454 Ma and $^{40}Ar/^{39}Ar$ hornblende and biotite ages as old as 2500–2450 Ma leads to the inference that some uplift occurred between ca. 2500 and 2454 Ma. It is unclear whether this event was Superior Province wide, or specific to the Kapuskasing area.

Circumstantial evidence from geochronology and inferences from paleomagnetic studies point to uplift at 2500–2454 Ma; however, its nature and cause are obscure. It seems unlikely that compressional tectonics could account for uplift at 2500–2454 Ma given the nature of tectonism in adjacent areas at this time. The Superior Province margin, 200 km south of the Kapuskasing zone, was undergoing rifting at this time, based on the nature of Huronian sedimentary sequences and associated mafic magmatism of 2491–2480 Ma age (Krogh et al. 1984). Associated felsic volcanic rocks, the Copper Cliff rhyolite, dated at 2450 ± 10 Ma, show that the magmatism extended into the rifting phase represented by intrusion of the Matachewan dyke swarm (2454 Ma). It is therefore difficult to explain convergent deformation at 2500–2454 Ma. Conversely, the formation of positive topography and consequent erosion at this time could have provided a source of quartz-rich sediments deposited in the thick Huronian sedimentary succession (e.g., Roscoe and Card 1992). Broad extensional uplift and heating could have occurred in the southern Superior Province as a consequence of rifting.

Isotopic resetting (2300–2250 Ma)

Evidence for this event appears in the form of 2.30–2.25 Ga Ar–Ar dates on biotite and muscovite from wide zones adjacent to the Ivanhoe Lake and Budd Lake faults (Hanes et al. 1994). The resetting is interpreted to result from hydrothermal alteration along faults or block uplift at this time. Paleomagnetic pole positions measured in the zone of reset Ar dates correspond to well-dated poles of 2.16–1.88 Ga age (Halls et al. 1994). A possible reconciliation lies in the suggestion that the paleomagnetic and Ar systems were set at different times by chemical rather than thermal effects.

Alkalic rock – carbonatite complex intrusion I (ca. 1885 Ma)

At least two alkalic-rock bodies of ~1.88 Ga occur in association with the Kapuskasing zone (Sage 1991). The spatial relationship suggests that the bodies were emplaced along preexisting structures and therefore that the Kapuskasing structure existed by this time. The Cargill complex has a U–Pb zircon age of 1885 Ma (L.M. Heaman, personal communication, 1985) and the Borden complex a less precise Pb–Pb isochron age of 1872 Ma (Bell et al. 1987). The structural control on emplacement may have been fault zones on a local scale, judging by position of the Cargill complex on the Lepage fault, but on a larger scale, the anomalously thick crust beneath the Kapuskasing uplift may have constituted a zone of lithospheric weakness.

Uplift 2 (1950–1850 Ma)

Several lines of evidence support an Early Proterozoic deformation event at 1950–1850 Ma. The Matachewan dyke swarm is deformed (Fig. 14), although direct observation of the relative ages of the dykes and Ivanhoe Lake fault zone is prevented by inopportune dyke distribution. Evidence for distortion includes systematic nonvertical dips (Card et al. 1981; Ernst and Halls 1984), regional variations in magnetization direction, paleomagnetic polarity domains (Halls and Bates 1990; Bates and Halls 1991*b*), as well as subswarm offset and broad warps (West and Ernst 1991). The deformation can be modelled as a broad ductile and ductile–brittle transcurrent deformation zone on the western flank of the Kapuskasing uplift that accommodated up to 70 km of dextral offset, or distributed northwest–southeast shortening (Halls et al. 1994). The deformation also involved some block tilting and uplift (Bates and Halls 1991*b*), and may have occurred at ~1950–1890 Ma, based on analysis of geochronological and paleomagnetic data. The Matachewan deformation can be inferred to be <2167 Ma (Buchan and Ernst 1994) and the Kapuskasing dykes (2.04 Ga) are locally offset by discrete faults in the Ivanhoe Lake zone. Paleomagnetic studies of the Kapuskasing dykes support a tilt of ~30° in a 15 km wide zone northwest of the Ivanhoe Lake fault zone (Symons et al. 1994). The dykes also contain evidence of emplacement pressures up to 0.57 GPa (5.7 kbar). Fault structures in the Ivanhoe Lake zone include protomylonite, cataclasite, pseudotachylite, and later microfaults formed in the brittle regime, probably during the Proterozoic. Biotite Rb–Sr dates young progressively southeastward to 1930 Ma in the Kapuskasing structural zone, implying that temperatures in the deep crust remained above ~280°C until that time. Additional geochronological evidence of activity at this time includes a whole-rock Ar–Ar date from the Ivanhoe Lake fault zone of 1947 Ma (Hanes et al. 1994), a Pb isochron on plagioclase from Matachewan dykes of 1960 Ma (Smith et al. 1992), and a whole-rock Rb–Sr age on pseudotachylite in the western Superior Province of 1950 Ma (Peterman and Day 1989). The lower age bracket for the deformation is uncertain; however, the Cargill complex (1885 Ma) is offset only slightly.

Normal faults in the central and northern Kapuskasing uplift are interpreted as collapse structures formed in response to thrust thickening (Percival and McGrath 1986). Based on aeromagnetic images, Halls et al. (1994) inferred a chronology for Early Proterozoic deformation: (1) southeast-verging thrusting, (2) west-side-down normal faulting, (3) sinistral transcurrent faulting along north- and north-northwest-trending structures, (4) dextral faulting along north-northeast–east-northeast-trending zones.

Deformation at 1900–1850 Ma could relate to Early Proterozoic orogenesis at the margins of the Superior Province (e.g., Gibb 1983), possibly in the Trans-Hudson orogen (Halls et al. 1994) or Penokean orogen (Percival and Peterman 1994). Whereas collisions with the Superior margin in the Trans-Hudson orogen were in the range 1820–1750 Ma (Hoffman 1989; Bleeker 1990), those in the Penokean are dated at 1860 and 1840 Ma (Sims et al. 1989).

Hydrothermal activity (1900–1700 Ma)

Hanes et al. (1994) reported a few Ar–Ar dates in this range and attributed them to minor hydrothermal activity, perhaps related to Hudsonian or Penokean activity.

Alkalic rock – carbonatite complex intrusion II (1145–1015 Ma)

At least three major bodies and associated lamprophyres occur in association with the Kapuskasing structure (Sage 1991). The Lackner Lake, Shenango Township, and Nemegosenda Lake complexes have respective Rb–Sr isochron ages of 1092, 1047, and 1015 Ma (Sage 1991), whereas lamprophyres have $^{40}Ar/^{39}Ar$ ages of ~1145 Ma (Hanes et al. 1988). Hydrothermal alteration and thermochemical paleomagnetic overprints probably relate to fluids associated with the magmatism (Costanzo-Alvarez et al. 1993). As with the ~1.88 Ga complexes, these bodies are variably associated with faults, suggesting larger scale control on their emplacement, possibly the Kapuskasing crustal root.

Late faulting (1100 Ma)

Minor shear zones cut lamprophyre dykes dated at 1145 Ma, while pseudotachylite and K-feldspars give ages in the 1100 Ma age range (Hanes et al. 1994). These effects could be related to alkalic rock intrusive activity near the Kapuskasing zone, or to distant structural effects of the Grenville orogeny.

Discussion

Existing models for the evolution of the Kapuskasing structural zone have generally attempted to explain single aspects of the geology or geophysics, and fail to provide comprehensive understanding of the region, particularly in light of new data acquired in the past decade. Because the structure is complex, having evolved over the period 2.8–1.1 Ga, and varies considerably along strike, no simple model is capable of accounting for all of the salient features. However, elements of most of the proposed models do enter the comprehensive evolutionary picture developed below.

One model, proposing a suture along the Kapuskasing structure between the unrelated eastern and western Superior Province, cannot be maintained in light of subsequent mapping and geochronology that indicate correlativity of the Abitibi and Wawa subprovinces at least since 2750 Ma.

The horst model succeeds in describing the surface geometry of the Chapleau, Groundhog River, and Fraserdale–Moosonee blocks, in that deep-crustal rocks are exposed in a linear structure; however, the structure in the third dimension is more consistent with a compressional origin than with extension. Furthermore, geochronological evidence suggests that the Kapuskasing structure existed prior to 1100 Ma, the proposed age of horst formation and associated emplacement of alkalic rock – carbonatite complexes.

A family of ductile sinistral transcurrent fault models has been proposed to account for northeasterly aeromagnetic and map trends in the Kapuskasing structure (Watson 1980; Goodings and Brookfield 1992; Williams et al. 1992, p. 1288). Other explanations for this phenomenon are possible, and based on unit deflection, the magnitude of any transcurrent displacement appears to have been minor (<20 km).

A model of Proterozoic dextral transcurrent shear accounts for the deformation pattern of the Matachewan dyke swarm west of the Kapuskasing structural zone (West and Ernst 1991), but does not accommodate the magnitude of uplift required, nor the variable along-strike geometry of blocks of the Kapuskasing zone.

The much-cited model of an east-verging Early Proterozoic thrust fault (Percival and Card 1983) explains many features of the Chapleau block, and associated normal faults account for the geometry of the Groundhog River and Val Rita blocks (Percival and McGrath 1986). However, the magnitude of post-Matachewan-dyke west-over-east displacement (55–70 km; Percival et al. 1989; Geis et al. 1990) does not seem possible given the semicontinuous nature of the Matachewan dyke swarm across the southern end of the Chapleau block.

All of the existing models are deficient to some extent in explaining the following elements of the Kapuskasing structure:
(1) the Borden Lake belt contains rocks younger than 2667 Ma that were incorporated late into the deep crust;
(2) the Chapleau, Groundhog River, Val Rita, and Fraserdale–Moosonee blocks are bounded by at least three sets of brittle faults and each has a distinct three-dimensional geometry;
(3) granulite-facies rocks occur in all blocks and indicate pressures of 0.8–1.1 GPa (8–11 kbar), requiring 25–30 km of uplift;
(4) the slow cooling history of the deep-crustal rocks indicates a protracted period of lower crustal residence following metamorphism;
(5) a broad zone of thick crust is associated with the Kapuskasing zone, and mid-crustal refraction velocities approach the surface beneath the Chapleau block. A positive gravity anomaly generally coincides with the distribution of high-grade rocks at surface;
(6) seismic reflection sections generally show truncations but no reflectivity marking the Ivanhoe Lake fault zone; prominent near-surface reflections in the Kapuskasing zone relate to contrasting lithological packages and appear to image Archean structure; prominent deep reflections within the Abitibi belt dip moderately northwest beneath the Chapleau block;
(7) the Matachewan dyke swarm is deformed west of the Kapuskasing zone; it is semicontinuous across the Ivanhoe Lake fault zone but deeper dyke levels are indicated in the Kapuskasing zone by mineral compositions and polarity domains.

These elements can be summarized in the following model for the late Archean and Early Proterozoic evolution of the Kapuskasing structural zone. Supracrustal rocks of the Kapuskasing zone (2750–2700 Ma) were buried by younger supracrustal rocks, tectonic shortening, and intrusion of mid-crustal tonalites (2700–2660 Ma). Metamorphism began during this period and continued in response to magmatic heat and crustal collapse (2660–2625 Ma). The Borden Lake conglomerate was deposited in a pull-apart basin and transported into the hot

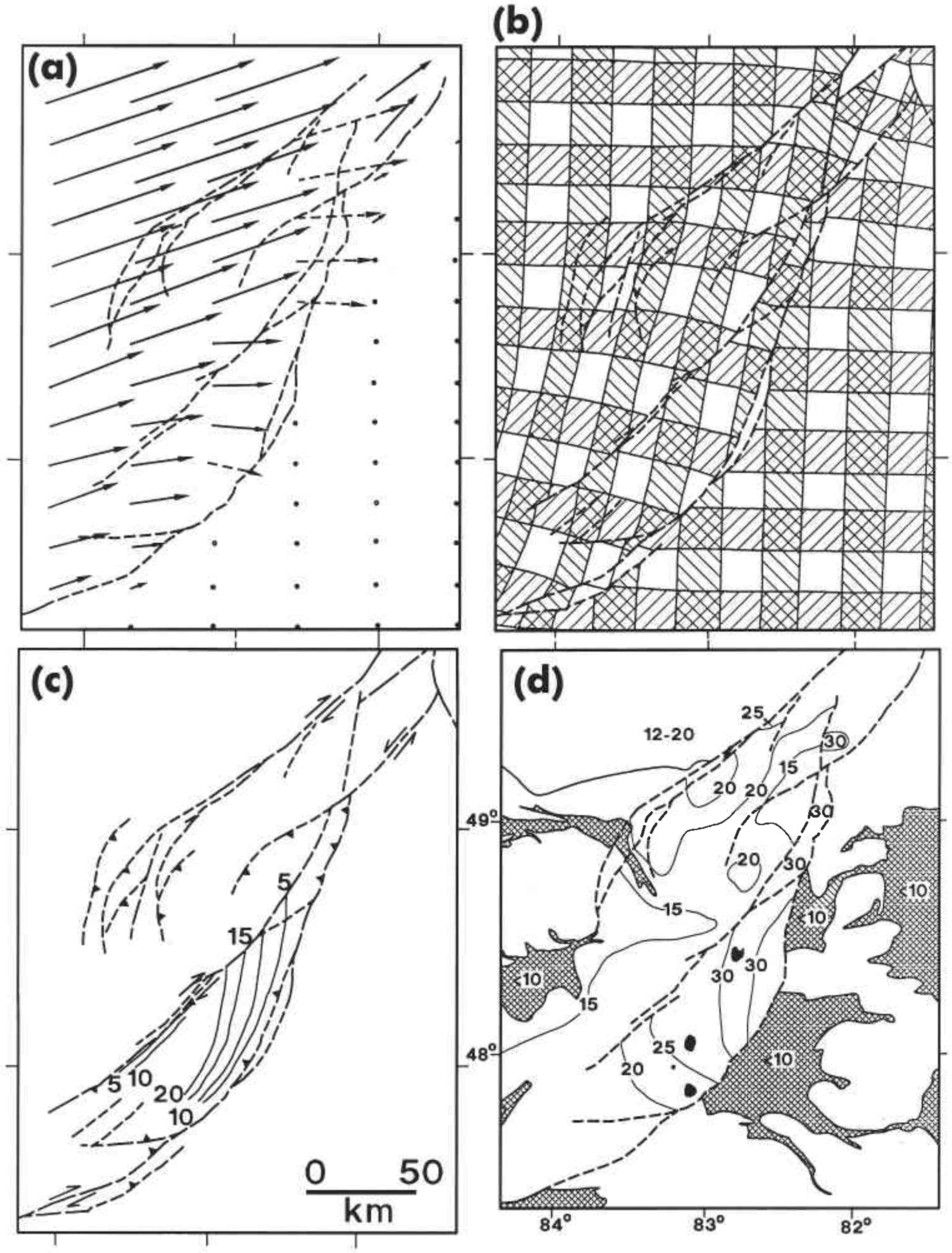

FIG. 19. Maps showing interpreted horizontal and vertical displacements in the Kapuskasing region. (*a*) Displacement vectors showing horizontal motion of northwestern block with respect to southeastern based on analysis of strain in the Matachewan dyke swarm. Arrow tail is pre-deformation position and head indicates present position. Dashed vectors are for material removed by erosion. (*b*) Two-dimensional strain map showing offset and rotation of the northwestern block with respect to the southeastern. Grid cells (undeformed) are 20 km squares. The map is drawn to accommodate 50 km of right-lateral offset and 30 km of shortening across the Ivanhoe Lake fault zone and associated structures. (*c*) Interpreted sense of movement on brittle Proterozoic faults (broken lines). Numbers on contours (solid lines) indicate depth (km) to the Ivanhoe Lake fault zone based on interpreted dip from reflection profiles. (*d*) Paleodepth contours (km) for the Kapuskasing region based on geobarometry (after Mäder et al.; Leclair et al. 1994). The map shows total amount of vertical uplift since ca. 2.6 Ga.

deep crust as a sliver along a downward vector on a dominantly transcurrent fault. Kapuskasing rocks remained at depth, where they cooled slowly and underwent intermittent minor deformation (2625–2585 Ma). The Superior Province was eroded by 10 km on average, elevating Kapuskasing levels from ~30 to ~20 km. Incipient breakup of the Superior craton (2500–2450 Ma) was recorded as new mineral growth at deep structural levels and by Matachewan dyke injection. A few additional kilometres of erosion preceded intrusion of Kapuskasing dykes (2040 Ma) into the intact crustal section. At ~1.9 Ga, when Kapuskasing levels had cooled to <300°C, stresses caused by plate collisions at the Superior margin were transmitted into the interior in the form of dextral transpression along northeast-trending faults and northwest-over-southeast thrusting, elevating Kapuskasing-level rocks and associated mid-crustal refraction velocities from ~18 to 3 km depths along the ~35°NW-dipping Ivanhoe Lake fault zone. Approximately 27 km of shortening is implied (Fig. 19). Some ductile(?) thrust move-

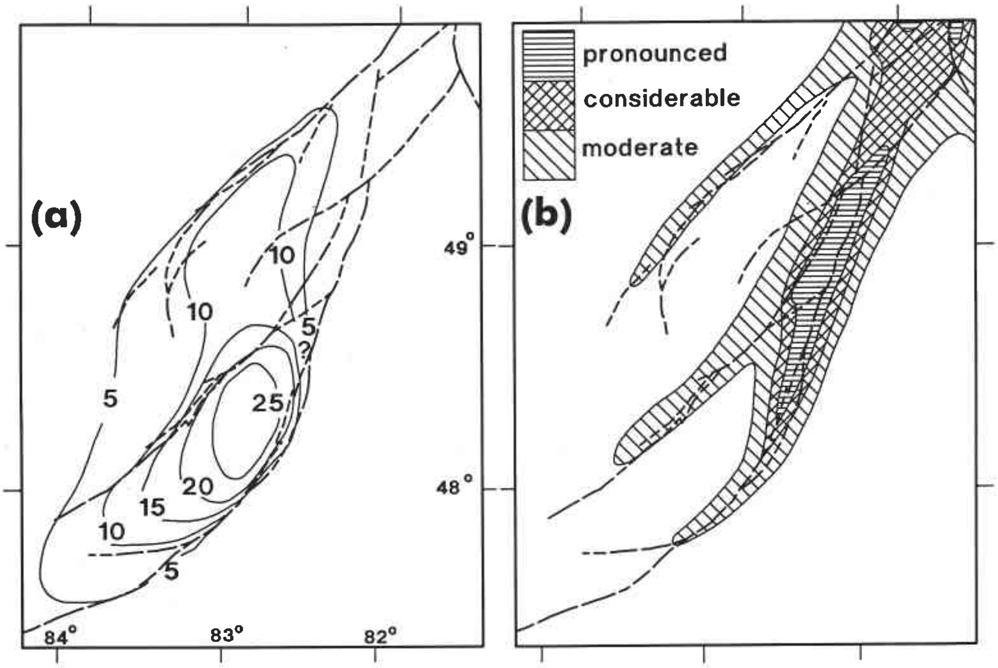

FIG. 20. Maps of distribution of ductile and brittle crustal thickening in the Kapuskasing region. (*a*) Amount of ductile lower crustal thickening (km) based on erosion magnitude and refraction seismic estimate of Moho depression. (*b*) Qualitative estimates of brittle upper crustal thickening by thrust duplication.

ment may have occurred on the parallel structures in the Abitibi footwall as well as on associated thrusts to the west (Zhang and Halls 1992). Because the main faulting occurred at relatively low temperature (<300°C at 20 km levels), the dominant fault mechanism was brittle, consistent with the dominantly cataclastic Ivanhoe Lake fault zone. The effect of brittle deformation and brecciation on the reflection character was to destroy existing reflectors rather than to produce layered mylonitic structures and new reflectivity. Formation of a crustal root accommodated shortening in the lower crust; its preservation was permitted by Moho temperatures in the 600–700°C range (~2.5–2.0 Ga, Parphenuk et al. 1994). Northwest-dipping normal faults and a conjugate set of strike-slip faults broke the Kapuskasing structure into separate blocks with variable geometry. During the relaxation phase, isostatic rebound reduced topography on the root and probably produced a few more kilometres of uplift along steep structures not coincident with the Ivanhoe Lake fault.

Geotectonic interpretation

The picture of horizontal and vertical displacements and deformation shown in Fig. 19 is initially puzzling. Horizontal components of thrust displacement across the Ivanhoe Lake fault zone appear nowhere to exceed 30 km, but vertical uplifts of more than 15 km are widespread. Qualitatively, the pattern of motions and deformation is in accord with the two-dimensional geodynamic model of Parphenuk et al. (1994), which considers deformation resulting from convergence of two lithospheric blocks. In that model, thrust faults in the upper crust are accompanied by growth of a broad crustal root zone beneath (Percival and Green 1989; Percival et al. 1989), similar to the structure inferred for the active Zagros continental collision zone (Snider and Barazangi 1986). Convergence of the blocks is accommodated in the brittle upper crust by thrust duplication and erosion, and in the high-viscosity upper mantle by a broad zone of mainly downward flow. The overall thickening of the crust and accompanying surface uplift results from the inevitable accumulation of dense, lower crustal material between the upper crust and mantle. The effective viscosities of the various layers control the style of deformation. Due to the difference in composition, and despite the difference in temperature, the lower crust has an effective viscosity much less than that of the adjacent upper mantle; both viscosities are significantly lower than that of the cool brittle upper crust. Although the basic geodynamic model of Parphenuk et al. is two dimensional and the Kapuskasing structure has lateral deformation components, the concept of ductile lower crustal flow coupled with upper crustal thrusting remains valid.

Figure 20*a* shows regions in which the crust appears largely to have been thickened by lower crustal flow and Fig. 20*b*, those regions in which there is evidence for upper crustal thickening by thrust duplication. Where the latter is negligible, one can estimate the total amount of lower crustal thickening by assuming that the pre-deformation crust was of uniform thickness and that all but 10 km of the exhumation deduced from the metamorphic pressure estimates is due to the deformation event. In regions of thrust duplication only total thickening can be estimated. Beneath these regions, the ductile lower crust may have thinned, or at least have been less inflated than in other areas. The total volume of lower crustal thickening suggested by Fig. 20*a* is not excessive in light of the observation from Fig. 19 of interblock displacement of 30 km of convergence and 50 km of dextral shear. A converging crust that included 5–10 km of ductile dense lower crust could provide this amount of material.

The presence of a widepread low-viscosity lower crustal layer would sustain local isostasy. Thus, a 1 km increment in lower crustal thickness due to lateral inflow of dense material would result initially in ~0.1 km of uplift and 0.9 km of Moho depression. The uplift would be continually renewed after erosion until the total amount removed reached a proportion of the total increment equal to the ratio of the density contrasts between mantle and injected material, and mantle and eroded materials. Thus, an estimate of 20 km average thickening along with an average 5 km of Moho depression under the Chapleau block, and density estimates of 2700, 3000, and 3500 kg/m^3 for, respectively, granitoid upper crust, lower crust, and upper mantle, would imply that about 10 km of granitoid upper crust and 10 km of dense lower crust were removed in response to the thickening.

The shape inferred for the lower crustal root zone suggests that lateral flow of ductile material at the base of the crust may be affected by local aspects of the crustal structure. Fault zones in the brittle crust may have constituted obstacles to flow, perhaps due to local thickening by duplication. Alternatively, the correlation in patterns may have arisen from the time sequence of events.

A test of this geotectonic interpretation is available in seismic reflection data, where most of the reflectivity appears to arise from late Archean structures. Regions of brittle crust that were uplifted as rigid blocks should retain their reflectivity, whereas regions deformed ductily should show little reflectivity. This pattern is apparent from comparisons of data from the Abitibi and Kapuskasing regions. Seismic sections from the Abitibi belt and on lines 8, 9, and 10 in the northern Kapuskasing uplift display strong reflectivity throughout the crust. However, the deeper parts of the section in the Chapleau block show very little reflectivity. If the interpreted amount of thickening is removed, a continuous pattern of moderately north-dipping reflectors can be traced smoothly from block to block throughout the region.

Suggestions for further research

Advances over the past decade have led to increased attention on outstanding problems. These include incomplete understanding of the timing, nature, and magnitude of Archean and Proterozoic events, and lack of information on the cause of seismic reflectivity, inhibiting use of the technique as a mapping tool in crystalline terrane.

Regarding the Archean history, a critical element lacking in the geochronological database is primary depositional or intrusive ages for major rock units of the Kapuskasing structural zone, including mafic gneiss, anorthosite, tonalite, and diorite. Model Sm−Nd ages provide maximum estimates for paragneiss and mafic gneiss (2750 Ma) and metamorphic zircon ages, minima (2696−2616 Ma); however, precise dates are required to test models of Archean crustal growth. For example, Krogh's (1993) model of tectonic underplating predicts progressively younger depositional ages with increasing depth. Seeing through the effects of the high-grade metamorphism represents a challenge for future geochronological studies.

An enigma in the late Archean tectonic history of the Kapuskasing zone is the origin of the Borden Lake conglomerate. Interpretations of the mechanism of emplacement of this unit into deep-crustal levels undergoing high-grade metamorphism have strongly influenced thinking on Kapuskasing tectonics. Further geochronological study aimed at determining the depositional age and metamorphic and structural history of the Borden Lake belt would provide valuable additional constraints.

Two major schools of thought apply to the uplift history of the Kapuskasing structural zone: (1) a protracted history of incremental Kapuskasing-related uplift events (2.63, 2.50, 2.3, 2.0, 1.85 Ga) is responsible for exhumation of the high-grade rocks and (2) a single Kapuskasing-related event at ~1.9 Ga, superimposed on regional exhumation, accounts for the observed isotopic and paleomagnetic constraints. The disparity between the two viewpoints results primarily from the discrepancy between Ar cooling dates and dyke paleomagnetic studies combined with Rb−Sr biotite dates. The cooling history in the interval 2.5−1.9 Ga has not been adequately resolved and could be improved through a regional study of U−Pb rutile dates to determine the time of cooling through the 400°C range. Dating of minerals within kinematically constrained fault zones provides a further means of determining uplift timing. A further major contribution to deconvolution of the early Proterozoic history would be acquisition of U−Pb ages on both quartz- and olivine-bearing Kapuskasing dykes. Previous assumptions of a single age for the swarm may have misguided interpretation.

Understanding of the geological nature of seismic reflectivity has not kept pace with acquisition of reflection data. Interpretation of the profiles could be improved through better correlation of geological and geophysical characteristics, through determination of physical properties and modeling. Owing to imperfect exposure, this goal has been only partly achieved in the Kapuskasing transect, such that it is difficult to identify reflectors with confidence. Drilling through shallow reflective sequences may be the most effective means of acquiring material for study.

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