Applications of seismic pattern recognition and gravity inversion techniques to obtain enhanced subsurface images of the Earth’s crust under the Central Metasedimentary Belt, Grenville Province, Ontario

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Accepted 2000 June 26. Received 2000 June 22; in original form 1999 February 9

SUMMARY
Project Lithoprobe’s Abitibi-Grenville transect seismic reflection lines 32 and 33 traverse the exposed Central Metasedimentary Belt (CMB) located in the Grenville province of the Precambrian Shield of Canada in southern Ontario. These seismic lines image a zone with a protracted deformational history spanning more than 300 Myr. Detailed examination of the commercially processed stacked sections reveals a number of significant deficiencies in some important areas. The image quality in these zones of reduced coherency needs to be enhanced to examine specific features and their relation to the surface geology. Examination of near-vertical seismic data from Lines 32 and 33 revealed that the signal-to-noise ratio was not improved by stacking, due to misalignment of signals even after static, normal moveout corrections and residual static corrections. The presumed reason is that reflected seismic energy following long ray paths in heterogeneous media suffers from relative advances and delays in its propagation, and hence arrives at slightly different times at the receivers, tending to be poorly aligned relative to its theoretical traveltime curves. A pattern recognition (PR) method for signal enhancement followed by energy stacking in moving time windows was used in this study to improve the images in spite of misalignments. Reprocessing has refined the geometry of the reflection profiles.

The objective of this paper is to use enhanced images of the seismic reflection data obtained by using a PR approach together with gravity data, using 2.5-D forward and 3-D inversion routines, to give an improved model of subsurface structure in the vicinity of lines 32 and 33.

Line 32 is dominated by southeast-dipping reflectors sloping into the lower crust. The listric geometry of the strong reflection packages of the CMB boundary thrust zone is interpreted to represent a crustal-scale ramp–flat geometry that accommodated northwest-directed tectonic transport of the CMB. This interpretation is also supported by the gravity data.

Key words: crustal structure, gravity, inversion, Moho reflection, pattern recognition, seismic reflection.

INTRODUCTION
‘Lithoprobe is Canada’s national, collaborative, multidisciplinary Earth science research project established to develop an understanding of the evolution of the Canadian landmass.

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foccuses on the late Archean Abitibi greenstone belt, which is part of the southern Superior Province, the central core of the north American craton, and on the Mesoproterozoic Grenville orogen, which is principally exposed in the southeastern Canadian shield. The Grenville province resulted from a Mesoproterozoic continental collision and consists of tectonically stacked slices of late Archean, Paleo-Mesoproterozoic rocks that are exposed at various crustal levels along the strike of the orogen’ (Clowes 1997). In 1991, several hundred kilometres of near-vertical seismic reflection data were collected in both the Abitibi and Grenville provinces with the aim of getting a better insight into the structure of the Earth’s crust in this area. This study focuses on two seismic lines, Lines 32 and 33 in southeastern Ontario (see Fig. 1) that traverse the Central Metasedimentary Belt (CMB) in the Grenville province.

Initial Lithoprobe interpretations of the seismic images from Lines 32 and 33 were based on conventionally processed stacked sections. Detailed examination of these sections reveals a number of significant deficiencies in some important areas. The image quality in these zones of reduced coherency needs to be enhanced if specific features and their relation to the surface geology are to be examined. The original images show a weak Moho signature and incoherent reflections towards the southern end of Line 32 and for the whole of Line 33. Are these artefacts of processing? Is there signal penetration down to the Moho? Can we characterize the crust using other physical properties (e.g. density distribution)? These are some questions that can be answered from an improved image of the Earth’s crust. In this paper we mainly focus on three objectives:

(i) to enhance images of the seismic reflection data from Lines 32 and 33 by the application of a pattern recognition (PR) approach for signal enhancement and energy stacking in time windows (Roy & Mereu 1996);
(ii) to generate 2.5-D forward models (Webring 1985) from Bouguer-corrected gravity data along these lines using structural information from seismic data and to estimate a 3-D distribution of the anomalous density below the surface using a 3-D inversion routine (Li & Oldenburg 1998);
(iii) to interpret Lines 32 and 33 reflection images in the light of known surface geology, gravity and results from coincident seismic refraction profiles.

Figure 1. Location map showing the 1991 LITHOPROBE reflection Lines 32 and 33 and the major geological subdivisions of the Central Metasedimentary Belt, Grenville Province, Ontario. Numbers indicate shotpoint numbers for the seismic line.
**GEOLOGICAL SETTING**

The CMB, located in the southwestern Grenville Province, Ontario, underwent large-scale tectonic movements during the Grenville orogeny (~1000 Ma). During this orogeny, allochthonous units of the CMB and parautochthonous units of the Central Gneiss Belt (CGB) were accreted to the Archean Abitibi and Superior provinces. Lithoprobe seismic reflection lines 32 and 33 extend ~130 km from north of Bancroft to Belleville, Ontario. Together they traverse the exposed CMB of the Ontario Grenville Province at a high angle to strike. Ca. 1400–1000 Ma metamorphic and igneous crystalline rocks were penetratively deformed and metamorphosed at upper to mid-crustal depths at ca. 1180–1060 Ma during crustal shortening and thickening, and were exhumed at ~950–1000 Ma. Geological and geophysical data from the Ontario Grenville Province are consistent with the interpretation that there was an extensive mountain belt and overthickened continental crust during Mesoproterozoic orogenesis (Wynne-Edwards 1972; Rivers et al. 1989). Thus, these seismic lines image a zone with a protracted deformatinal history spanning more than 300 Myr.

The northern end of Line 32 is located within the basal portion of the CMB, where it is bounded by a ~10 km thick southeast-dipping dikele thrust zone termed the Central Metasedimentary Belt boundary thrust zone (CMBbtz) (Fig. 1; Easton 1988; Hamner 1988; Hamner & McEachern 1992). The northwest-directed thrust carried rocks of the CMB and juxtaposed them against the CGB to the north (monocyclic allochthonous and polycyclic parautochthonous belts, respectively, of Rivers et al. 1989). The CGB consists of fault-bounded domains of reworked granulite to amphibolite facies Archean and Proterozoic rocks of the Laurentian craton as well as <1800 Ma supracrustal and igneous rocks that are interpreted to be affiliated to the palaeomargin of continental Laurentia (see Davidson 1995; Culshaw et al. 1997; Carr et al. 2000). Lines 32 and 33 cross internal lithotectonic domains within the CMB, namely the Bancroft, Elzevir and Mazinaw terranes. The Bancroft terrane consists mainly of <1400 and >1250 Ma marbles, calcisilicates and siliciclastics, interpreted as continental shelf deposits, with minor metavolcanic rocks, as well as distinctive 1180 Ma gabbros (McEachern & van Bree-men 1993) and 1270–1220 Ma nepheline syenites and syenites (Miller et al. 1997). The Elzevir terrane comprises mafic metavolcanic, volcaniclastic and metasedimentary rocks and carbonates intruded by a variety of plutonic suites. The CMB underwent plutonism and metamorphism at 1250–1230 and 1130–1070 Ma (Lumbers et al. 1991; Easton 1992). The Mazinaw terrane projects into the subsurface beneath the Palaeozoic cover rocks at the end of Line 33. It contains metavolcanics and marbles as well as intrusive rocks.

**SEISMIC DATA PROCESSING**

The optimum combination of processing steps to achieve the scientific objectives included some standard procedures together with several signal enhancement operations. Fig. 2 outlines the major steps in the processing flow. Reprocessing was performed from raw field shot gathers. Refraction statics were computed using Hampson & Russell GLI-3D software. In addition to computing the static shifts, the near-surface velocity models were also correlated to the surface geology. Velocity filtering using the F-K filter was performed for certain shot gathers where high-amplitude, low-velocity surface waves were masking weaker reflections. Geometrical spreading corrections using a regional velocity function were applied. Velocity analysis in the CDP domain was performed using the constant-velocity

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**Figure 2.** (a) Major processing steps applied for the reprocessing of Lines 32 and 33. (b) Major steps in the PR algorithm.
stacking method. Dip moveout corrections were performed on the pre-stack CDP gathers, which focused on some of the SE-dipping reflectors seen between 2 and 8 s two-way traveltime (TWT) on Line 32.

An important pre-stack signal enhancing step is noise removal using a PR technique and subsequent energy stacking in time windows (Roy & Mereu 1996). The stratification of the rock units in the sedimentary basins is near horizontal with less lateral heterogeneity, hence the ray paths are fairly predictable and arrive at the surface receivers at their predicted traveltimes. Both static corrections for removal of the weathering layer and normal moveout corrections make the reflected signals well aligned and hence the stacking performs well. Many of the crustal experiments are carried out in areas where the crust has undergone some tectonic deformation and the layers have lost their original stratification. It is in these areas that there is a potential danger of losing a lot of information during processing. For such non-conventional targets, data-adaptive procedures such as the PR approach can be more effective. The rationale for using this method is that reflected seismic energy following long ray paths in vertically and laterally heterogeneous media suffers from relative advances and delays in its propagation, and hence arrives at slightly different times at the receivers, tending to be poorly aligned relative to its theoretical traveltime curves. Static corrections cannot correct for these deep-seated effects. Conventional stacking routines often fail to give a good image of the crust if the reflected signals are not properly aligned, hence a PR method was applied on the pre-stack processing stream in an effort to increase the signal-to-noise ratio (SNR). The discrimination between signal and noise is based on attributes such as (i) lateral continuity, (ii) amplitude, (iii) polarity and (iv) frequency of the signal. The trace-to-trace matching algorithm for pattern selection and energy stacking in a moving time window take into account misalignment present in the reflected signals. This is a time-domain noise filter and is applied to CDP gather in the form

\[
W_j(d_j^{\text{before}}) = d_j^{\text{after}},
\]

(1)

where \(d_j^{\text{before}}\) is a CDP gather before PR filter, \(d_j^{\text{after}}\) is a CDP gather after PR filter and \(W_j\) are the weights for each sample point decided from the PR algorithm. These weights are either 0 or 1. The arbitrary parameters that control the algorithm are: (i) \(F\) (controls the amplitude threshold), (ii) \(N_{\text{w1}}\) (the window size for the coherency test), (iii) \(N_{\text{w2}}\) (the window size to compute the energy stack) and (iv) \(N_t\) (the number of traces used for the coherency test). Optimum values of these parameters were chosen after considerable testing. A 2-D window \(N_t \times N_{\text{w1}}\) is slid along the entire CDP gather in search of coherent events based on the attributes mentioned above. After the selection of the reflection pattern, an energy function \(E\) is produced. The energy of the peaks is computed over a small time window \(N_{\text{w2}}\) across the whole CDP gather. The arrival time of this energy, \(E\), is determined from the average of the arrival times of the signals in that time window. A distinction is made between peaks with positive and negative polarities. As the window is moved, the amplitude builds up near a real reflected signal and then decreases again. This leads to a slight widening of the reflector. A flow chart showing the major steps in the PR algorithm is presented in Fig. 2(b).

An example of the application of the PR method is shown in Fig. 3. Note the significant improvement of the stacked signal in the vicinity of the reflector by the application of the PR approach (compare Figs 3c and d). It can be seen from Figs 3(a) and (b) that the strong event is accepted but along with the random noise, a variety of well-aligned signals (or coherent noise), but not horizontal events, have been muted. However, these signals are of much lower amplitude and coherent only over 3–4 traces. Hence, a combination of high-amplitude threshold factor \(F\), high \(N_t\) and low \(N_{\text{w2}}\) may lead to a failure in recognition of weak coherent signals. It should be emphasized that a proper selection of PR parameters is essential for the success of the algorithm, along with a trade-off between how much of the data is signal and what should be considered as noise.

The post-stack section is then migrated, bandpass filtered, energy balanced, coherency filtered and reduced to the form of line drawings for display.

**RESULTS FROM SEISMIC DATA**

Seismic images and their correlation with geology

Upper-middle crust

Comparison of conventionally processed and reprocessed stacked sections from Lines 32 and 33 using a PR approach indicate distinct improvements in the subsurface near the southern end of Line 32 and for the whole of Line 33 (see Figs 4 and 5). Lines 32 and 33 are oriented in a nearly N–S manner and extend from the CMBBz in the north to the Palaeozoic
Figure 4. (a) Coherency filtered, migrated stacked section from Line 32 processed using conventional stacking routines. (b) Coherency filtered, migrated stacked section from Line 32 processed using the PR approach. Major SE-dipping reflectors can be observed from 0–8 s TWT. The Moho signature is poor except for in the northern part of Line 32.

rocks in the south, near Lake Ontario. Both profiles show very high reflectivity up to 8 s TWT, although there are lateral variations in the strengths of reflections.

Near the northern end of profile 32 (i.e. CDPs 300–1200), there is a package of strong, laterally coherent, apparently south-dipping reflectors (~20–30°) that extend from the surface to about a TWT of 8 s to the central portion of the profile, where reflectivity dies out. The top of this reflecting package (event A, Fig. 4b) when projected to the surface coincides with the trace of the CMBbtz. The zone of reflections is interpreted to represent a listric ductile shear zone that accommodated northwest thrusting of the CMB over the CGB. The character and apparent dip of the reflectors are consistent with the surface lithological units, which comprise gneissic rocks with high strain boundaries.

A zone of arcuate reflections (event B, Fig. 4b) overlies event A (CMBbtz). Package B is less reflective than package A. This package of reflection has an apparent dip towards the SE and is laterally continuous for about 30 km and dies out towards the SE. This reduction of reflectivity is due to the fact that the data quality deteriorates and the CDP gathers do not show any coherent reflection events. Hence, the PR algorithm fails to find a reflected signal. The north end of profile 32 (i.e. CDPs 300–1500) is underlain by a south-tapering wedge of relatively non-reflecting crust about 10 km thick seen between 7 and 10 s TWT (Fig. 4b). The improved seismic images presented in Figs 4(b) and 5(b) suggest a gradual soling of the mid-crustal reflectors (events A and B) into the lower crust.

Further south of reflection packages A and B along profile 32 (i.e. south of CDP 1500), the reflective nature of the crust reappears. At least three major SE-dipping reflectors can be identified that are fairly strong and continuous and extend from near the surface to about 8 s. The Bancroft terrane is not highly reflective; however, a distinct apparent SE-dipping reflecting zone penetrating up to almost 8 s separates the non-reflective Bancroft terrane from a more reflective zone to the south represented by reflection packages C. This reflector can be extrapolated to the surface near CDP 1500. Several other reflectors (event D) are juxtaposed against reflector C. The approximate dip of all these reflectors is ~20° towards the southeast. The reflective packages C and D are interpreted to represent Elzevir terrane rocks thrust westwards above the Bancroft terrane. The plutonic bodies present in the Elzevir terrane are not mapped by the seismic data due to either their small size or their steep dip.

Line 33 crosses Line 32 at the north end (at approximately CDP 350 or SP 201) and extends southwards from the Elzevir terrane to the Mazinaw terrane into the Palaeozoic cover rocks. It intersects the Moira Lake Shear Zone (MLSZ) at CDP 1000. The reflective nature of profile 33 (Fig. 5b) is different from that of profile 32. The reflections in the upper crust of Line 33 are very gently dipping (~10°), generally more heterogeneous and form short segments (i.e. <10 km). Towards the north end of the line, there are no major dipping reflectors except for package E, which becomes strong at certain parts of the profile. This reflector dips gently towards the south and can be roughly traced throughout the entire profile. Reflection package E can be interpreted as a ramp for the NW transport of Mazinaw terrane rocks. The MLSZ is a steep late extensional fault and is not well imaged by the seismic reflection data. However, there is a clear break of reflector E between CDPs 500 and 1000 at about 4.7 s TWT that may be correlated to the MLSZ. At about CDP 1200, the reflective nature of the upper crust is enhanced (Fig. 5b). A number of small antiformal reflectors (F) can be seen...
Figure 5. (a) Coherency filtered, migrated stacked section from Line 33 processed using conventional stacking routines. (b) Coherency filtered, migrated stacked section from Line 33 processed using the PR approach. The reflectors are more gently dipping than those observed on Line 32. The lower crust is less reflective and a clear Moho signature is absent.

scattered from near the surface to about 5 s, which are underlain by the reflector E. These antiforms probably represent the thrusted Mazinaw terrane on top of the Elzevir terrane. The Mazinaw terrane in this area is covered by the Palaeozoic rocks. The Robertson Lake Mylonite Zone (RLMZ), which is a shallow thrust feature and exhibits significant normal motion, is observed on the surface near CDP 2000, Line 33. It is not clearly defined by near-vertical reflections, although it has been imaged by Zelt et al. (1994) in a wide-angle survey.

Lower crust–Moho

Although the Moho is present as a relatively sharp boundary, as seen from refraction results, it is not observed on near-vertical reflection data from Lines 32 and 33. Except for CDPs 300–1700 of Line 32, the crust below 8 s TWT for both Lines 32 and 33 appears to be non-reflective. The PR approach, which greatly enhanced the reflectors in the upper crust, failed to observe any Moho reflectivity. There are a few weak and discontinuous segments occurring at about 11–12 s between CDPs 300 and 1700 of Line 32 (event M, Fig. 4b). The reflection package M has been utilized to interpret Moho depths to be ~38–42 km assuming an average crustal velocity of 6.5 km s⁻¹ (migration velocity based on the average between near-surface velocities and velocities in the lower crust obtained from wide-angle refraction data), which is consistent with that interpreted by Winardhi & Mereu (1997).

The reflective package M on Line 32 has been interpreted to represent the limits up to which the Laurentian Archaean craton (Carr et al. 2000; White et al. 2000) north of the GFTZ was underthrust below the CGB (see Fig. 12). This is in general agreement with the reflective lower crust in Archaean terranes, where the Moho is marked by the gradual die-out or termination of reflectivity (Ludden et al. 1993; Calvert et al. 1995). It has also been suggested by Kellet et al. (1994) from their study of the magnetic and aeromagnetic relief that the Archaean rocks of the foreland extend well south of the Grenville Front. This can also be seen from the velocity anomaly map of refraction Line MG of the 1992 Abitibi–Grenville refraction experiment (Winardhi & Mereu 1997), where the Archaean crust extends 140 km south of the Grenville front below the CGB rocks.

Correlation of near-surface velocity variations

The near-surface (top 2 km) lateral variation of velocity was also studied to see if there is a correlation with the surface lithology. Velocities were computed by first-break analysis of
the data set. A 1-D velocity profile obtained along with the error bars and surface geology for Lines 32 and 33 is shown in Fig. 6. The near-surface velocities range from 5.8–6.2 km s$^{-1}$. The presence of very high-velocity rocks near the surface agrees well with the velocity anomalies as well as delay time analysis for the near-surface velocity from refraction studies in this area by Winardhi & Meru (1997). At the beginning of Line 32, a high-velocity region near CDP 150 probably corresponds to the high-grade metamorphic rocks of the CGB and the transitional zone of the CMNbOtz, which comprises of gneisses and tectonites. The velocities of these rocks can vary between 5.0 and 7.0 km s$^{-1}$ (Schon 1996). The Bancroft terrane consists of marble, meta-plutonic rocks and siliciclastic rocks with an average velocity of 5.8-6.0 km s$^{-1}$, in agreement with published velocities of such rocks. Nepheline syenite has a relatively lower velocity, i.e. 5.6 km s$^{-1}$ (Schon 1996), and this is probably represented by the local minima in the velocity profile.

The Belmont domain of the Elzevir terrane consists mainly of marble and mafic metavolcanics of medium metamorphic grade indicated by a slightly higher velocity, i.e. 6.0–6.2 km s$^{-1}$. The Elzevir terrane contains numerous diorite–gabbro intrusions, including the Thane, Umirflaville and Tudor intrusions, which are crossed by Line 32. These plutonic suites are very thin and cannot be imaged by the velocities; however, belts of gabbros and metavolcanic rocks are visible in the velocity diagram as localized high-velocity regions with a velocity of up to 6.5–6.7 km s$^{-1}$, which agree well with the velocity measurements for gabbro and diorites (Fig. 6). The marble belts are represented by average velocities of 6.0–6.1 km s$^{-1}$. Between CDPs 200 and 500 of Line 33, higher velocities are observed, i.e. 6.3–6.5 km s$^{-1}$. This corresponds to the Grimsthorpe domain of the Elzevir terrane, which is dominantly composed of metavolcanics and gabbros. This region is followed by a low-velocity zone of 5.2–5.5 km s$^{-1}$ corresponding to the lower-grade Belmont domain rocks observed towards the south. This zone continues up to the MLSZ. The Mazinaw terrane rocks are similar to the Elzevir terrane rocks. The Mazinaw terrane is marked by an increase in velocity from 5.5 to 5.9 km s$^{-1}$ between CDPs 1000 and 1500 on Line 33. The lower-velocity Palaeozoic rocks, with an average velocity of 5.4 km s$^{-1}$, surround the Mazinaw terrane. Near-surface velocity studies suggest the fact that the observed lithology also has its expression in the seismic velocity, and this information can be utilized in defining lithological boundaries.

**Moho reflectivity**

Compared to the reflections in the upper crust (see Figs 4 and 5), the Moho reflections are more scattered and less continuous and the crust–mantle boundary is poorly defined. If we assume that the noise is a stationary process, then the limits of signal penetration can be found where the seismic signal is overwhelmed by the noise so that the amplitude and frequency contents ceases to change (Barnes 1994). This limit exists since the seismic energy decreases in both amplitude and frequency content as it propagates due to the prevailing noise (Mayer & Brown 1986). To compute the amplitude decay curves, CDP gathered traces were processed (for example, by bandpass filtering, f-k filtering, spreading corrections, NMO corrections, static corrections) and conventionally stacked without the application of the PR method. Amplitude decay curves were then computed for all of the stacked traces (with trace normalization
applied to balance laterally varying noise bursts) and spatially averaged over an area to obtain the representative decay curve for that area. Fig. 7 shows a representative amplitude decay curve from the northern part of Line 32 and from a CDP gather from Line 33. There is a $10^{0.5}$ dB difference in the background noise level from Line 32 to Line 33. Local fluctuations in time and space of this amount are common, but systematic differences of this amount between similar environments are uncommon. We attribute this difference to differences in acquisition conditions. These curves clearly show that for the initial part of Line 32, there is a small rise in the decay curve at about 11–12 s, beyond which the curve becomes stationary, reflecting the background noise. This Moho signature can be observed almost up to CDP 1000. For the southern ends of Lines 32 and 33, the amplitude decay curve falls gradually and becomes stable beyond 8 s TWT. This corresponds to the depths of signal penetration for most of Lines 32 and 33. The reasons for such a high degree of attenuation may be energy losses from both elastic scattering from small-scale reflectors and anelastic losses. Average estimates of the effective quality factor $Q$, which is a measure of the seismic wave attenuation (using the spectral ratio method), for Lines 32 and 33 are about 200–300 (Roy 1997). Such values of $Q$ are very low for the crystalline crust in a low-heat-flow regime. However, it should be kept in mind that the effective quality factor $Q$ is a combined effect of different loss mechanisms (Knopoff 1964) and does not reflect the intrinsic quality factor $Q$. Since we know that the metabasaltic, metasedimentary and other volcanioclastic rocks in the CMB were thrusted and deformed along ductile faults, elastic losses from the scattering may be one of the major factors contributing to the low $Q$ factor.

**RESULTS FROM BOUGUER GRAVITY DATA**

2.5-D modelling of potential data

Bouguer gravity anomaly data for the study area were obtained in digital format from the National Gravity database (Geological Survey of Canada, Ottawa). Theoretical responses were calculated for 2.5-D structural models and compared with the observed data using a forward modelling routine (Webing 1985). A 2.5-D model is one in which the model is constructed as an ensemble of prisms of polygonal cross-section allowing for finite lateral extent in the along-strike direction. Polygons were designed based on a priori information from the seismic data. The assigned densities were based on published sample density compilations (Schon 1996). To eliminate edge effects at the region of interest, the models were extended by 500 km. The depth extent of the model was extended to 150 km. The parameters were interactively varied during successive iterations to reduce the root-mean-square misfit error between the calculated and observed data. Additional details of the modelling and examples of modelling along the western Canadian margin are given by Dehler & Clowes (1992) and Clowes et al. (1997).

Results from 2.5-D modelling of gravity data

Fig. 8(a) is a 2-D Bouguer gravity anomaly map with seismic lines 32 and 33 superimposed. A gravity profile is chosen along the seismic profile (Fig. 8b) and the result of the 2.5-D modelling is shown in Fig. 8(c). The start of the seismic line is taken as the reference point (0 km) and distances are measured relative to that. The main feature of Fig. 8(a) is a long-wavelength component of the anomaly from $-65$ to $-32$ mgal in the SE corner that does not coincide with the strike direction of major structural boundaries, possibly because the low anomaly at the SW corner of the map area is due to a thick Palaeozoic cover (close to Lake Ontario). In spite of this gravity feature, the seismic lines are at a high angle to the strike (from Fig. 1), hence construction of 2.5-D models is legitimate.

The long-wavelength component is controlled by lower crustal structures (for example, the topography of the Moho and density variations in the mantle), whereas the higher-frequency components are controlled by upper crustal density variations. The SE-dipping reflectors seen on the seismic section have also been modelled from the gravity data. Densities of

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**Figure 7.** (a) Amplitude decay curve from the northern end of Line 32 showing limits of signal penetration. A clear Moho signal can be observed close to 11 s TWT. (b) Amplitude decay curve from Line 33, showing no signal penetration beyond 8 s TWT.
Figure 8. (a) Bouguer gravity anomaly map of the study area with locations of Lines 32 and 33 superimposed. The NE–SW strike of the CMBbtz can be seen on the map by a change in gravity anomaly from low in the NW (CGB) to high in the SE (CMB). Dashed box indicates region over which 3-D inversion was performed. (b) Actual (crosses) and 2.5-D (solid line) gravity profiles along Lines 32 and 33. (c) The density model showing the polygons. The densities are in SI units (kg m$^{-3}$).

rocks in the top 5 km range from 2550–2670 kg m$^{-3}$. Densities gradually increase from the Bancroft terrane to the Elzevir terrane. This may suggest that rocks in the Elzevir and Mazinaw terranes have been uplifted from greater depths compared to those in the Bancroft terrane and/or that they are more mafic in composition. This compositional difference is also reflected by the increase in near-surface velocities from the Elzevir terrane to the Bancroft terrane. The presence of more mafic rocks in the Elzevir terrane in contrast to the large tracts of marble and siliciclastic rocks in the Bancroft terrane may have its expression in the lower crust. Similar density variations are also observed in the 3-D inversion results discussed below. The absence of density measurements of surface samples from this region prohibits direct calibration of the 2.5-D models.

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3-D inversion of potential field data

For the 3-D analyses, gravity inversion algorithms developed by Li & Oldenburg (1998) have been used. These algorithms have considerably more flexibility in defining subsurface structures than procedures available in the 1970s. Li & Oldenburg (1998) demonstrated the application of the method at the scale of mining problems. As in any inverse problem the user has a set of observations, some estimate of the uncertainties and a theoretical relationship that enables one to compute the predicted data for any model (m). The model (m) represents the spatial distribution of the physical property under investigation, in this case density. The goal is then to find the model that gives rise to the observations. The problem is numerically solved by discretizing the Earth into cells of constant density. The model (m) is a vector, and the inverse problem is solved by minimizing the model objective function subject to the constraint that the observed data be reproduced (Oldenburg et al. 1998). The data misfit is a sum-of-square measure assuming the observations are contaminated with Gaussian noise and an estimate of the standard deviation is available. For gravity and magnetic problems the relationship between the model parameters and the data is linear, and the solution to the inverse problem is obtained by iteratively solving a single matrix system of equations.

The vertical component of the gravity field produced by the density \( \rho(x, y, z) \) is given by

\[
F_z(r_0) = \gamma \int \rho(r) \frac{z - z_0}{|r - r_0|^3} \, dr,
\]

where \( r_0 \) is the vector denoting the observation location and \( r \) is the source location (Li & Oldenburg 1998). \( V \) represents the volume of the anomalous mass and \( \gamma \) is the gravitational constant. The data from a typical gravity survey are a set of measurements acquired over a 2-D grid. The data are processed to obtain an estimate of the anomalous field due to the anomalous mass below the 2-D grid, and the goal of the gravity inversion is to obtain quantitative information about the distribution of the anomalous density in the ground.

The inverse problem (Li & Oldenburg 1998) is solved by finding a model \( (m) \) that minimizes a model objective function \( \phi_m \) subject to satisfying the data constraints. The inverse problem can be stated mathematically as

\[
\text{minimize} \quad \phi_m = \| W_m (w)(m(x, y, z) - m_0(x, y, z)) \|^2
\]

subject to \( \phi_d = \| W_d (\phi_{obs} - Gm) \|^2 = \phi_d^* \).

To solve the inverse problem a global objective function \( \phi = \phi_3 + \beta \phi_m \) is minimized, where \( \beta \) is a trade-off parameter that controls the relative importance of the model norm and the data misfit. \( W_m \) is the model-weighting matrix, \( w(z) \) is the depth-weighting function that counteracts the geometrical decay of the sensitivity with the distance from the observation location, \( \phi_3 \) is the data misfit and \( \phi_d^* \) is the target misfit to be achieved after the inversion. The model-weighting function can be chosen to penalize roughness in three spatial directions and also be close to a reference model. It can be represented by

\[
W_m = x_0 W_y W_x + x_1 W_z W_y + x_2 W_z W_x + x_3 W_x W_y W_z ,
\]

where \( W_z \) is the weighting function that determines the closeness of the model to a reference model. \( W_x, W_y \) and \( W_z \) are the spatial derivatives of the model in three directions, \( x, y \) and \( z \) respectively. \( x_0, x_1, x_2, \) and \( x_3 \) are the control parameters that determine the relative importance of the components in the objective function. The non-uniqueness of the inverse problem is minimized by choosing a proper model objective function that produces a smooth model and the depth-weighting function \( w(z) \), which places the anomaly at its proper depth. The system of equations is solved using subspace techniques for the gravity inversion.

An inherent difficulty in gravity problems is that there is no depth resolution. Consequently, structures concentrate near the surface in a simple smallest or flattest model regardless of the true depth of the causative bodies. This is due to the rapidly diminishing amplitude of the kernel with depth. The kernels of the surface data are not sufficient to generate a function that possesses significant structure at depth. To overcome this, Li & Oldenburg (1998) devised a depth-weighting function in the inversion scheme that approximately cancels the natural decay of the kernel and gives cells at different depths equal probability of entering the solution with a non-zero density. These weights were empirically determined by Li & Oldenburg (1998), and their tests suggest that the model constructed by minimizing the model objective function subject to fitting the data places the recovered source at approximately reasonable depths. This ensures that the anomalous distribution of density obtained from the inversion can be interpreted to represent density variations of the Earth’s crust, and hence a correlation to the geological environment is sensible.

Results from 3-D inversion of gravity data

3-D gravity inversion was carried out for a 200 x 300 km area and up to 70 km depth. The area is the same as that shown in Fig. 8(a); however, coordinates have been changed to UTM coordinates to facilitate the inversion algorithm. A problem in the processing and interpretation of gravity data is the separation of the long- (regional) and short-wavelength components of interest. Regions over larger areas can be removed by fitting a trend or a polynomial to the data. In this study a constant regional gradient was assumed for two reasons: (i) the gravity data over a larger area encompassing the study area (see Fig. 8a) do not show major changes in the trend; and (ii) regionals are controlled by changes in the depth of the Moho or large-scale density variations in the lower crust or upper mantle. However, neither seismic reflection nor refraction data (Winardhi & Mereu 1997) show such variations.

The 3-D volume of the Earth represented by this area is divided into cells. The grid spacings in the \( x \) and \( y \)-directions were taken as 5.0 km and that in the \( z \)-direction was taken as 2.5 km. The region for inversion was extended on all sides by four cells to reduce edge effects. An average density for crustal rocks of 2650 kg m\(^{-3}\) was chosen as the background density. A reference anomaly model (that is, the absolute density minus the background density) was created assuming a gradual change in density from 2600 kg m\(^{-3}\) at the surface to 2900 kg m\(^{-3}\) at the lower crust and 3200 kg m\(^{-3}\) below the Moho to constrain the inversion. The objective function was chosen to give the smoothest and flattest distribution of the density anomaly. At this point we would like to emphasize that the results from the 3-D inversion do not give absolute densities, but only give a quantitative indication of the anomalous density distribution in the subsurface. Therefore, a direct correlation of the 2.5-D model densities and the 3-D model densities is not possible.
The results of the 3-D inversion are shown in Figs 9 and 10. Smaller wavelength anomalies are concentrated at smaller depths, whereas longer-wavelength anomalies are at greater depths, as shown by the depth sections in Fig. 10. Higher densities occur below the Elzevir and Mazinaw terranes than beneath the Bancroft terrane. The CGB shows a very gradual increase in density from 0–40 km depth. This may reflect more homogeneous internal CGB rocks and may also explain the relatively non-reflecting crust in the CGB as seen on Lines 30 and 31 (White et al. 1994). The juxtaposition of higher-density CMB rocks against the lower-density footwall rocks of the CGB also support the idea of mid-crustal exhumation along regional-scale décollement zones, i.e. the CMBtz. Density images indicate the amalgamation of two different crustal units—the CMB and the CGB—to the NW, augmenting the current idea of NW-directed tectonic transport during the Grenville orogeny. The presence of an anomalous density distribution up to lower crustal levels between the CGB and CMB suggests that this process and the accretion of the two units was a crustal-scale process.

**DISCUSSION**

Seismic images of the CMB show crustal-scale fault zones marked by arcuate reflectors truncated at their bases by linear SE-dipping reflectors analogous to the Grenville Front Tectonic Zone (Green et al. 1988; Milkeret et al. 1992), and support models of NW-directed crustal shortening during the Grenville orogeny. Fig. 11a is a map showing the location of all the reflection and refraction profiles in the Abitibi-Grenville province; Table 1 compares the general nature of crustal reflectivity found from other reflection profiles in the Grenville Province and surrounding Archaean Superior, Abitibi and Southern provinces.

Table 1 indicates that the TW of the interpreted Moho is fairly consistent, between 11 and 13 s in all lines. Assuming an average crustal velocity of 6.5 km s⁻¹, the depth of the Moho ranges from 36–42 km. However, the nature of the Moho reflectivity is highly variable in the Abitibi and Grenville provinces. In most of the profiles mentioned in Table 1, it is marked by a reduction in the reflections from the mid-lower crust, suggesting a fairly non-reflective mantle, except for Line 48 in the Superior Province, where relics of a subduction zone have been inferred. These characteristics of the lower crust and Moho may arise from (i) a gradual transition in the acoustic properties, rather than a sharp crust–mantle boundary (Winardhi & Meru 1997) or (ii) a first-order Moho, characterized by irregular interfaces that do not generate reflections that can be stacked effectively (Nemeth et al. 1996). The absence of a distinct Moho in the Grenville Province reflection lines is consistent with the global pattern of Archaean crust (Durrheim & Mooney 1991, 1994) and a common feature in convergent terranes (Mooney & Brocher 1987). The Moho is better determined in the relatively undisturbed parautochthonous Archaean provinces to the north of the Grenville Front, for example, in Lines 27, 28, 29, 23, 24, 25, 15, 16A, 17 and 48. On the other hand, Lines 32, 33, 30 and 31 and GLIMPCE Line J, all located south of the Grenville front in the CGB and CMB, representing reworked a Proterozoic margin, show a poor Moho signature from near-vertical reflection data. This is contrary to other areas of the Canadian shield (Lucas et al. 1993; Hajnal et al. 1996), where the Moho is part of the Proterozoic tectonic
process and is always a strong reflector. Hence, the presence or absence of a reflective Moho may be related to its tectonic history. Poor Moho signals from near-vertical data do not imply that a Moho is absent; wide-angle PnP signals are quite clear even in these tectonic areas (Winardhi & Mereu 1997; Zelt et al. 1994; Grandjean et al. 1994; Hughes & Luetgert 1992). Nemeth et al. (1996) illustrated that the difference in Moho reflection signatures from near and wide-angle data is probably a result of the different frequency spectra of the two signals and the dissimilarity in their amplitude due to the differences in the angle of incidence of the wave front. The simplest model of the Moho capable of generating high-amplitude reflections at wide-angle incidence and low-amplitude reflections at normal incidence is a smooth gradient transition zone at least 300 m thick, and not more than 2 km thick, given the 35 Hz dominant frequency of reflection data (Meissner 1967) and 4 Hz dominant frequency of pre-critical PnP amplitudes (Zelt et al. 1994).

The overall structure and reflection character shown in Figs 4 and 5 agree well with those presented by White et al. (1994). However, the new images show a more detailed geometry of the reflectors between 2 and 8 s TWT. The seismic reflection data from Line 32 image the CMBtz as a crustal-scale ductile shear zone that accommodated SE over NW crustal stacking. SE-dipping reflective bands of the CMBtz penetrate up to mid-crustal depths of 25–30 km. These reflection surfaces are steep at the surface and become listric at depth. They do not penetrate the Moho, as indicated by the absence of any major relief of the Moho topography in the CMB and CGB. CMB terranes were gradually translated towards the Laurentian craton over the mid-crustal thrust system or a ramp in a compressional regime (Forsyth et al. 1994). Refraction data (Winardhi & Mereu 1997) indicate higher velocities of mid-crustal rocks for the CMB than for the CGB. However, such a relief is not observed for the lower crustal velocities. This implies
Table 1. A table comparing the crustal reflective style in different parts of the Abitibi–Grenville.

<table>
<thead>
<tr>
<th>Location of reflection line</th>
<th>Nature of the crust</th>
<th>Moho reflectivity</th>
<th>Depth of Moho (TWT s)</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lines 32 &amp; 33, CMB</td>
<td>Reflective crust up to 8 s, with distinct SE-dipping shear zone reflectors</td>
<td>Very poor except for northern end of Line 32</td>
<td>12.5</td>
<td>This paper</td>
</tr>
<tr>
<td>Lines 30 &amp; 31, CGB</td>
<td>Upper crust has poor discontinuous reflectors, reflectivity enhanced in the lower crust</td>
<td>Not clear. Probably a decrease in reflectivity</td>
<td>12.5–13</td>
<td>White et al. (1994)</td>
</tr>
<tr>
<td>Reflection profiles in lakes Ontario and Erie</td>
<td>Reflective crust up to 6 s with distinct E-dipping shear zone reflectors</td>
<td>Not recorded</td>
<td>–</td>
<td>Forsyth et al. (1994)</td>
</tr>
<tr>
<td>GLIMPCE Line J</td>
<td>Non-reflective upper crust, reflective lower crust truncated by highly reflecting E-dipping shear zones down to 20 km</td>
<td>Not defined by a single reflection; marked by a die-out of reflectivity</td>
<td>11–13</td>
<td>Green et al. (1988)</td>
</tr>
<tr>
<td>Line 28, Abitibi</td>
<td>Three-layered crust. Crust reflective up to 10 s</td>
<td>Discontinuous but can be traced clearly</td>
<td>13</td>
<td>Ludden et al. (1993)</td>
</tr>
<tr>
<td>Lines 27 &amp; 29, Northern Abitibi</td>
<td>Three-layered crust</td>
<td>Marked by termination of reflectivity</td>
<td>12</td>
<td>Ludden et al. (1993)</td>
</tr>
<tr>
<td>Lines 16, 16A &amp; 17, Pontiac Subprovince</td>
<td>Three-layered crust. Reflections represent imbrication of slices</td>
<td>Marked by decrease of reflectivity</td>
<td>12–13</td>
<td>Ludden et al. (1993)</td>
</tr>
<tr>
<td>Lines 12 &amp; 14, southern Abitibi</td>
<td>Subhorizontal discontinuous reflections cannot be correlated to geology</td>
<td>Poor</td>
<td>12</td>
<td>Jackson et al. (1990)</td>
</tr>
<tr>
<td>Lithoprobe line in the Eastern Grenville</td>
<td>Highly reflective crust</td>
<td>Distinct Moho</td>
<td>13–16</td>
<td>Eaton et al. (1995)</td>
</tr>
<tr>
<td>Lines 21 &amp; 21A, Noranda camp of Abitibi Belt</td>
<td>Three-layered crust</td>
<td>Not recorded</td>
<td>–</td>
<td>Verpaelst et al. (1994)</td>
</tr>
<tr>
<td>Line 15, western Quebec</td>
<td>Reflective crust, but GFTZ not well defined</td>
<td>Well defined showing crustal thinning under GFTZ</td>
<td>11–12</td>
<td>Kellet et al. (1994)</td>
</tr>
<tr>
<td>Lines 52, 53 &amp; 54, Grenville Province of western Quebec</td>
<td>Highly reflective crust with intermittent transparent zones</td>
<td>Well defined by termination of reflectivity</td>
<td>11–13.5</td>
<td>Martignole &amp; Calvert (1996)</td>
</tr>
</tbody>
</table>

that the CMBlitz is associated with uplift that resulted from upward exhumation of mid-crustal blocks along a décollement. Similarly, in the 3-D gravity models, higher densities are observed in the CMB than in the CGB. Within the CMB, densities are higher in the Elzevir than in the Bancroft terrane. Fig. 12 shows our combined interpretation of Lines 32 and 33 as observed from the seismic and gravity data analysis. Fig. 12(a) clearly shows how the major thrust boundaries are continuous over the two lines. Higher densities in the CMB as compared to the CGB support our interpretation of the presence of décollement zones along which mid-crustal exhumation may have taken place. Note the close similarity between Figs 12(b), (c) and (d). Amalgamation of the domains within the CGB and CMB occurred prior to 1160 Ma and they were accreted to the Laurentian margin starting at 1120 Ma. The present-day crustal architecture is a result of the latest phase of the orogeny at ca. 1120–980 Ma (Carr et al. 2000).

**CONCLUDING REMARKS**

(i) Data-adaptive techniques such as the pattern recognition method for signal enhancement and energy stacking are particularly useful in places where the regional geology is complicated by different episodes of tectonic activity. The results

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obtained using a PR method show improved subsurface images from Lines 32 and 33 in the CMB. Due to some inherent limitations of the PR method, the top 1 s TWT cannot be imaged.

(ii) The seismic section from Lines 32 and 33 and its interpreted geometry indicate that the crust in the CMB is dominated by crustal-scale compressional structures from 0–8 s TWT. Mylonitic fabrics and their orientations and nappes within the CMB boundary thrust zone determine the character of the seismic reflection data. The listric geometry of the strong reflection packages of the CMB boundary thrust zone is interpreted to represent a crustal-scale ramp–flat geometry that accommodated NW-directed tectonic transport of the CMB. The N-directed movement of the CMB rocks over the footwall rocks of the CGB have created relatively thin terranes (Bancroft, Elzevir and Mazinaw terranes) with an apparent difference in reflection character. An interesting feature of the crust imaged by Lines 32 and 33 is the lower crustal transparency, except for a small stretch of reflectivity in the northern part of Line 32, which we interpret as the extension of the Laurentian crust present to the north of the Grenville Front below the CGB. The transparent Moho signature is explained in terms of low signal penetration arising from the scattering and absorption of energy.

Accurate estimates of velocities can help to distinguish lithological boundaries. A velocity of 5.8–6.2 km s⁻¹ at the near surface agree with the mid-crustal exhumation of the rocks to the surface. Belts of syenite and marble in the Bancroft terrane are marked by localized minima in the velocity profile.

Figure 11. Location map of other LITHOPROBE reflection and refraction lines as well as the 1982 COCRUST experiment lines.
(5.2–5.6 km s⁻¹), whilst tracts of gabbros or metavolcanics rocks in the Elzevir terrane can be located from the local high in the near-surface velocity variations (6.5–6.7 km s⁻¹).

(iii) Advanced imaging techniques such as gravity modelling and inversion play a significant role in characterizing the physical properties of the Earth’s crust. 2.5-D modelling of Bouguer gravity data indicates a complex geometry of the upper crust. In general, higher densities occur below the Elzevir and Mazinaw terranes than the Bancroft terrane. Similar features are also observed in the 3-D depth sections. The juxtaposition of higher-density CMB rocks against the lower-density footwall rocks of the CGB also supports the idea of mid-crustal exhumation along regional-scale décollement zones, i.e. the CMBdzt. A different density distribution of the CGB and CMB existing down to the lower crust supports the current idea of the amalgamation of two crustal blocks during the Grenville orogeny.

(iv) This integrated study using enhanced seismic images, density models obtained from 2.5-D modelling and anomalous density distributions from 3-D inversion gives a better picture of the crustal architecture and the tectonic processes that may have been associated with it.

**ACKNOWLEDGMENTS**

We would like to thank all the reviewers for carefully going through the manuscript and their constructive suggestions. We thank K. Vasudevan and R. Maier of the Lithoprobe Seismic Processing Facility at University of Calgary, B. Duna and J. Brunet from the University of Western Ontario and John Amor from the University of British Columbia for their technical assistance with the INSIGHT seismic processing package that we used in carrying out the conventional processing. We
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