TECTONICS, VOL. 18, NO. 3, PAGES 430-451, JUNE 1999

# Tectonic evolution of the Superior Boundary Zone from coincident seismic reflection and magnetotelluric profiles

D. J. White, A. G. Jones, and S. B. Lucas
Geological Survey of Canada, Ottawa, Ontario

Z. Hajnal
Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, Canada

**Abstract.** The Superior Boundary Zone (SBZ) forms the northwestern margin of the Archean Superior craton and constitutes a tectonic foreland of the ~1.8 Ga Trans-Hudson Orogen. The Superior Boundary Fault (SBF) separates the SBZ from the adjacent Reindeer Zone, a collage of Paleoproterozoic juvenile intraoceanic rocks. Lithoprobe seismic reflection and magnetotelluric data were acquired along two profiles crossing the SBZ in an attempt to better constrain the deformation and crustal geometry resulting from Trans-Hudson orogeny. Analysis and interpretation of spatially coincident regional seismic and magnetotelluric data acquired along the southern 200 km profile indicate the following: (1) the Reindeer Zone accretionary collage forms an east dipping, eastward steepening, crustal-scale tectonic stack of moderately conductive rocks near the SBZ. (2) The SBZ is characterized at shallow depths (< 5-6 km) by steep to moderately east dipping reflectivity that is associated with the limbs of third generation folds (F3 and D3) and east-side-up shear zones. At greater depth, the SBZ crust is highly resistive and is contiguous to the east with resistive crust beneath the Superior craton proper. (3) The SBF is recognized in the subsurface as an abrupt resistivity contrast between the Reindeer Zone and the SBZ, extending subvertically to 15 km depth. (4) Moderately conductive rocks of the Reindeer Zone extend eastward for 30 km beneath the SBZ at depths of 15-45 km. Seismic reflection data from a second crossing located 100 km NE along strike indicate a similar crustal structure with some notable exceptions: (1) The SBF is recognized in the subsurface as a truncation of interpreted collisional fabrics and extends subvertically to ~ 30 km depth. (2) There is no compelling evidence for the eastward continuation of Reindeer Zone lower crust beneath the SBZ. To explain the present-day SBZ crustal structure, we propose that the nature of the SBZ evolved over an ~200 Myr convergent margin history from a lower plate collisional thrust belt setting at > 1.88-1.81 Ga, through lithospheric delamination at ~ 1.82 -1.80 Ga to a steep transpressive plate boundary at 1.80-1.72 Ga.

## 1. Introduction

The Superior Boundary Zone (SBZ) forms the northwestern margin of the Archean Superior craton that was deformed during ~1.8 Ga continent-continent collision associated with the Hudsonian orogeny [*Bleeker*, 1990a, b; *Weber*, 1990]. The SBZ extends from south central North America northward to central Canada where it is exposed on the Canadian Shield (Figure 1) and continues to the northeast beneath Hudson Bay, emerging again in the Cape Smith Belt in northern Quebec [*Hoffman*, 1989]. Recently, the SBZ in northern Manitoba (Figure 1) has received considerable attention during Lithoprobe multidisciplinary studies, which have advanced our understanding of the role of the Superior craton in the Trans-Hudson Orogen (THO). Prior to these studies, the THO was considered to have resulted from continent-continent collision between the Archean Hearne and Superior cratons (Figure 1), trapping the intervening Reindeer Zone, a collage of juvenile intraoceanic rocks [cf. *Lewry and Collerson*, 1990; *Lucas et al.*, 1996a]. It has subsequently been recognized [e.g., *Lucas et al.*, 1993; *Lewry et al.*, 1994; *Hajnal et al.*, 1995] that a third Archean block (the "Sask craton" [*Ansdell et al.*, 1995]) played a prominent role in the collisional history of the orogen, with the Superior craton arriving only during the later stages of orogeny (~1.80 Ga).

Lithoprobe seismic reflection profiles 2 and 3 [*Lucas et al.*, 1993, 1994; *Lewry et al.*, 1994; *White et al.*, 1994] acquired in 1991 across the SBZ in northern Manitoba (Figures 1 and 2) show reflections dipping moderately beneath the SBZ to a depth of 25 km. This result, interpreted as juvenile Reindeer Zone rocks dipping beneath the SBZ, is in marked contrast to most previous tectonic models proposed for the SBZ [e.g., *Green et al.*, 1985; *Bickford et al.*, 1990; *Bleeker*, 1990a]. As well, it is particularly surprising in that better preserved sections of the Superior craton margin elsewhere indicate that the craton formed the lower plate above which a craton-verging fold and thrust belt developed, for example, Cape Smith Belt [*St. Onge and Lucas*, 1990], Belcher Belt [*Hoffman*, 1987], and the Fox River Belt [*Bleeker*, 1990a, b; *Weber*, 1990]. The principal evidence for Superior craton forming the lower plate in the Manitoba segment of the orogen is the absence of a continental margin (Andean) magmatic arc [*Lewry*, 1981]. New seismic reflection data from the Phanerozoic-covered portion of the SBZ in Manitoba suggest that a Superior-verging thrust belt developed prior to the formation of the east dipping, west verging structures [*Lucas et al.*, 1996b].

In this paper, we present reprocessed versions of the 1991 regional seismic reflection data (lines 2 and 3, Figures 1 and 2) along with new spatially coincident regional magnetotelluric data. The reprocessed seismic data have been migrated in a manner which preserves steeply dipping reflections in the vicinity of the SBZ and also properly accounts for data acquisition oblique to regional geological strike. In addition, regional and high-resolution seismic reflection data are presented from a second



Paper number 1999TC900002.
0278-7407/99/1999TC900002$12.00

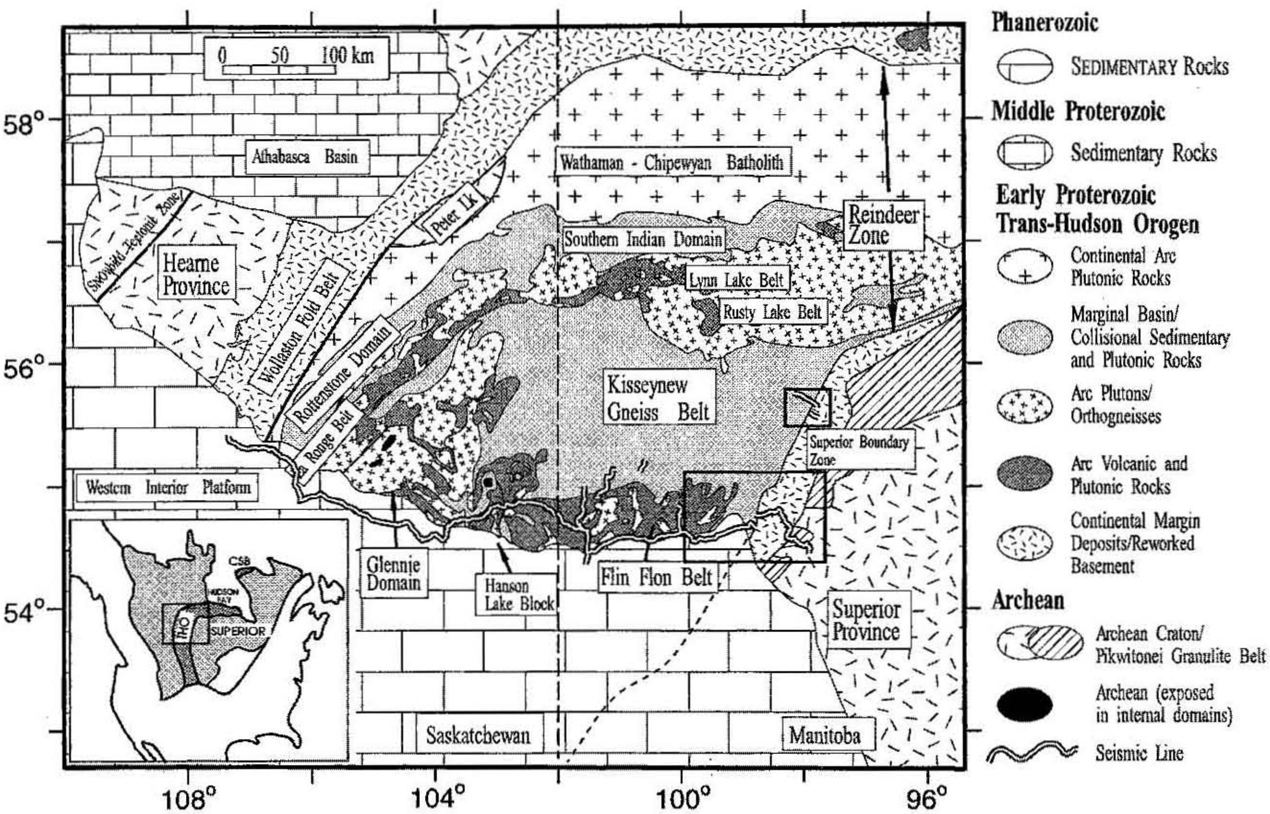


**Figure 1.** Geological map of the Trans-Hudson Orogen showing locations of the 1991 Lithoprobe seismic reflection lines. Study areas along the Superior Boundary Zone indicated by bold rectangles are shown in detail in Figures 2 and 3. Inset (lower left) shows location of map relative to North America. Abbreviations are as follows: THO, Trans-Hudson Orogen and CSB, Cape Smith Belt.

traverse across the SBZ located ~ 100 km along strike to the north (lines 1A and 1B, Figures 1 and 3) providing some measure of the structural variability along the margin. The high-resolution data are particularly important as they allow ground truthing of the cause of reflectivity within the shallow crust of the SBZ. The seismic and magnetotelluric data are interpreted in conjunction with new geological and geochronological constraints that have become available during detailed studies focused on this region over the past 5 years. We attempt to reconcile the apparently conflicting views of the SBZ provided by the surface geology and the recent seismic interpretations by developing a tectonic model that is consistent with both the geological interpretations and the present crustal structure as imaged seismically and electrically.

## 2. Geologic Setting

The Superior craton in the transect area (Figures 1-3) comprises the Gods Lake Domain and the Pikwitonei Granulite Belt, which together constitute an oblique crustal cross section [*Fountain and Salisbury*, 1981] with deeper crustal levels exposed successively northwestward. Gods Lake Domain consists primarily of 3.0-2.75 Ga granitoid rocks, with subordinate east trending greenstone belts [*Weber*, 1990]. Metamorphic grade within Gods Lake Domain increases northwestward from greenschist facies to upper amphibolite facies. The Pikwitonei Granulite Belt includes granulite metamorphic grade, mainly enderbitic orthogneiss and enderbite with minor amounts of high-grade equivalents of the Gods Lake Domain. Granulite-grade metamorphism within the Pikwitonei Granulite Belt is dated at 2.637-2.695 Ga [*Heaman et al.*, 1986], with closure of titanite (U-Pb) at 2.18 Ga. Uplift of the Pikwitonei granulites to upper crustal levels had presumably occurred by 2.17 Ga, the age of crosscutting dykes emplaced at shallow crustal levels [*Bleeker*, 1990a, b; *Weber and Mezger*, 1990; *Zhai and Halls*, 1994], perhaps in association with extensional thinning prior to rifting of the continent.

The SBZ comprises both Paleoproterozoic continental rift (~ 2.0 Ga) margin rocks (Ospwagan group) and reworked Superior craton gneisses [*Bleeker*, 1990a, b]. The eastern boundary of the SBZ is marked by a metamorphic/deformation front separating the E-W and NE-SW structural fabrics that characterize the Pikwitonei Gneiss Belt and the SBZ, respectively. The SBZ amphibolite-grade gneisses are, in part, derived from retrograded Pikwitonei granulites [*Weber*, 1990; *Machado et al.*, 1990] that were hydrated and structurally overprinted during the Hudsonian orogeny [*Bleeker*, 1990a, b]. The Superior Boundary Fault (SBF in Figures 2 and 3) generally forms an abrupt subvertical structural contact between the western SBZ and the adjacent Kisseynew Belt [*Bleeker*, 1990a, b]. Late, brittle-ductile and brittle transcurrent displacements have overprinted the SBF [*Bleeker*, 1990a, b].

The Kisseynew Belt and the Flin Flon Belt (Figure 2) of the easternmost Reindeer Zone lie immediately west of the SBZ. The Flin Flon Belt is a predominantly low-grade collage of 1.92-1.88

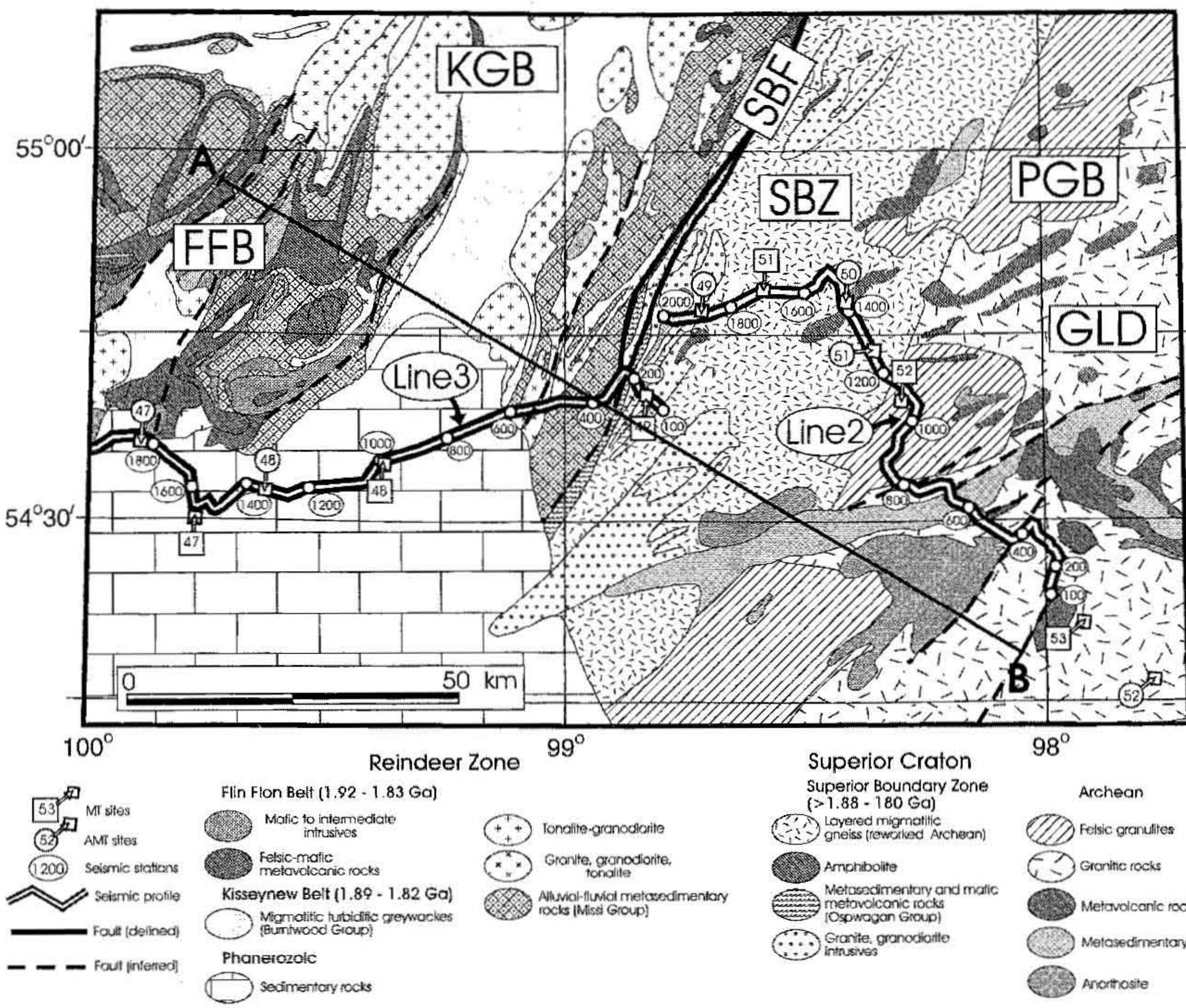


**Figure 2.** Location of the southern profile across the SBZ including seismic lines 2 and 3 and colocated magnetotelluric (MT)/audiomagnetotelluric (AMT) sites. See Figure 1 for context within the Trans-Hudson Orogen. Abbreviations are as follows: FFB, Flin Flon Belt; KGB, Kisseynew Gneiss Belt; SBZ, Superior Boundary Zone; SBF, Superior Boundary Fault; PGB, Pikwitonei Granulite Belt; GLD, Gods Lake Domain.

Ga arc and oceanic volcanic sequences (Amisk collage [*Lucas et al.*, 1996a]), stitched together by 1.88-1.83 Ga calc-alkaline plutonic rocks and unconformably overlain by 1.85 -1.83 Ga continental sedimentary rocks [*Lucas et al.*, 1996a; *Ansdell*, 1993; *David et al.*, 1996]. Rocks of the Amisk collage record the evolution from an oceanic to a continental setting for deformation and magmatism over a period of ~ 50 m.y. (1.90-1.85 Ga). The Kisseynew Belt is a high-grade complex of metasedimentary and metaplutonic rocks. Migmatitic, amphibolite-grade turbiditic greywackes (Burntwood Group) are predominant, with lesser amounts of alluvial-fluvial metasediments occurring along the southern (Missi Group) and eastern (Sickle Group) margins of the Kisseynew Belt [*Zwanzig*, 1990]. These metasediments were presumably shed primarily from the adjacent Flin Flon and Lynn Lake belts (Figure 1) [*Zwanzig*, 1990; *Ansdell et al.*, 1995] during 1.85-1.835 Ga [*Gordon et al.*, 1990; *David et al.*, 1996; *Ansdell and Norman*, 1995].

Detailed structural, geochronological, and geochemical studies within the eastern Reindeer Zone during the past few years have greatly advanced our understanding of how this region developed. Intraoceanic collisions at ~ 1.88-1.87 Ga produced an accretionary complex (Amisk collage) of regional extent that represented the early infrastructure of the Flin Flon Belt. Associated crustal thickening due to successor arc magmatism and concurrent deformation [*Lucas et al.*, 1996a] resulted in unroofing of the accretionary complex and deposition of 1.85-1.84 Ga Missi and Burntwood sediments in the Kisseynew basin, perhaps in a back arc setting [*Ansdell et al.*, 1995]. The Flin Flon and Kisseynew belts were likely separated from the Superior craton by an ocean basin at the time of Missi sedimentation (~ 1.85 Ga) as (1) ages of earliest Paleoproterozoic to latest Archean detrital zircons within the Missi and Burntwood metasediments (2.4-2.5 Ga [*Ansdell and Norman*, 1995]) are unknown in the Superior Province and (2) paleomagnetic poles indicate a 750 km longitudinal separation of

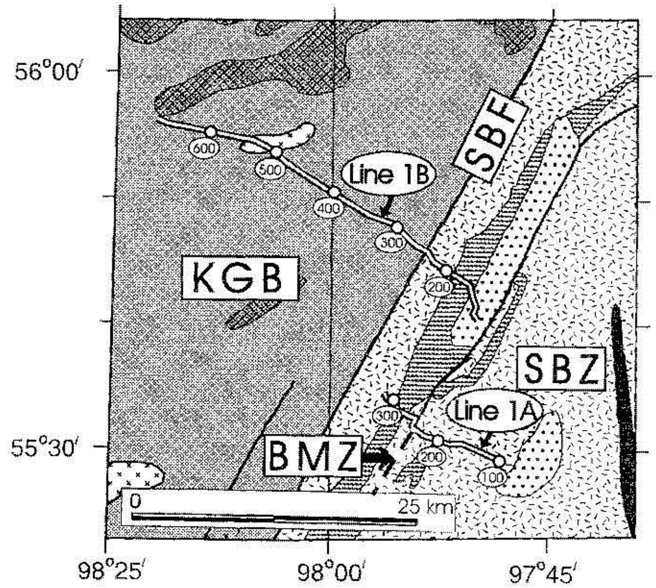


**Figure 3.** Location of the northern profile across the SBZ including regional seismic lines 1A and 1B. High-resolution lines (1AHR and 1BHR) are coincident with line 1A and the eastern part of line 1B. See Figure 1 for context within the Trans-Hudson Orogen and Figure 2 for geology legend and abbreviations.

the Flin Flon Belt and Superior craton at 1.85 Ga [*Symons et al.*, 1995].

Collision of the Flin Flon and Kisseynew belts with Sask craton at ~1.84 Ga is characterized by significant crustal shortening with regional metamorphism following with continued deformation at 1.82-1.805 Ga. Contractional deformation in the Kisseynew Belt included thrusting and nappe formation [*Zwanzig*, 1990, 1995] which persisted to peak metamorphic conditions (1.815-1.818 Ga [*Gordon et al.*, 1990; *Ansdell and Norman*, 1995]). Tectonic transport associated with both the prepeak and synpeak metamorphic deformation is generally thought to be top-to-the-southwest [*Lewry*, 1990; *Connors*, 1996]. Terminal collision by (1) synchronous peak metamorphism and anatectic granite intrusion (1.822-1.805 Ga [*Gordon et al.*, 1990; *David et al.*, 1996; *Machado*, 1990]), (2) synpeak to late peak metamorphic deformation associated with SE-NW compression within the eastern Kisseynew Belt [e.g., *Connors*, 1996], and (3) the close similarity of late metamorphic deformation in both regions [*Zwanzig*, 1990; *Bleeker*, 1990a, b].

The early deformation history of the SBZ is not as well preserved as in the adjacent Reindeer Zone but has been thoroughly documented by *Bleeker* [1990a, b]. Early deformation (F1-F2) consists of > 1.883-1.82(?) Ga nappe development and ductile deformation under medium-to-high-grade conditions interpreted as having occurred at midcrustal levels in an east vergent fold-and-thrust belt [*Bleeker*, 1990a, b]. This phase deformed both the Ospwagan group cover sequence and the underlying Archean gneissic basement rocks. The timing constraints on deformation come from Molson dykes (1.883 Ga [*Heaman et al.*, 1986]) which cut the earliest structures and peak metamorphism (1.820-1.809 Ga) which overprints F1-F2 deformation fabrics.

The third main deformation episode (F3) within the SBZ occurred at postpeak metamorphism (1.8-1.77 Ga) and produced upright, high-amplitude, doubly plunging folds that are observed throughout the SBZ. These folds have axial planes dipping at 65°-90° SE and trend 10°-20° clockwise relative to the N30°E trend of the SBZ. They deform the earlier recumbent F1-F2 structures into their present near-vertical attitudes. The en echelon pattern of these folds provides evidence for sinistral transpression of the SBZ during this stage of deformation, with associated assumed sinistral strike slip along the SBF [*Bleeker*, 1990a, b; *Weber*, 1990]. Widespread mylonitization associated with predominantly Superior-side-up ductile reverse faulting trending parallel to the SBZ was concurrent with but outlasted F3 folding [*Fueten and Robin*, 1989]. Postcollisional, intracontinental deformation in the vicinity of the SBZ appears to have occurred, at least episodically, until ~1.72 Ga [*Machado*, 1990], accommodating transpression due to oblique convergence between the Superior craton and the Reindeer Zone [*Gibb*, 1983; *Green et al.*, 1985; *Hoffman*, 1989; *Bleeker*, 1990a, b].

**Table 1.** Acquisition Parameters

| Parameter | Value: Regional | Value: High Resolution |
|---|---|---|
| Number of vibrators | 4 Hemi-50 | 2 Hemi-50 |
| Source array | 100 m, tapered | 10 m, tapered |
| Source station spacing | 100 m | 10 m |
| Number of sweeps | 8 | 2 |
| Sweep frequencies | 12-56 Hz | 30-130 Hz |
| Sample rate | 4 ms | 2 ms |
| Sweep length | 16 s | 12 s |
| Record length | 16.384 s correlated | 4.096 s correlated |
| Number of channels | 240 | 240 |
| Geophone group spacing | 50 m | 20 m |
| Geophones/group | 9 over 40 m | 9 over 20 m |
| Geophone frequency | 10 Hz | 30 Hz |
| Spread geometry | 60-180 split | 60-180 split |
| Nominal fold | 60 | 120 |
| Recording system | MDS-18 | MDS-18 |

## 3. Seismic Reflection and Magnetotelluric Profiles

In 1991 and 1992, Lithoprobe acquired complementary seismic reflection and magnetotelluric (MT) profiles across the SBZ in two locations separated along strike by approximately 100 km (Figure 1). These data were acquired as part of an orogen-wide seismic reflection/MT profile [cf. *Lucas et al.*, 1993; *Jones et al.*, 1993]. Seismic reflection data were acquired using both regional and high-resolution acquisition parameters (Table 1). The MT data were acquired using one of two recording frequency bands at each site, 10,000 Hz to 0.0005 Hz ("MT" sites) and 10,000 Hz to 10 Hz ("AMT" sites, for audio-MT). Seven component recordings were made at all sites, comprising three components of the magnetic field, two horizontal components of the electrical field, and two remote components of the magnetic field. The MT frequency band is designed to sense the resistivity structure from the near surface (50 m) to the upper mantle (100 km), whereas the AMT band senses only to the middle crust (25-30 km).

The first transect crossing the SBZ (Figure 1) is located adjacent to the exposed Flin Flon Belt and forms part of the primary orogen-wide Lithoprobe geophysical transect across the THO. Two seismic lines (lines 2 and 3, Figure 2) have been combined to provide a profile across the SBZ and the adjacent Kisseynew Belt, oriented at an acute angle of 40°-60° clockwise from the SBF. A total of 10 alternating MT/AMT sites are colocated along the line 2-line 3 profile shown in Figure 2. A gap in spatial coverage exists in the SBZ between MT sites 52 and 53 because of the presence of a high-voltage DC power line radiating energy at 2-4 Hz, which precluded observation of the natural signals for more than one decade in frequency on either side of this band because of sidelobe leakage.

The second transect crossing the SBZ (Figure 1) is in the vicinity of Thompson, Manitoba, located ~ 100 km to the north of the line 2-line 3 corridor. A 30 km seismic line (line 1B, Figure 3) crosses the SBF nearly perpendicularly and continues eastward for 6 km into the SBZ. The eastern 10 km of line 1B were also profiled using high-resolution parameters (line 1BHR). In addition, a second seismic line was acquired in this vicinity using both regional (line 1A) and high-resolution (line 1AHR) acquisition parameters. Line 1A/1AHR starts ~ 2 km east of the SBF and continues for 10 km to the east within the SBZ.

## 4. Southern Profile

### 4.1. Seismic Lines 2 and 3

The seismic data for lines 2 and 3 have previously been presented in small-scale line drawing format in a series of overview papers [*Lucas et al.*, 1993, 1994; *Lewry*, 1994; *White et al.*, 1994]. However, the original line 2 migrated seismic images suffered from enhanced migration noise in the deeper part of the section and elimination of east dipping reflections at the west edge of the section as a result of migration. To improve the seismic image in the vicinity of the SBZ, the original line 2 data have subsequently been reprocessed (poststack) and remigrated to preserve the prominent east dipping fabric observed in the unmigrated images. The processing stream applied to the seismic data is listed in Table 2. The remigrated line 2 data have been merged with the data from line 3 to form a continuous profile across the SBZ (Plate 1). Although the orientations of lines 2 and 3 differ by about 30° and they are offset along strike by 15 km, a comparison of these sections shows a good correlation of the major reflection package at 9-13 s and conformity of dips at lesser times, indicating the general validity of combining these lines. The seismic data presented here (Plates 1, 2 and 3 and Figures 7 and 8) are plotted with no vertical exaggeration for an assumed crustal velocity of 6.0 km/s, whereas a whole crustal average velocity of 6.6 km/s determined from refraction modeling (B. Nemeth, personal communication, 1998) is appropriate for this region.

**Table 2.** Processing Sequence

| Process | Parameter Value | |
|---|---|---|
| | Regional | High Resolution |
| Geometrical spreading compensation | | |
| Crooked line binning | 2000 m x 25 m bins | 2000 m x 10 m bins |
| Automatic gain control | 0.5 s | 0.2 s |
| Predictive deconvolution | 5-window | 4-window |
| | 24 ms p.d. | 2 ms p.d. |
| Refraction statics | 2-layer EGRM* | |
| Velocity analysis | 5500-7200 m/s | 5500-7200 m/s |
| Normal move-out correction | | |
| Residual statics | 0.5-10.0 s | 0.5-3.0 s |
| | +/-24 ms | +/-24 ms |
| First-break mutes | | |
| Trim statics | 0.5-8.0 s | 0.0-3.0 s |
| | +/- 20 ms | +/- 20 ms |
| Stack | 60-fold | 120-fold |
| Time-variant frequency filter | 20-55 Hz, 0-5 s | 40-130 Hz, 0-1 s |
| | 12-56 Hz, 5-16 s | 30-130 Hz, 1-4 s |
| Multiwindow automatic gain control (AGC) | | |
| Random noise attenuation | | |
| Migration | fk Stolt (6000 m/s) | fk Stolt (6000 m/s) |
| Coherency filter | | |

*Extended Generalized Reciprocal Method

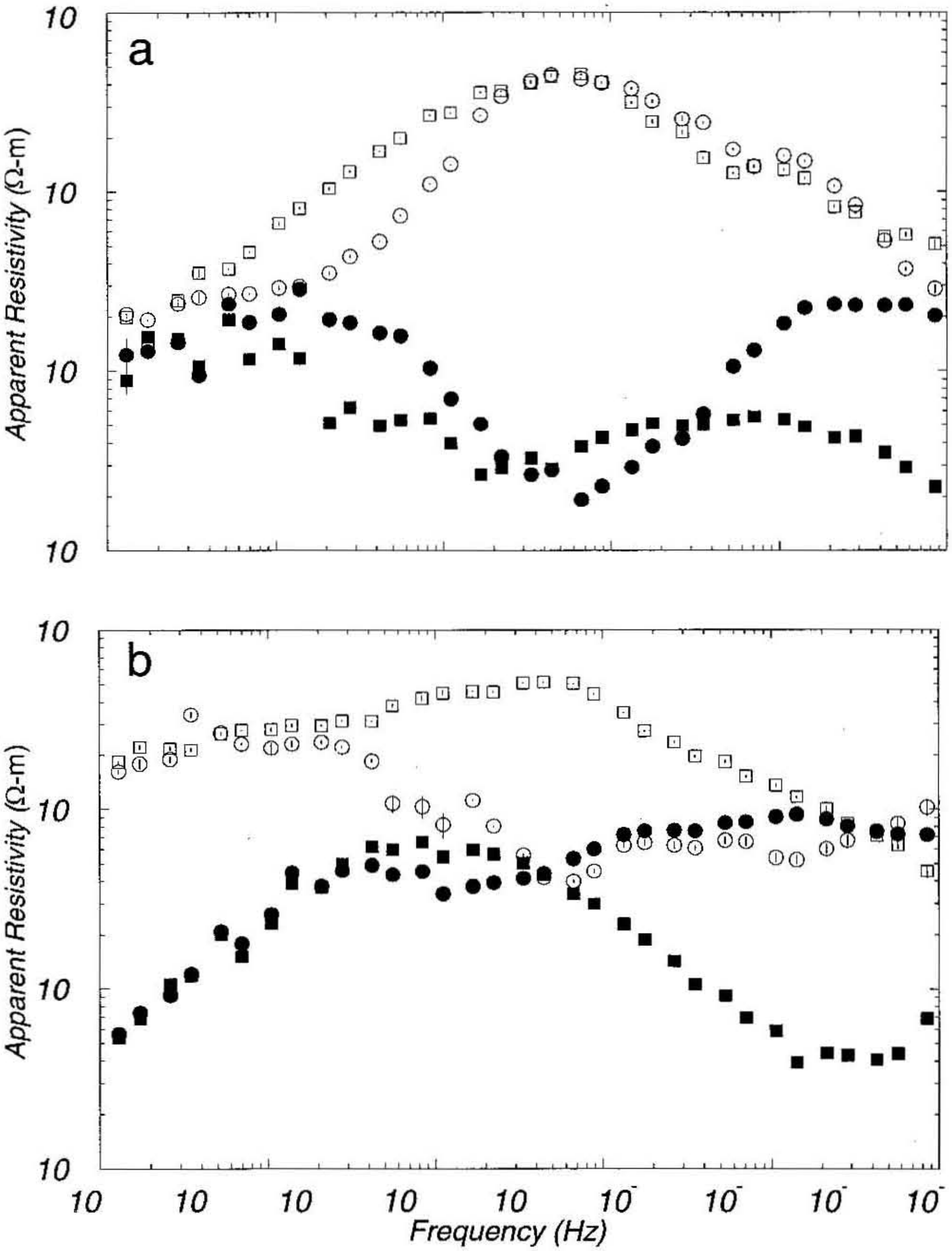


**Figure 4.** Apparent resistivity curves for (a) sites mt-53 (open symbols) and mt-47 (solid symbols) and (b) sites mt-49 (open symbols) and mt-48 (solid symbols). E-parallel mode data (circles) and E-perpendicular mode data (squares) are shown.

Thus depths taken directly from the plots for depths greater than 10 km will underestimate the true depths by a maximum of 8%.

The reflectivity on the composite line 2-line 3 seismic section (Plate 1) is characterized by a predominance of east dipping bands of reflections. In the following summary of reflectivity, true strikes or dips (where stated) have been estimated using the changes in reflection geometry that occur where there are directional variations in the seismic line. (1) In the lowermost crust, there is a planar band of reflections (A in Plate 1a-1c) with a true east dip of 20°± 5° and a true strike of N0°E ± 9°. The top of this band projects updip to the Namew Gneiss Complex within the western Flin Flon Belt (not shown in Figure 2) at the west end of line 3 [cf. *White et al.*, 1994; *Leclair et al.*, 1997], and at depth, between stations 800 and 1400, this zone appears to sole into the reflection Moho (M). Farther east, this zone may extend to greater depth. (2) The reflection Moho (M?) is tentatively identified on the unmigrated data (Plate 1a) at the east end of line 3, where there is an abrupt decrease in subhorizontal reflectivity at approximately 15 s, indicating a crustal thickness of > 50 km using an average crustal velocity of 6.6 km/s. (3) A second prominent band of east dipping reflectivity (B) is identified at the east end of line 3 (unmigrated, Plate 1a) and apparently continues beneath the western part of line 2. The reflections beneath the west end of line 2 move to beneath line 3 upon migration and have true dips of 45°± 7°. The true strike of the reflections is N45°E ± 12° which is slightly east of the strike of the SBF trace. (4) A third prominent band of east dipping reflectivity (C) has an apparent dip of ~ 40° when migrated (Plates 1b and 1c), with some discordant reflections having apparent dips > 60°. This band of reflections projects to the surface in a zone that straddles the SBF and extends into the easternmost Kisseynew Belt. The near-surface truncation of these reflections is an artifact of merging the data from lines 2 and 3. (5) The band of reflections C intersects, or in some cases, soles into a zone of reflections (D) which have shallow eastward apparent dips at midcrustal depths (4-9 s) and can be traced eastward from within 10-15 km of the SBF to the eastern boundary of the SBZ. (6) An east dipping fabric is observed in the shallow crust (E at < 4 s or 12 km) above the C reflections across the map width of the SBZ. The true strike of these reflections is N43°E ± 9°. (7) The shallow crust between the bands B and C is generally diffractive, comprising discrete packages of reflections with both east and west apparent dips, although east dipping reflections predominate. (8) The lower crust is poorly reflective, and the reflection Moho is not distinguished along all of line 2, although this may be due in part to reduced energy penetration in this area. Weak reflections (F) in the middle-to-lower crust beneath the eastern SBZ and the Pikwitonei Granulite Belt have shallow apparent dips to the NNW. (9) East of the SBZ, the shallow crust is less uniformly reflective and is characterized by reflections (G) with NW apparent dips that generally correlate with major shear zones (probably Archean in age) mapped at the surface in the Archean Cross Lake domain (Figure 2). The true geometry of these reflections (especially after migration) is uncertain because of the crooked nature of the eastern part of line 2. (10) The upper crustal zone of NW dipping reflectors (G) is underlain by a band of reflections (H) between 5 and 8 s that are subhorizontal at the east end of the line but shallow to the NW where they project to the surface near the boundary between the SBZ and Pikwitonei Granulite belt [cf. *White et al.*, 1994]. At the east end of the line 2, these reflections are truly subhorizontal.

To obtain a better representation of the true dip of the reflectors in the vicinity of the SBZ (Plate 2), a psuedo-three-dimensional (3-D) migration has been applied to a subset of the data accounting for the obliquity of the acquisition line (N80°E-N85°E) relative to the predominant geological strike direction (N30°E± 5°) near the SBF. The migration procedure moves each reflector to its true position in three dimensions using the assumed strike direction (i.e., the surface-mapped geological strike) and then projects the reflector on to a vertical plane which is perpendicular to the geological strike (i.e., a dip plane). True strike information provided by the seismic data indicates that both reflection packages B and E have strikes (N43°E±9° and N45°E±12°, respectively) that are subparallel with the surface strike of the SBF, whereas the lower crustal reflection package A has a strike of ~ 0°. The true strike of reflection package C is unknown from the seismic data but is assumed to be consistent with packages B and E. Thus the reflection geometry shown in Plate 2 is appropriate for all of the reflection packages shown except for A which is oversteepened by the process. On the remigrated line drawing (Plate 2), the reflections in the upper crust adjacent to the SBZ are much steeper (< 80°) than on the original migrated data (< 45°, Plate 1b) and are more consistent with the steep dips (~ 65°-85°) mapped at the surface in this area.

### 4.2. MT Data Analyses, Modeling, and Inversion

The MT time series were reprocessed subsequent to acquisition using a variant of the *Jones and Jödicke* [1984] robust scheme. The responses derived were then analyzed for galvanic distortion using the Groom-Bailey approach extended for multiple sites and multiple frequencies [*McNeice and Jones*, 1996]. Analyzing data in the frequency band of 1000-0.001 Hz from all sites simultaneously in frequency bands of half a decade, a geoelectric strike of ~N20°E-N40°E is observed in general agreement with the geological strikes observed in the vicinity of the SBZ. Accordingly, the regional responses were derived by fitting the multifrequency extended Groom-Bailey decomposition model to the data in the frequency band 100-0.1 Hz with an assumed strike of N40°E. Frequency-independent galvanic distortion parameters were determined for each site, and the two apparent resistivity curves were coalesced at high frequencies to deal with local anisotropy, leaving only the individual site gains as unknowns. In the adopted 2-D reference frame, the data with the electric field in the direction N40°E describes the response of Earth for currents flowing along the structure and is called the "E-parallel" mode of induction, whereas the data with the electric field in the direction N130°E describes the response for currents flowing across the stucture, called the "E-perpendicular" mode.

The regional responses from two sites, mt-47 and mt-53, on either side of the SBF are shown in Figure 4a. At site mt-53 to the east of the SBF, both apparent resistivity curves rise from around 2000 Ω m at the highest frequencies to around 30,000 Ω m at 3 Hz, and both phase curves (not shown) are below 45° at all frequencies above 1 Hz. At site mt-47, to the west of the boundary, the two curves are consistent with the mt-53 response to 200 Hz (~ 2 km depth). At lower frequencies, the apparent resistivity curves decrease with decreasing frequency, and the phases (not shown) lie above 45° for much of the frequency range. This extreme difference in electrical response is indicative of a very large difference in electrical resistivity structure beneath the two sites below about 2 km depth. Shown in Figure 4b are the apparent resistivity curves for the two sites that straddle the SBF, sites mt-48 (west) and mt-49 (east). The pronounced difference in response at high frequencies (>1 Hz) on either side of the SBF is indicative of contrasting crustal structure. Note that at low frequencies, the E-parallel mode data (circles) are consistent with each other, whereas the E-perpendicular mode data remain apart, as required by the physics of a strong vertical contact.

Figure 5 shows the apparent resistivities and phases along the profile in pseudosection format. Qualitative interpretation of the MT phases is more reliable than interpretation of the apparent resistivities because the phases are not affected by "static shifts" for each mode. Figure 5 shows low phases, <30°, at high frequencies at all of the eastern sites, indicative of a highly resistive upper and middle crust. In contrast, most sites show high phases (>45°) in both modes at frequencies which are sampling the deep crust (<1 Hz), representative of a more conducting lower crust. Note the strong contrast in high frequency E-parallel phase response (100 - 9 Hz) between sites amt-49 and mt-51 (Figure 5a). Site amt-49 phases are in the range 50° - 60°, whereas site mt-51 phases are in the range 30° - 40°, requiring a strong contrast in resistivity between these sites.

Objective modeling of these data was undertaken using the rapid relaxation inverse (RRI) MT inversion scheme of *Smith and Booker* [1991] which attempts to find the smoothest resistivity structure (i.e., a least structure model) that will fit the data to within error tolerances. An error floor of 4% in phase (1.14°) and 10% in apparent resistivity was applied, with the larger value for apparent resistivity chosen to compensate for the final static shift factors which remain after all other scaling effects have been removed. To aid the inversion scheme, three fictitious sites were

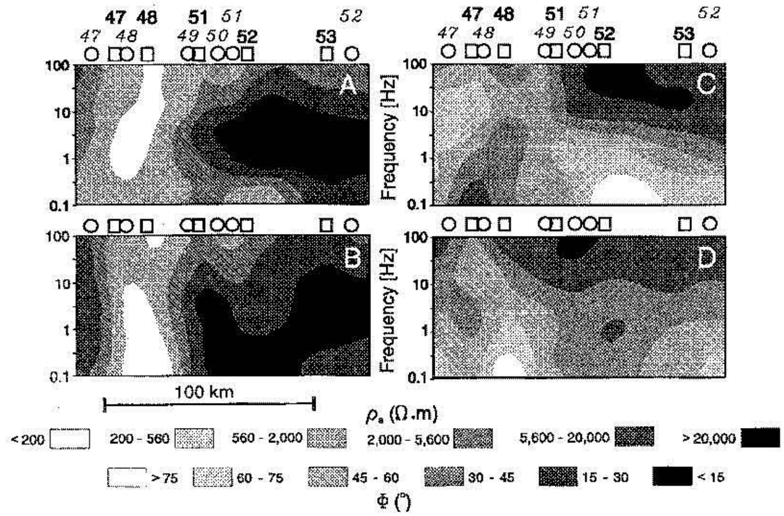


**Figure 5.** Pseudosections of apparent resistivity ($\rho_a$) and magnetotellurc (MT) phase response ($\phi$). The apparent resistivities and phases (a) and (c) for the E-parallel mode and (b) and (d) for the E-perpendicular mode of induction. High phases (above 45°) are indicative of a more conducting region at depth, whereas low phases suggest a more resistive region at depth. Circles and squares denote audiomagnetotelluric and magnetotelluric sites, respectively.

introduced in the region of the DC power line between sites mt-52 and mt-53. A discontinuous jump in resistivity was permitted without penalty at the approximate depth of the crust-mantle boundary of 44 km. Although the MT data do not require such a jump, it was determined that its inclusion led to models having a slowly varying upper mantle conductivity. In addition, the site gain for site mt-48 was solved for as part of the inversion.

Data at 20 frequencies in the range 96 - 0.1 Hz were used from the MT sites, and eight frequencies in the range 96 - 9 Hz from the AMT sites were used for 18 sites for a total of 1152 data (four data per frequency per site). The crustal portion of the final model is compared with the coincident seismic data in Plates 1b and 2. This resistivity model fits the data to within a normalized RMS error of 6.5, which means an average phase misfit of 7.5°. A significant fraction of this phase misfit (>25%) is concentrated at the sites at the boundary, amt-49 and mt-51, and results from the program's attempts to produce a smooth model in a region of large resistivity contrast. Permitting an unpenalized discontinuity in resistivity from the surface to 20 km depth reduces this misfit to levels comparable to that at other sites. The model with this discontinuity is the same as the model shown in Plate 1b with the minor difference that there is a sharp vertical boundary between sites amt-49 and mt-51 to a few kilometers depth. The existence of this boundary is consistent with the MT data; although not formally required by it, its inclusion reduces the misfit levels for sites amt-49 and mt-51.

### 4.3. Interpretation

Interpretation of the seismic data is based on the identification of regional variations in reflectivity and their spatial correlation with surface lithologic units and structures and with the electrical conductivity structure (where available). In some cases, individual reflections project to surface in the vicinity of well-mapped structures of known age providing well-constrained interpretations. In other instances, the relative age of discordant seismic fabrics can be inferred by cutoff relationships, but these are rarely well defined. In the absence of such specific information, we implicitly assume that the broad-scale reflectivity pattern is determined primarily by the last major episode of regional deformation.

The combined interpretation of the seismic data and the resistivity model is shown in Plate 1c and Figure 6. West of the SBZ, a crustal-scale east dipping imbricate zone [cf. *White et al.*, 1994] is imaged (Plates 1b and 1c) which generally corresponds with a zone of moderate resistivity (< 1000 Ω m, Plate 1b). The planar band of reflections (A) observed in the lower crust has been interpreted as the eastward continuation of a crustal-scale detachment [*Lucas et al.*, 1993, 1994; *White et al.*, 1994] that projects to the surface as the Namew Gneiss Complex [*Leclair et al.*, 1997]. Structurally above A lies a second reflective band B interpreted as flooring an imbricate stack of rocks in the upper crust associated with the eastern Flin Flon and Kisseynew belts. The upper crust within the Kisseynew Belt along line 3 (Burntwood and Missi groups) is characterized by relatively steep reflections without significant lateral continuity. The bottom of this unit indicated in the interpretation (i.e., the base of the orange unit in Plate 1c) is not based on observed reflections or the resistivity pattern but is included to indicate that the metasedimentary rocks of the Kisseynew Belt are likely relatively thin [cf. *Lucas et al.*, 1994]. The shallow conductivity anomaly beneath sites mt-47 and amt-48 is of a local nature as its effects are not visible in the data from the sites on either side (amt-47 and mt-48). Its cause is undetermined as it lies below the Phanerozoic cover.

The combined interpretation of the seismic and MT data indicates that the SBF is subvertical to a depth of ~ 15 km (Plate 1c and Figure 6). The MT data require an abrupt westward decrease in resistivity from mt-51 to amt-49 (Plate 2). The shallow conducting anomaly beneath site amt-49 is a local feature possibly due to occurences of metapyroxenite [*Manitoba Energy and Mines*, 1993]. This anomaly is not present in the data from site mt-49, attesting to its local nature. At greater depth, the transition occurs from very high resistivities (>10,000 Ω m ) characteristic of the Superior Craton farther east to intermediate resistivities (<3000

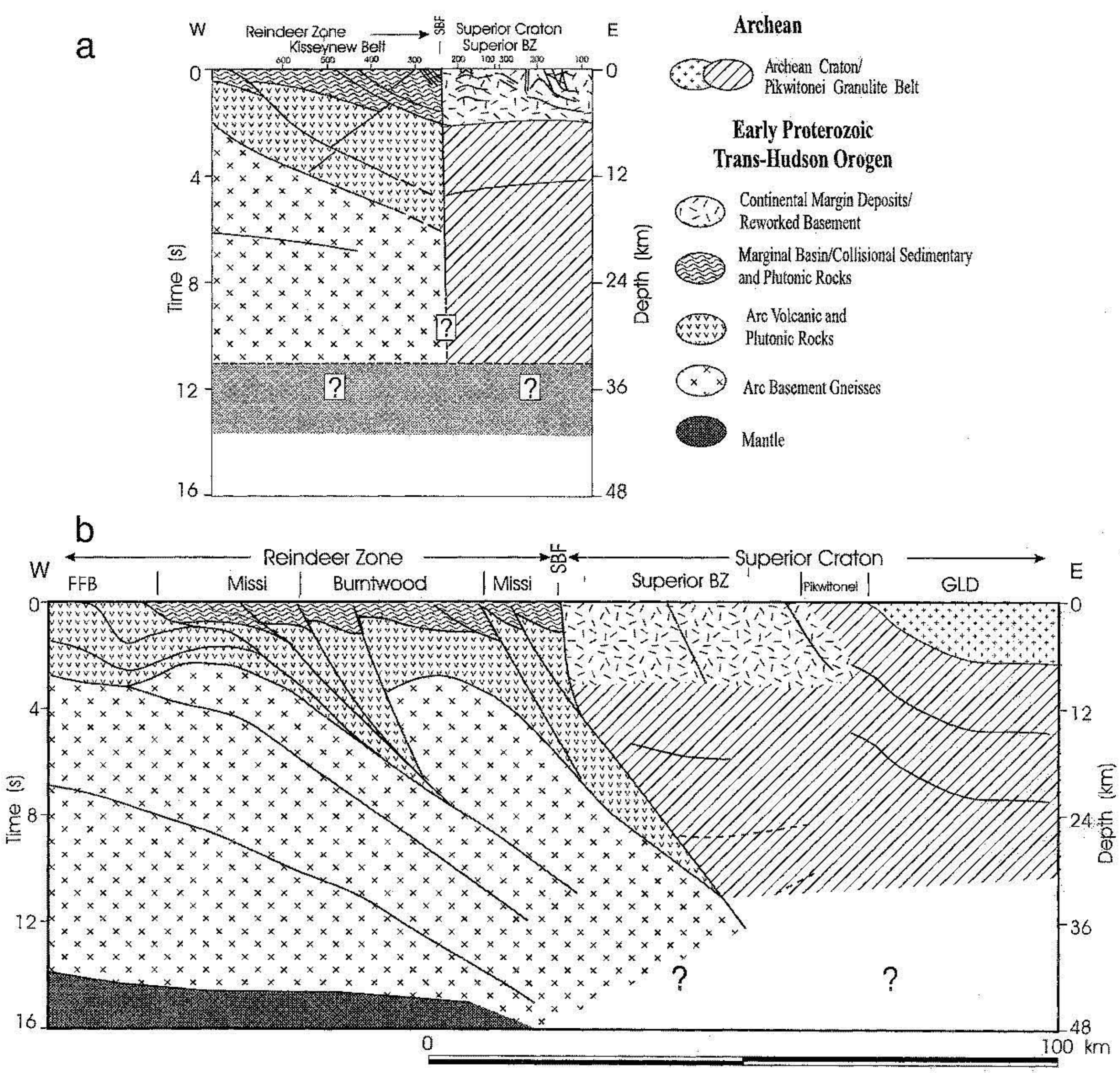


**Figure 6.** Interpretation cartoons for (a) the northern (line 1A/1B) and (b) southern (line 2/3) SBZ profiles. The cartoon in Figure 6b represents the crustal structure interpreted in Plate 1c after projection on to the line A-B (shown in Figure 2) which is perpendicular to the local geologic strike.

Ω m) associated with the eastern Reindeer Zone. This large contrast in resistivity (Plate 1b) is observed to a depth of at least ~ 15 km, clearly identifying contrasting material properties of the SBZ (i.e., Superior craton) and the Reindeer Zone across the SBF. The steeply east dipping reflective band C is interpreted as a band of shear zones that have accommodated Superior-side-up displacement in analogy with similar structures that are well known within the SBZ (see below). The true dip of the C reflections (< 85 ° east) is consistent with the dips observed at the surface within the westernmost SBZ and easternmost Reindeer Zone [*Weber*, 1990; *Manitoba Energy and Mines*, 1993]. The SBF likely lies within this reflective band as the easternmost reflections within this band project to the surface in the vicinity of the SBF.

Continuity of low resistivities from the Reindeer Zone beneath the SBZ at depths of >15 km indicates that the subhorizontal midcrustal reflections (D in Plates 1 and 2) are associated with Reindeer Zone lithologies extending beneath the SBZ.

The upper crustal east dipping fabric (E in Plates 1 and 2) in the SBZ, which has previously been inferred to be "Hudsonian tectonic overprint" [cf. *Lucas et al.*, 1993; *White et al.*, 1994], is most likely associated with F3 folds and associated D3 Superior-side-up ductile reverse faults that are documented throughout the SBZ in this area [*Fueten and Robin*, 1989; *Bleeker*, 1990a, b] and in the subsurface from borehole information [cf. *White et al.*, 1997]. At the surface, these structures generally strike 10°-20° clockwise of the SBF strike (~N30°E), with a mean dip

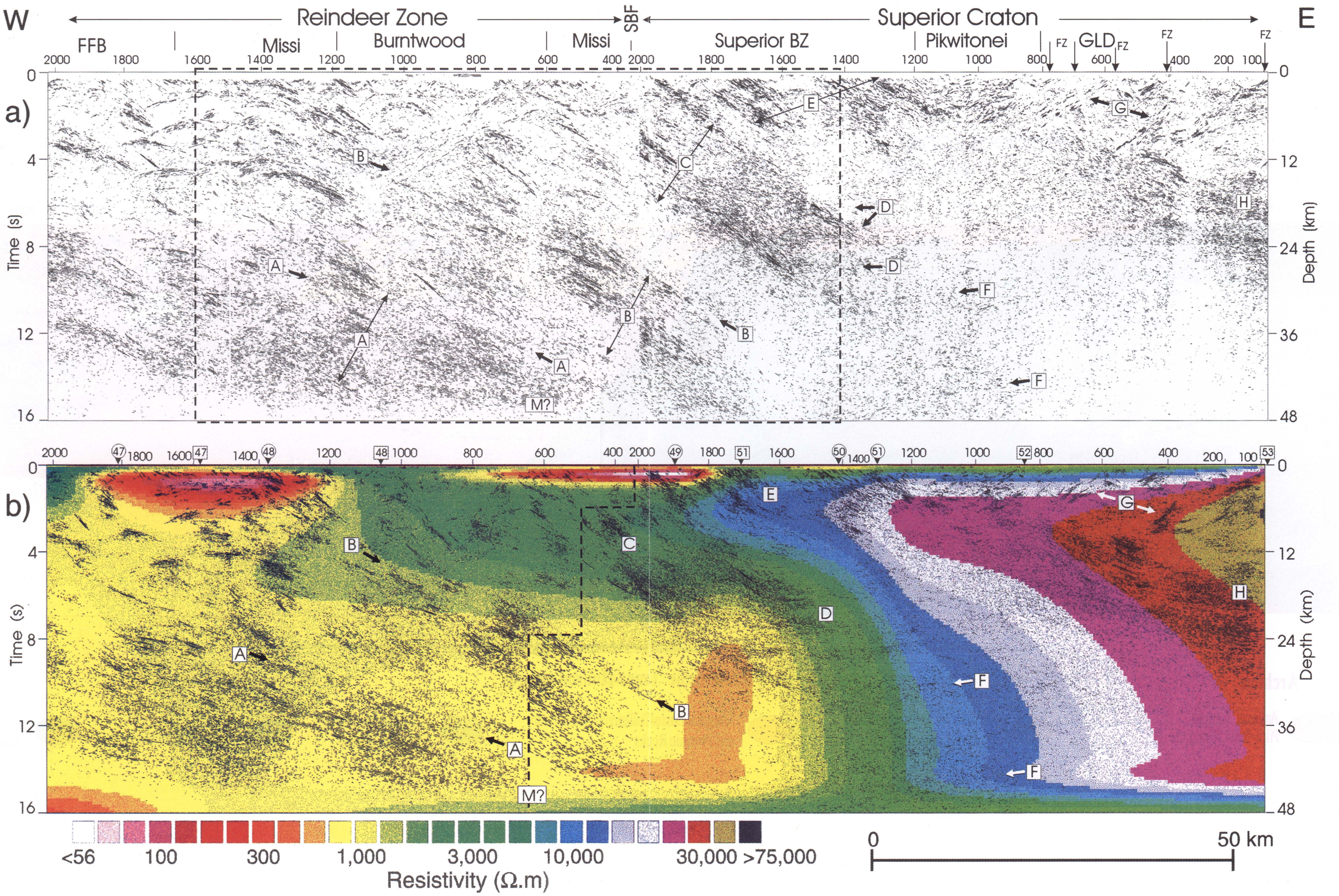


**Plate 1.** Merged seismic images and resistivity model for the southern profile (for location, see Figure 2). (a) Unmigrated seismic data. Data from lines 2 and 3 have been merged at stations 2000 (line 2) and 350 (line 3). The dashed box indicates the profile segment of Plate 2. (b) Migrated seismic data with the crustal portion of the final least structure resistivity model overlain for comparison. The seismic data from lines 2 and 3 were migrated prior to merging; the stitching line is indicated by the irregular dashed line. Merging included a region of overlap (approximately stations 350-650, line 3) due to the westward migration of east dipping reflections from line 2. Owing to their along strike separation of approximately 15 km, some latitude was allowed in how the sections were merged. An attempt was made to provide the best match of the prominent reflectivity in the middle crust (B). (c) Migrated seismic data with interpretation overlain. The steep geometry of the SBF indicated in the interpretation is based on the resistivity model and the corrected seismic data shown in Plate 2. FZ is fault zone; other labels are referred to in the text.

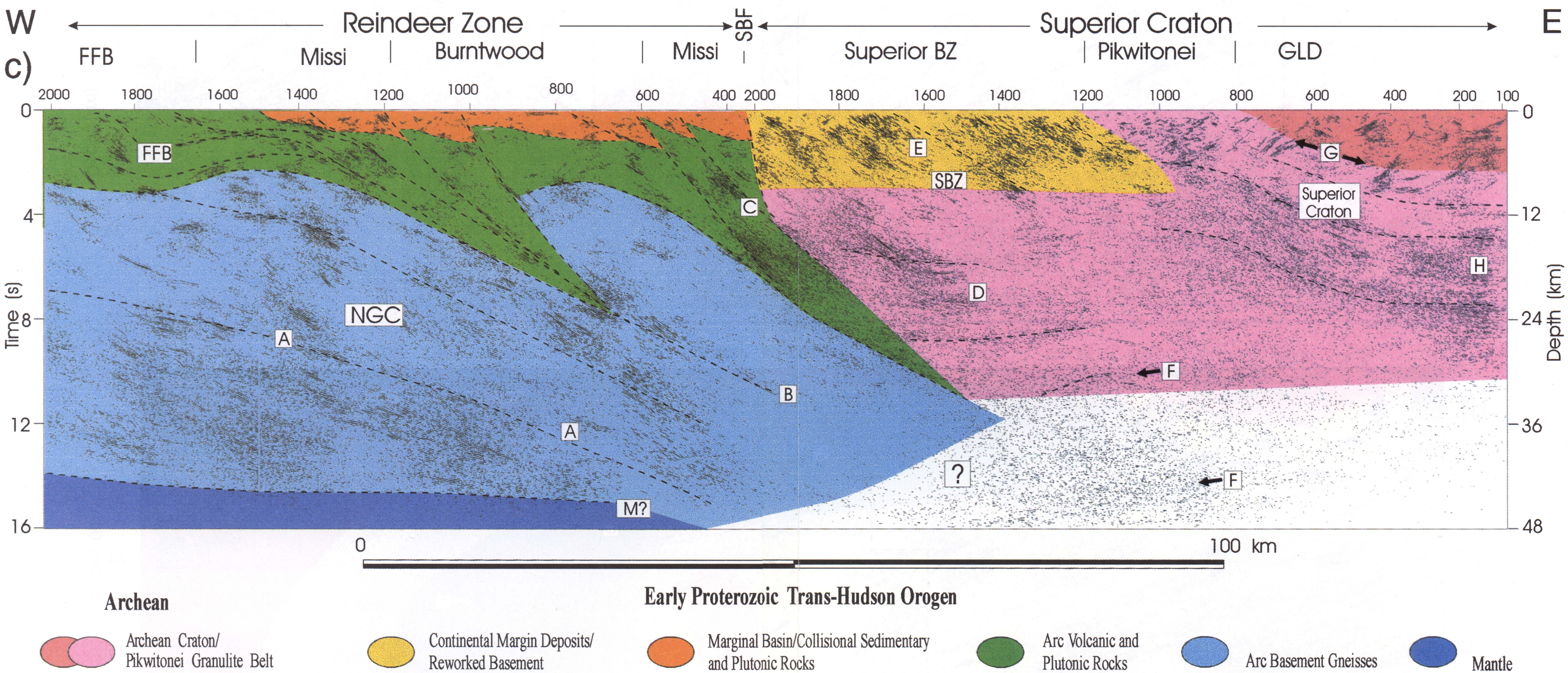


**Plate 1.** (Continued)

of ~45°-65°SE for F3 fold axial planes and ~80°-85°SE for syn- to post-F3 mylonites. The true dips of the E reflections (45°-60°, Plate 2) and their strike (N20°E), determined seismically, compare well with the orientations of the F3 structures. The zone of D3 folding/faulting inferred from the seismic data is not restricted to the SBZ but appears to extend westward into the Reindeer Zone.

At lower crustal depths, Reindeer Zone rocks are inferred to dip eastward toward the Superior craton (B in Plates 1c and 2, and Figure 6) as far east as the surface trace of the SBF. The associated low-resistivity zone can be extended at least 40 km east of the SBF (Plate 1b). The eastward continuation of the Reindeer Zone beneath the SBZ, as indicated by the MT data, is consistent with earlier interpretations [e.g., *Lucas et al.*, 1993; *White et al.*, 1994] that concluded that rocks of the Reindeer Zone formed a crustal wedge that delaminated the Superior crust at lower crustal levels.

Within the Superior craton proper (common depth point reflection stations 100-1200, line 2, Plate 1), the seismic image shows an upper crustal domain (west dipping reflections G) overlying a westward shallowing zone (reflective band H). The midcrustal seismic fabric (H) that shallows westward from ~ 12 km depth to the near surface is consistent with this region representing an oblique crustal section as argued by *Weber and Mezger* [1990]. Although this seismic signature could be interpreted as representing a thrust zone of pre-Hudsonian or Hudsonian age [e.g., *Lucas et al.*, 1993; *White et al.*, 1994], we interpret it as an Archean crustal fabric (originally horizontal) that was subsequently exhumed to the west primarily during late Archean deformation [cf. *Lucas et al.*, 1996b] and/or Proterozoic rift-related extensional thinning. The Pikwitonei uplift would have included the present-day SBZ which was subjected to Hudsonian retrograde metamorphism at ~1.8 Ga. We prefer this model as (1) it satisfies the above timing constraints, (2) it readily explains isolated occurrences of (unretrogressed) granulite-grade metamorphism within the SBZ, (3) the ~ 12 km of structural relief for the Pikwitonei uplift inferred from the seismic profile is generally consistent with that inferred from the surface geology [cf. *Weber and Mezger*, 1990], and (4) analogous east dipping seismic fabrics are observed well into the Archean Superior Province along more recent seismic data acquired 200 km to the south [*Lucas et al.*, 1996b].

## 5. The Northern Profile

### 5.1. Regional Lines 1A/1B

Lines 1A and 1B (Figure 3 and Plate 3) form a composite line that extends from the Kisseynew Belt across the SBF into the SBZ. The upper crust within the Kisseynew Belt is characterized by an erratic pattern of reflectivity (I in Plate 3) with both east and west apparent dips. Prominent upper crustal east dipping reflectivity is observed in the SBZ along line 1A (J in Plate 3) and the high-resolution lines (N and P in Figure 7) where it is prominent down to at least 2.0 s. At 5-9 s (Plate 3), there is an abrupt lateral change in reflectivity with strong associated diffractions (K) that originate beneath the surface location of the SBZ and extend beneath the Kisseynew Belt. After migration, these reflections and collapsed diffractions lie completely beneath the SBZ (Plate 3b). In contrast to the southern crossing, there are no prominent east dipping reflections in the lower crust. The reflection Moho is not readily identified, indicating that the base of the crust is either not a distinct reflector or it resides at times >16 s.

### 5.2. Interpretation

The large-scale crustal geometry observed along the northern crossing of the SBZ (Plate 3) is fundamentally different than that observed along the southern crossing, although there are some crustal features common to both. In the SBZ upper crust, the depth extent of the prominent east dipping fabric (J in Plate 3), associated with F3 folds and associated fault structures (see below), indicates a similar depth (at least 6 km) of Hudsonian (~1.8 Ga) tectonic overprint on Superior crust to that observed to the south. The upper crustal reflectivity pattern within the adjacent Reindeer Zone is similar to that observed at the east end of line 3, although there appears to be more evidence for both east and west dipping reflections along line 1B. Interpretation of these oppositely dipping reflections is difficult because of the lack of surface geological information in this area and the uncertainty of obvious geometrical discordances that would suggest timing relationships. The east dipping reflections are interpreted as east-over-west collisional structures analogous to those observed along the southern profile, whereas the significance of the west dipping structures is uncertain.

The SBF is interpreted as a vertical structure extending from the surface to ~ 30 km depth. This interpretation is based on the abrupt lateral change in crustal reflectivity observed between the SBZ and the adjacent Reindeer Zone on both the regional (1B in Plate 3) and high-resolution (1BHR in Figure 8) seismic images and the prominent diffractions (downward concave portion of K, Plate 3) that result from this truncation of west dipping reflectors (also K in Plate 3). It is notable that the SBF does not appear to penetrate the entire crust but rather is underlain by subhorizontal reflections below 10 s within a lower crustal layer of unknown affinity. The age of the SBF is suggested by the fact that it truncates Reindeer Zone and SBZ collisional fabrics and thus appears to be postcollisional in age (< 1.80 Ga) consistent with geological evidence [cf. *Bleeker*, 1990a, b]. This interpretation is consistent with the seismic signatures of other steep intracontinental faults (for example, Tabbernor Fault [*Hajnal et al.*, 1995]) or plate boundary faults (for example, Walls Boundary Fault [*McGeary*, 1989] and San Andreas Fault [*McBride and Brown*, 1986]).

The notable difference between the northern (lines 1A and 1B) and southern (lines 2 and 3) profiles is the absence of prominent east dipping reflectivity within the middle and lower crust of the Reindeer Zone and SBZ along the northern profile. This suggests that in contrast to the southern profile, rocks of the Reindeer Zone do not continue eastward beneath the SBZ with moderate dips at middle crustal depths. Instead, reflections at 8 s are west dipping beneath the SBZ and are subhorizontal below the Reindeer Zone. However, we cannot entirely eliminate the possibility that there are steep east dipping midcrustal reflections due to the short length of lines 1B and 1A.

### 5.3. High-Resolution Lines 1AHR/1BHR

Line 1AHR (Figure 7), which lies entirely within the SBZ, is characterized primarily by east dipping reflectivity in addition to two prominent bands of arcuate reflections (L and O). The arcuate reflection band L that extends from about 0.3 to 0.9 s is apparently truncated to the west, in the vicinity of the Burntwood mylonite zone, by a less reflective zone (N) of east dipping reflectivity. Both L and O are underlain by an east dipping zone (P). When migrated, the arcuate reflective bands (L and O) collapse into tight

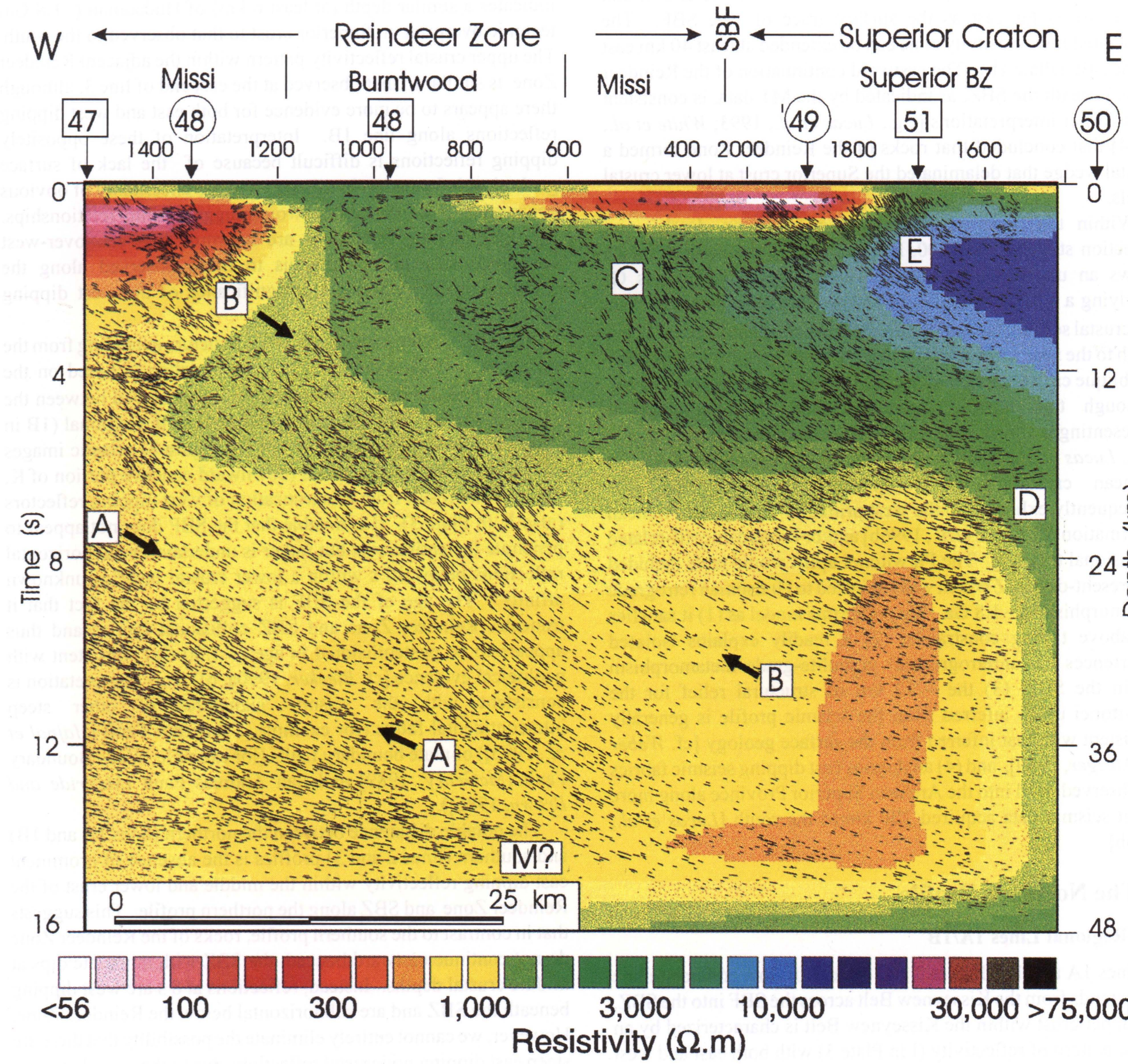


**Plate 2.** Seismic data corrected for acquisition oblique to strike with the final resistivity model overlain for comparison. A line-drawing of the data from the eastern end of line 3 and the western end of line 2 has been corrected assuming a geological strike of N40°E and acquisition lines oriented at N80°E and N85°E, respectively.

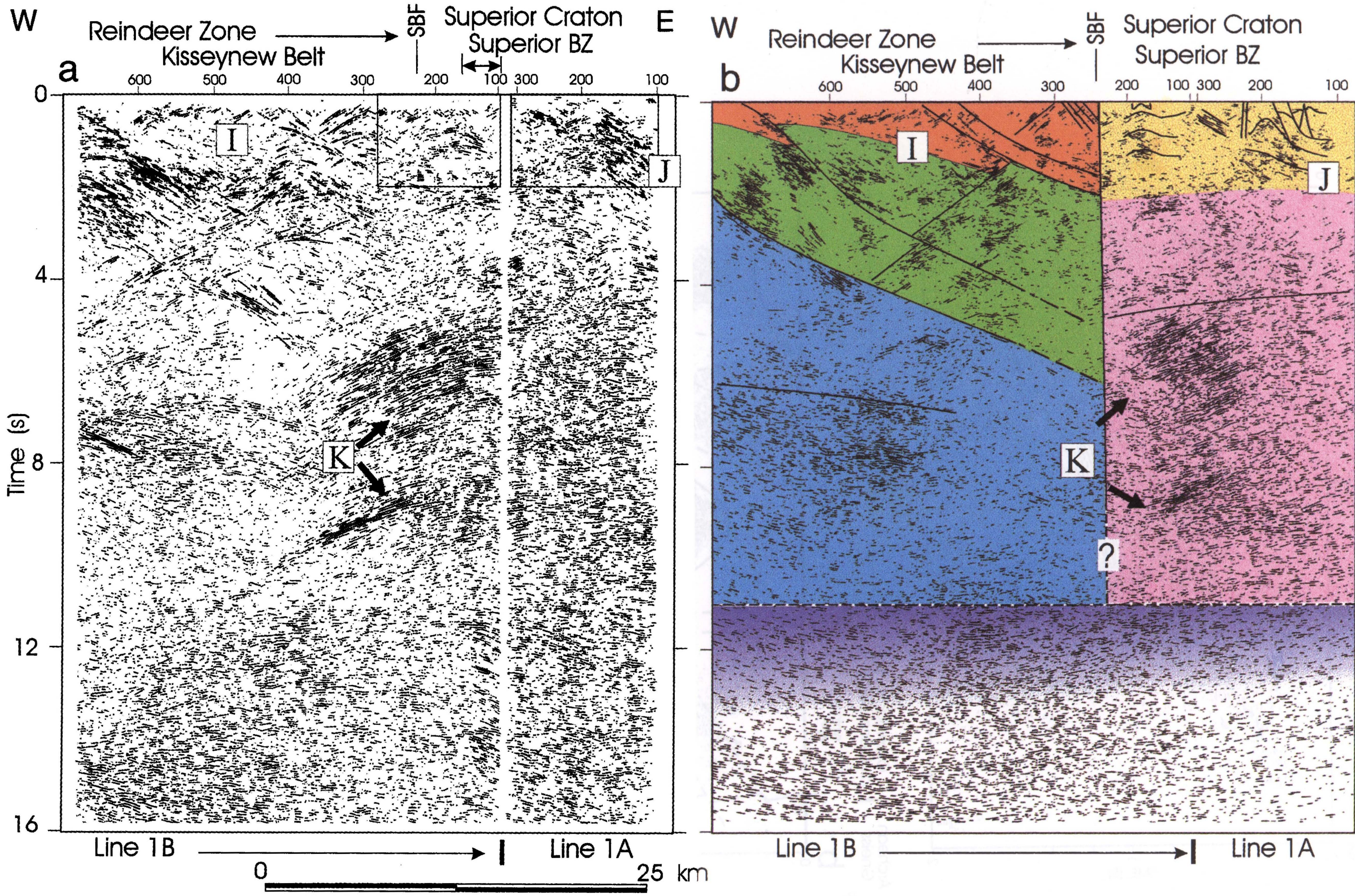


**Plate 3.** Seismic images from the northern profile, lines 1A and 1B. (a) Unmigrated data. (b) Migrated seismic data with interpretation overlain. The data were merged prior to migration. Owing to their along-strike separation of approximately 10 km, some latitude was allowed in how the sections were merged. An attempt was made to provide the best match of the prominent reflectivity in the middle crust (K). Bold rectangles in Plate 3a indicate location of the corresponding high-resolution data shown in Figures 7 and 8. Labels are referred to in the text. Colors are defined in the legend of Plate 1c.

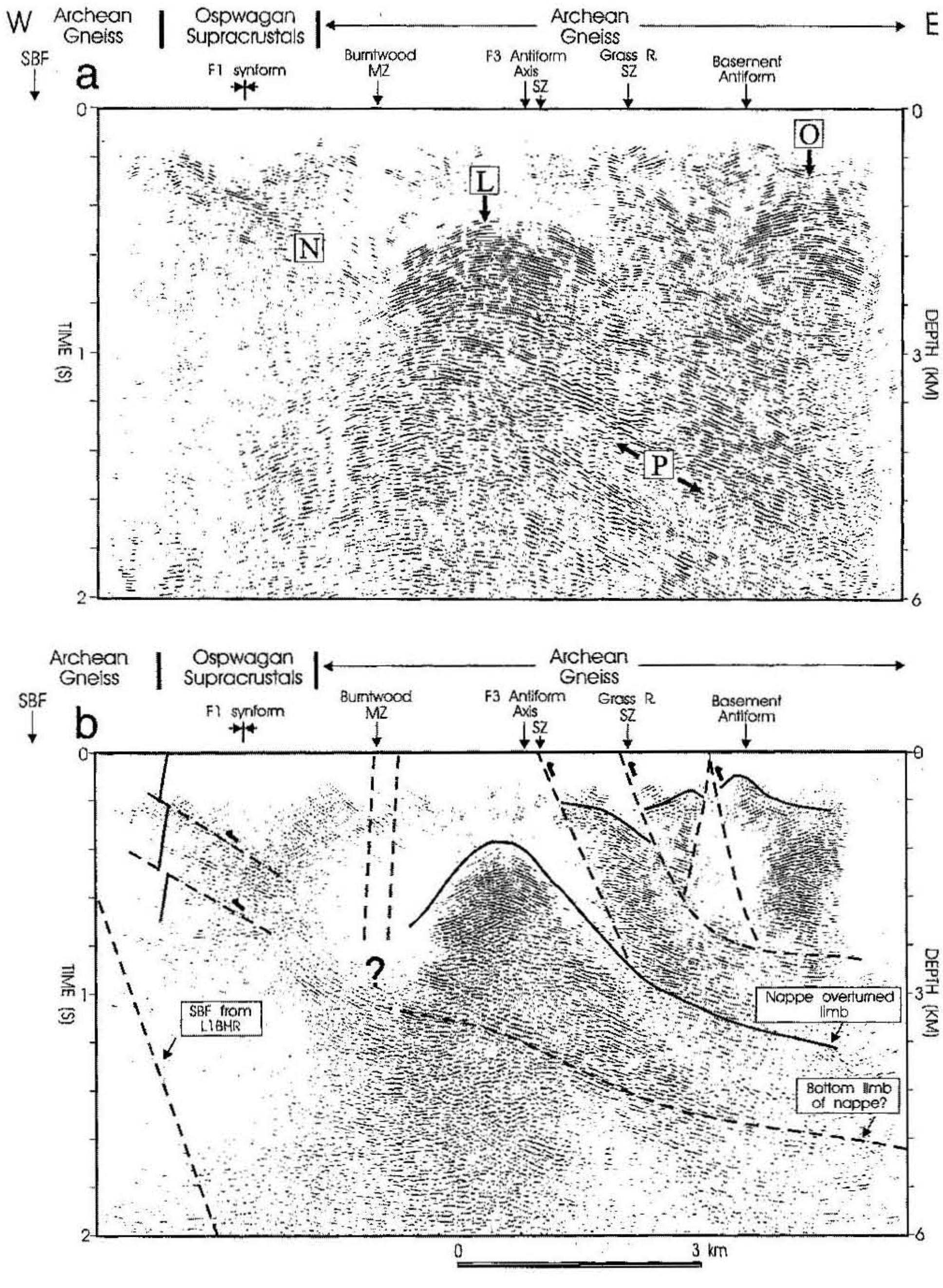


Figure 7. Interpreted 1AHR high-resolution seismic data: (a) unmigrated data and (b) migrated data. Labels are referred to in the text; SZ is shear zone and MZ is mylonite zone.

antiformal patterns separated by zones of reduced reflectivity (Figure 7b).

Line 1BHR (Figure 8), which straddles the SBF, is characterized by a vertical sequence of arcuate reflections (Q) within the SBZ that extend to at least 1.5 s. The reflections become diffractive in nature near the SBF. West of the SBF, a series of moderately east dipping reflections (R) are observed in an otherwise transparent upper crust. After migration, the antiformal reflections (Q) are clearly truncated beneath the SBF, and dipping reflectors (R) have apparent dips of ~ 45° to the east.

### 5.4. Interpretation

The high-resolution images provide structural details that are more readily comparable to the scale of geological structures that have been mapped at the surface in this region and generally corroborate *Bleeker's* [1990a, b] recent structural synthesis. As described below, much of the reflectivity observed in the high-resolution images is likely due to primary lithologic contacts and early (F1/F2) transpositional fabrics [cf. *White et al.*, 1997], but the gross geometry of the images reveals mainly F3/D3 and younger (postcollisional) structures.

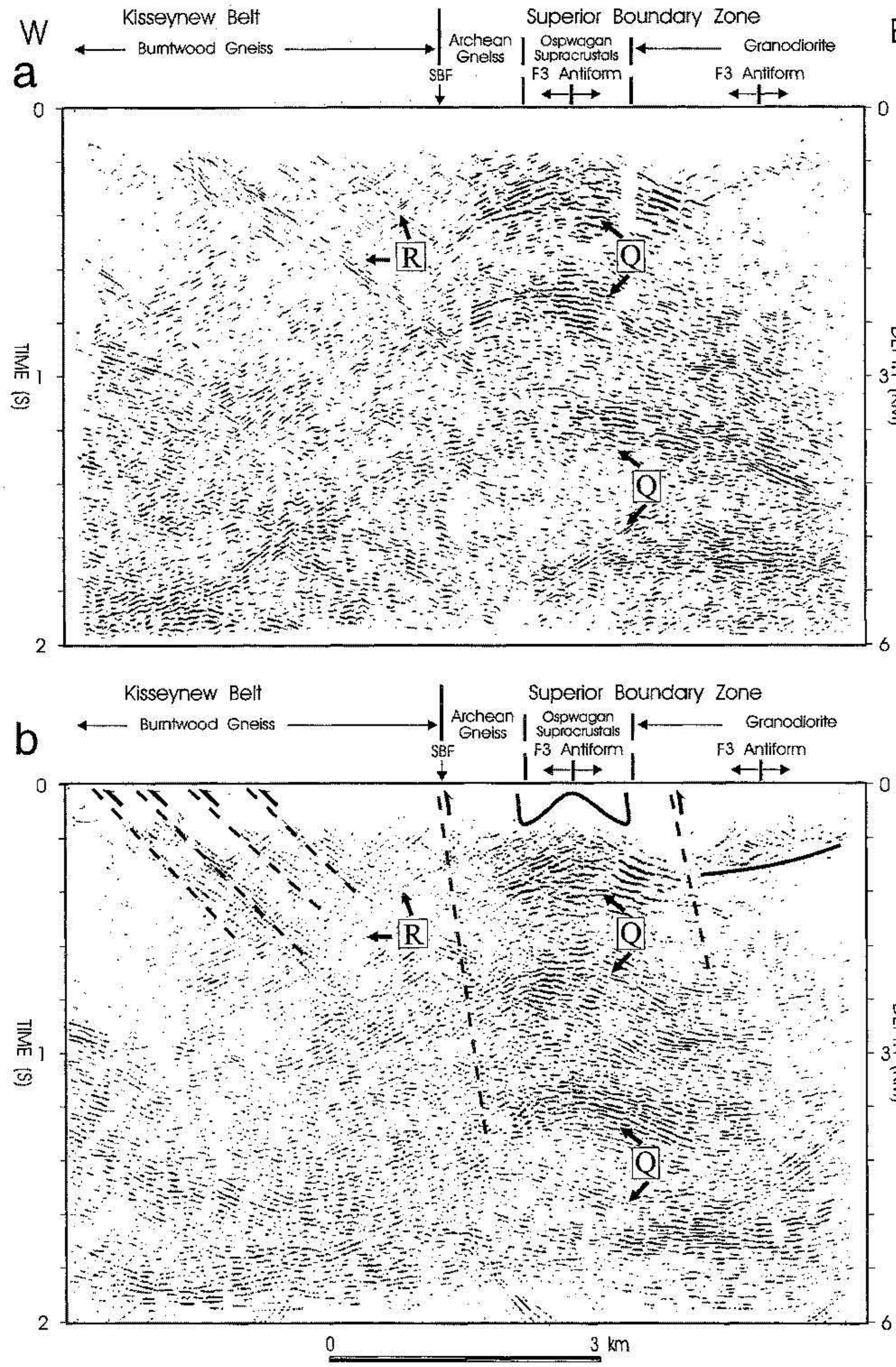


**Figure 8.** Interpreted Line 1BHR high-resolution seismic data: (a) unmigrated data and (b) migrated data. Labels are referred to in the text.

Figure 9 portrays a 3-D perspective of the surface geology along line 1A and a cartoon of the seismic interpretation. Zones of high reflectivity along line 1AHR are interpreted as being associated primarily with the Ospawagan group supracrustal rocks, based on the results of rock property measurements and borehole geologic logs [*White et al.*, 1997]. East of the Burntwood mylonite zone (Figures 7 and 9), the highly reflective regions outline an F1 Ospwagan group cored nappe, folded and faulted by upright F3 structures [cf. *Bleeker*, 1990a, b]. Farther east, the overturned basement-cover contact is imbricated along steeply SE-dipping reverse faults that postdate F3 folding and appear to be listric to the SE at 3-4 km depth. This interpretation suggests that these structures accommodate Superior-side-up displacement, similar to the bulk of the late, ductile reverse faults found

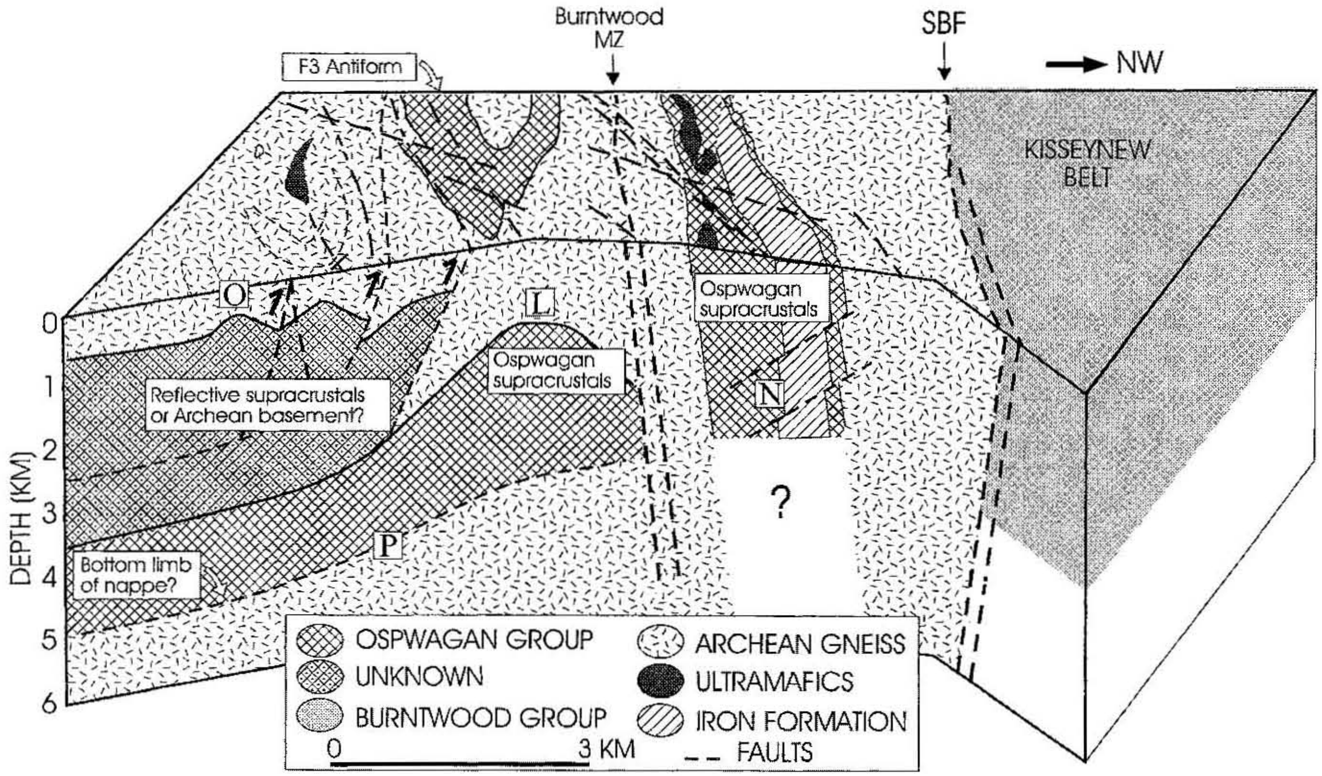


**Figure 9.** Interpretation cartoon for the line 1AHR high-resolution profile. The detailed subsurface information northwest of the Burntwood mylonite zone is from mine and drill hole information.

throughout the SBZ [cf. *Fueten and Robin*, 1989; *Bleeker*, 1990a, b].

The F3 antiform is truncated to the west by the Burntwood mylonite zone which is interpreted as a post-F3 subvertical fault consistent with subsurface drill hole information. The seismic signature changes dramatically across the Burntwood mylonite zone from reflective in the east to less reflective in the west, because of a westward change across the structure to known steeply dipping lithologies. Subsurface constraints from boreholes and underground mining [*White et al.*, 1997] indicate that the prominent moderately east dipping reflections along this part of the line are associated with faults rather than lithologic contacts which are steeply west dipping.

Superior Province basement forms essentially all of the SE half of line 1BHR. The reflection fabric (Q in Figure 8) within the SBZ may represent a gneissosity in the reworked Archean basement, probably Archean in age, reoriented during Hudsonian collisional/postcollisional deformation. An F3 antiform is well imaged at ~ 1 km depth on the seismic section and appears to be cored by Superior basement rocks, which are truncated by the apparently steeply SE dipping SBF. Surface geology indicates that Ospwagan group supracrustal rocks occur in the synformal keels both east and west of the F3 antiform, but their absence on the seismic section indicates that they do not project below ~ 1 km depth. The series of moderately SE dipping, strong, planar reflections that characterize the shallow (<1 km depth) seismic data are interpreted as syn- to post-F3 east-side-up discrete shear zones and/or faults, similar to the post-F3 structures mapped in the SBZ [*Fueten and Robin*, 1989; *Bleeker*, 1990a, b].

## 6. Discussion

### 6.1. A Model for the Tectonic History of the SBZ

In marked contrast to most of the tectonic models proposed for the SBZ [for example, *Green et al.*, 1985; *Bickford et al.*, 1990; *Bleeker*, 1990a, b], the reprocessed seismic images and the electrical resistivity models clearly demonstrate that rocks of the Reindeer Zone dip beneath the SBZ at middle to lower crustal depths (to at least 45 km) along the line 2/line 3 (southern) transect. In order to accommodate the apparently conflicting views of the SBZ provided by surface geology and geophysical data, we have developed a tectonic model that is consistent with both the geological observations and interpretations and the present crustal structure as imaged seismically and electromagnetically. This model also accounts for the observed along-strike variation in SBZ crustal structure as imaged along the northern profile, where there is no compelling evidence for significant underthrusting of Reindeer Zone rocks beneath the SBZ. The critical aspect of the model is that the nature of the SBZ changed dramatically over its ~200 m.y. convergent margin history, from a lower plate collisional thrust belt setting at >1.88-1.80 Ga to a steep transpressive plate boundary accommodating highly oblique convergence with the Reindeer Zone at 1.80-1.72 Ga [*Bleeker*, 1990a, b; *Machado*, 1990]. We draw analogy to the west central Alps, where seismic reflection profiling has imaged a wedge of colliding lithosphere (Adriatic subplate) that appears to have delaminated the (lower plate) European continental lithosphere at deep crustal levels [*Nicolas et al.*, 1990] in an overall tectonic regime of dextral transpression [*Schmid et al.*, 1989]. A tectonic

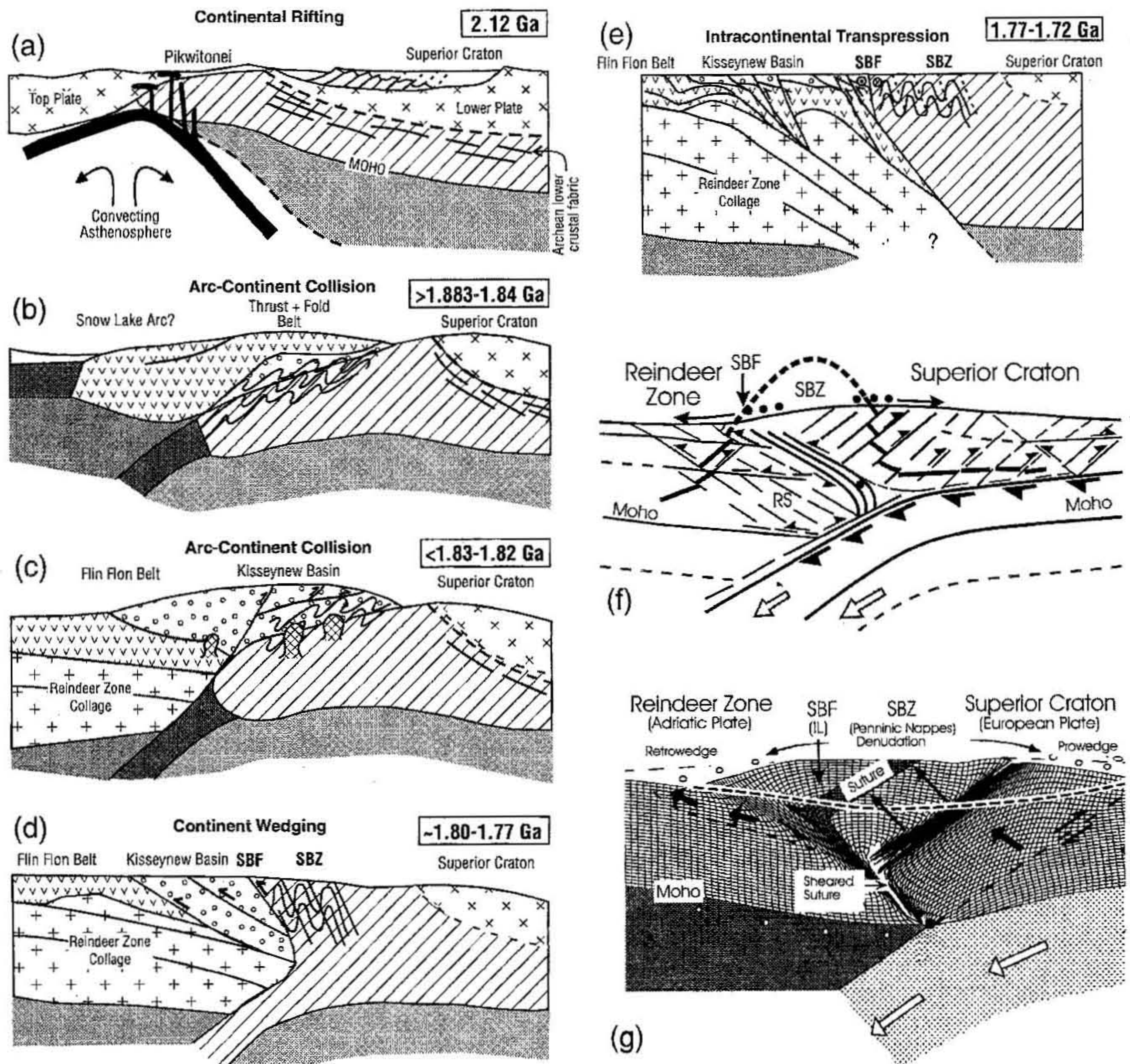


**Figure 10.** Cartoons illustrating the proposed tectonic stages for the SBZ margin. See text for details. (a) Lithospheric thinning and subsequent rifting of the Superior craton resulted in exhumation of the Pikwitonei granulites, upwarping of the Archean lower crustal fabric, and injection of the 2.17-2.12 Ga mafic dykes. (b) Arc-continent collision involving an unknown (Snow Lake arc?) outboard terrane resulted in an early east verging thrust and fold belt. (c) Continent-continent collision between the SBZ and the emerging continental lithosphere of the Flin Flon Belt results in renewed east vergent fold/thrusting in the SBZ, back thrusting in the adjacent Reindeer Zone accompanied by peak metamorphic conditions, and anatectic intrusion of granitoids. (d) Further convergence results in wedging of the Flin Flon crustal block into the edge of the Superior craton, with associated east-side up folding and faulting on both sides of the SBF. (e) Intracontinental transpression continued until 1.72 Ga resulting in the crustal geometry that is preserved today. (f) Cartoon modified from *Beaumont and Quinlan* [1994] showing expected general deformation patterns controlled by asymmetric mantle subduction for a small compressional orogen. RS is retroshear zone; dashed line indicates expected structural relief across the prowedge. (g) Predicted model evolution (modified from *Beaumont et al.* [1996]) for the advanced collisional stage of an Alpine-type compressional orogen as indicated by the deformed model grid. Arrows indicate direction of mass transport, and dashed line represents denuded surface. Labels refer to the SBZ and Alpine analogues, respectively. IL is Insubric Line.

model for the Paleoproterozoic history of the SBZ, based on the geophysical and geological data, is presented in Figure 10.

### 6.2. Precollision history

The last pervasive episode of regional metamorphism within the western Superior Province (represented by granulite-grade metamorphism dated at 2.637-2.695 Ga in the Pikwitonei Belt) was followed by uplift of the Superior crust by ~ 2.17- 2.18 Ga. The seismic interpretation is consistent with the magnitude of crustal uplift, although it does not constrain the timing or cause of uplift. Crustal exhumation is the result of some combination of (1) late Archean deformation, (2) extensional thinning associated with continental rifting as has been previously suggested [e.g., *Green et al.*, 1985; *Weber and Mezger*, 1990], and (3) Hudsonian orogeny as is suggested by the broad zone of granulite exposure within the promontory of the Superior margin north of line 1 (see Figure 1). Exhumation of the Pikwitonei granulites is depicted in Figure 10a as the result of simple shear extension [cf. *Wernicke*, 1985] prior to rifting of the Superior craton with attendant reorientation of the inferred subhorizontal Archean seismic fabric. The onset of rifting of the Superior craton is dated at 2.17-2.12 Ga [*Zhai and Halls*, 1994; *Heaman et al.*, 1986] and was followed by development of the Manikewan ocean [*Stauffer*, 1984] outboard of the Superior margin.

### 6.3. Early Collision History: 1.88-1.84 Ga

Early collision (Figure 10b) between an inferred outboard terrane and the Superior continental margin was underway by 1.883 Ga, as Molson dykes crosscut early (F1) SBZ structures and resulted in the development of an east verging thrust and fold belt [*Bleeker*, 1990a, b] prior to 1.820 Ga, the inferred age of high-grade metamorphism in the SBZ [*Machado*, 1990]. This part of the history of the SBZ is not well constrained geochronologically or by the seismic data presented here, although much of the reflectivity observed on the high-resolution lines is thought to be associated with F1-F2 deformed continental margin and Superior basement lithologies that were subsequently refolded during later (F3) deformation. *Lucas et al.* [1996b] have recently reported seismic reflection evidence for an east verging SBZ thrust belt (probably F2 age) preserved below the Phanerozoic cover south of line 2.

The F1 (~1.88 Ga) collisional history of the SBZ margin is problematic, as the colliding terrane has not been identified. The Kisseynew Belt rocks currently juxtaposed along the SBF are 1.85-1.84 Ga old and do not contain a significant Superior Province detrital zircon signature. Furthermore, paleomagnetic data indicate that the Superior craton was likely separated from the Flin Flon Belt at ~1.90-1.84 Ga [*Symons et al.*, 1995]. However, the Snow Lake arc assemblage in the eastern Flin Flon Belt may be a suitable candidate for the outboard terrane, as it contains some geochronological and isotopic evidence of interaction with Superior age crust at ~1.89 Ga [*David et al.*, 1996]. In contrast, the older crust involved at the time of early magmatism and deformation in the western portion (Amisk collage) of the Flin Flon Belt has a unique 2.5 Ga signature [e.g., *David and Syme*, 1995] not found in the Snow Lake arc assemblage. This hypothesis could be tested with paleomagnetic pole determinations for the Snow Lake assemblage and Amisk collage at 1.88 Ga, and comparison with determinations from Molson dykes of the SBZ [cf. *Zhai and Halls*, 1994]. Evidence for 1.864 Ga mafic to ultramafic magmatism along the buried portion of the SBZ [cf. *Lucas et al.*, 1996b] suggests that the colliding terrane (Snow Lake assemblage?) could have rifted away from the Superior margin prior to terminal collision at ~ 1.82 Ga [cf. *Weber*, 1990].

### 6.4. Collisional History: 1.84-1.81 Ga

Strong geologic evidence for the terminal collision of the Superior craton with the Reindeer Zone first appears at ~1.82-1.80 Ga. Closure of the 700 km of E-W separation between the Flin Flon Belt and Superior craton between 1.85 Ga and 1.82-1.80 Ga would require a convergence rate of 1.5-2.0 cm/yr, which is certainly reasonable compared to modern day tectonic processes. A component of the widespread magmatism recorded in the Flin Flon and Kisseynew belts up until ~1.83 Ga may be associated with the westward subduction of oceanic crust accommodating closure between the Superior craton and the Reindeer Zone accretionary collage to the west [cf. *Ansdell et al.*, 1995].

The age of F2 deformation within the SBZ prior to peak metamorphism at ~ 1.82 Ga is unknown. However, a diorite (1.83 Ga [*Bleeker et al.*, 1995]), anatectic granites (1.82-1.80 Ga [*Machado*, 1990]), and peak metamorphism (1.82-1.805 Ga [*Machado*, 1990]) in the SBZ are considered to be associated with late F2 deformation resulting in the development of an east verging fold and thrust belt [*Bleeker*, 1990a, b]. Within the Reindeer Zone, structures that developed during 1.84-1.80 Ga, record predominantly S-SW overthrusting of juvenile Reindeer Zone units on the Archean Sask craton [*Lewry*, 1990; *Ansdell et al.*, 1995]. Terminal collision (Figure 10c) of Superior craton with this N-NE dipping crustal scale thrust stack, which apparently developed largely prior to collisional interaction with Superior craton, resulted in the observed eastward tilting and further crustal shortening within the Reindeer Zone.

### 6.5. Late collisional to Postcollisional History: 1.81-1.72 Ga

It is during the period between 1.82 and 1.80 Ga that the fundamental change in the geometry and kinematics of deformation along the SBZ occurred [*Bleeker*, 1990a, b]. The geophysical interpretations indicate a steep attitude to the SBF in the present day upper and middle crust at both locations where it is crossed, generally consistent with *Bleeker's* [1990a, b] model for collisional/postcollisional sinistral transpression. However, the persistence of east dipping reflections and conductive rocks to ~ 45 km beneath the SBZ (southern corridor) that are contiguous with Reindeer reflections suggests that this SBZ deformation may have occurred above a wedge of Reindeer Zone juvenile crust that indented and delaminated the Superior craton (Figure 10d). This model is directly analogous to that described above for the western Alps, with Superior craton being equivalent to the European (lower) plate and the Reindeer Zone being equivalent to the Adriatic indentor [cf. *Schmid et al.*, 1989; *Beaumont et al.*, 1996].

The moderately to steeply SE dipping reflections observed within the SBZ are probably associated with the F3 folding of earlier (S1-S2) foliations and compositional layering and the syn- to post-F3 ductile shear zones (Superior-side-up back thrusts) dated at ~1.80-1.77 Ga [*Bleeker*, 1990a, b; *Machado*, 1990]. Postcollisional, intracontinental deformation in the vicinity of the SBZ (Figure 10e) appears to have occurred until ~1.72 Ga [*Bleeker*, 1990a, b; *Machado*, 1990], manifest by ductile to brittle strike-slip faulting along the SBF and steep reverse faulting (Superior side up) within the SBZ and the adjacent Kisseynew Belt. The SBF was clearly active during this period as demonstrated by the truncation of F3 collisional fabrics by the interpreted subsurface expression of the SBF (along line 1A/1B). Intracontinental (sinistral) transpression was also accommodated by strike-slip motion along major orogen-parallel faults within the Reindeer Zone [*Bleeker*, 1990a, b; *Hajnal et al.*, 1995], similar to the faults of the Periadriatic system [*Schmid et al.*, 1989].

In contrast to the line 2/3 (southern) profile where the SBF is subvertical to 15 km depth and then dips eastward beneath the SBZ to > 40 km, the SBF along line 1A/1B (northern profile) is interpreted as subvertical to 30 km depth where it is underlain by lower crustal material of unknown affinity. This suggests a somewhat different mode of deformation than that interpreted for the southern profile. If the deep crust of unknown affinity represents Superior crust, then this geometry may be the result of crustal delamination as observed to the south except that delamination occurred at a higher level within the crust and the SBF has been steepened over its entire depth extent. Alternatively, the apparent difference in crustal structure between the two profiles may be associated with along-strike variations in the physical properties of the colliding lithosphere or the 3-D collisional geometry. The possibility that physical properties may have played a role is suggested by temperatures at ~1.81 Ga within the Kisseynew Belt which were 200° higher than in the adjacent Flin Flon collage [*Menard and Gordon*, 1996]. Thus, during collision with the Superior craton, the Kisseynew Belt was likely warmer and perhaps more amenable to homogeneous crustal thickening.

### 6.6. Relation to Plastic Deformation Models

Although we have invoked a rigid indentor model to explain the observed crustal geometry, "indentation" can be achieved by asymmetric detachment and underthrusting of mantle lithosphere in a lithosphere with plastic-viscous rheology [cf. *Beaumont and Quinlan*, 1994]. The general pattern of crustal vergence observed near the SBZ along the southern profile (line 2/3) can be accommodated by a model in which the Superior plate delaminated near the base of the crust, with the Superior mantle lithosphere being subducted westward (Figure 10f). Comparison of a generic form of the mantle subduction model [from *Beaumont and Quinlan*, 1994] with the cartoon model from our study (Figure 10e) allows the following observations. The prominent east dipping reflectivity characterizing the SBZ corresponds to the retroshear zone, which is subject to the most intense deformation within the model and thus should correspond to the predominant attitude of reflectivity (i.e., east dipping). The model also predicts an increase in uplift eastward across the retroshear zone, perhaps manifest in the SBZ by the absence of Paleoproterozoic supracrustal rocks away from its western margin and the exposure of granulite facies rocks (Paint Lake Domain [*Fueten and Robin*, 1989]). A well-coupled lower crust is indicated by the narrow width of the SBZ, as the width of the deformation zone is inversely proportional to the coupling strength at the detachment level in models driven by basal detachment [*Ellis*, 1996]. An apparent discrepancy between the models is the absence of west dipping shear zones (pro-shears) of Hudsonian age in the SBZ or Superior crust to the east, although this may be the result of erosion accompanying uplift within the SBZ. However, a seismic line acquired south of the Shield Margin in a region where higher structural levels are preserved shows west dipping reflections associated with an east verging thrust belt [*Lucas et al.*, 1996b].

Pursuing the analogy of the SBZ and the midwestern Alps, *Beaumont et al.* [1995] have suggested comparison of the SBZ seismic model with a plastic deformation model developed specifically for an "Alpine-type" compressional orogen [*Beaumont et al.*, 1996]. In this geodynamical model, the Alpine orogen is characterized by three distinct convergent phases which closely resemble the stages depicted in Figures 10a-10c for the SBZ: subduction, a transition from subduction to collision, and continental collision with accompanying orogenic vergence reversal. The transition from subduction to collision and subsequent vergence reversal in the model results from increased buoyancy of the subducting lithosphere which in the case of the SBZ would correspond to attempted subduction of the buoyant Superior lithosphere. Figure 10g depicts this final collisional stage of orogeny. At this advanced stage of the collision, the remnants of the suture zone between the colliding plates have been realigned (see Figure 10d) with the retroshear zone (i.e., the SBF), and only the roots of the early provergent fold and thrust belt (i.e., the SBZ) remain along lines 1 and 2/3. A complicating factor in the case of the SBZ is that the collisional/postcollisional history is three-dimensional and largely associated with sinistral transpression rather than pure compression relative to the Reindeer Zone [cf. *Bleeker*, 1990a, b; *Ansdell et al.*, 1995]. Sinistral strike-slip motion along the SBF may represent the transcurrent component of oblique slip occurring along low-angle shear zones (i.e., the retroshear) in a predomimantly compressive regime, or it may be the result of strain partitioning in a largely transcurrent regime where strike-slip motion occurs along distinct steeply dipping shear zones [cf. *Braun and Beaumont*, 1995]. The latter scenario is likely appropriate for the postcollisional stage (<1.80 Ga) of deformation along the SBF as indicated by its subvertical attitude.

## 7. Conclusions

The present-day crustal structure in the vicinity of the SBZ, as interpreted from seismic and electromagnetic images, comprises the following: (1) The Flin Flon accretionary collage and Kisseynew Belt form an east dipping, eastward steepening, crustal-scale tectonic stack of moderately conductive rocks near the SBZ. (2) The SBZ is characterized at shallow depths (< 5-6 km) by steep to moderate east dipping reflectivity that is associated with late F3 and D3 fold limbs and east-side-up shear zones. At greater depth (5-10 km), the SBZ crust is highly resistive and is contiguous to the east with resistive crust beneath the Pikwitonei granulite belt and Superior craton proper. Westward shallowing reflectivity within the Pikwitonei granulite belt is consistent with 12 km of structural relief. (3) The SBF extends subvertically to at least 15 km depth (southern profile) and to as much as 30 km depth (northern profile). Geophysically, it is recognized in the subsurface as a truncation of interpreted ~ 1.8 Ga collisional seismic fabrics beneath the western SBZ (northern profile) or as an abrupt resistivity contrast between the Reindeer Zone and SBZ (southern profile). (4) Along-strike variations in the lower crustal structure near the SBZ are observed between the southern and northern profiles. Moderately conductive rocks of the Reindeer Zone apparently extend eastward beneath the SBZ at lower crustal depths along the southern profile; whereas along the northern profile, the seismic reflection data provide no compelling evidence for a similar eastward continuation of Reindeer Zone lower crust beneath the SBZ.

We propose a tectonic model that attempts to reconcile both geological observations and interpretations and the present-day crustal structure from the geophysical interpretations. The key aspect of this model is that the nature of the SBZ changed dramatically over its ~200 m.y. convergent margin history. Specifically, this model draws analogy to the west central Alps to suggest that the Reindeer Zone lithosphere collided with and delaminated the Superior craton lithosphere near the base of the crust. The essence of the model is as follows: (1) Collision of Superior craton prior to 1.883 Ga with an unknown terrane (perhaps Snow Lake arc terrane) resulted in an east verging fold (F1) and thrust belt within the SBZ. (2) Terminal collision of Superior craton with the Flin Flon accretionary collage commenced at ~1.83 Ga with Superior craton forming the lower plate, resulting in renewed deformation (F2) and synchronous peak metamorphism in the eastern Flin Flon Belt and SBZ by ~1.805 Ga. (3) A fundamental change in the geometry and kinematics of deformation along the SBZ occurred between 1.82 and 1.80 Ga, with juvenile crust of the Reindeer Zone indenting and delaminating the Superior craton. Sinistral transpression resulted in strike-slip faulting along the SBF with upright folding (F3) and steep reverse faulting (east side up) within the SBZ between 1.80 and 1.77 Ga and eastward tilting and further crustal shortening within the Reindeer Zone. (4) Intracontinental transpressional deformation continued until at least 1.72 Ga with sinistral strike-slip motion accommodated along the near-vertical SBF. The along-strike variation in crustal structure observed in two crossings of the SBZ may indicate different levels of crustal delamination within the Superior craton or collision with warmer Reindeer Zone lithosphere to the north.

**Acknowledgments.** The authors thank W. Bleeker, J. Hall, and an anonymous reviewer for reviewing the original manuscript. Phoenix Geophysics Ltd. is recognized for their high-quality data acquisition. This is Geological Survey of Canada Contribution 1998193 and Lithoprobe Publication 974.

## References

Ansdell, K., U-Pb constraints on the timing and provenance of fluvial sedimentary rocks in the Flin Flon and Athapapuskow Basins, Flin Flon Domain, Trans-Hudson Orogen, in *Radiogenic Age and Isotopic Studies: Report 7, GSC Pap. 93-2*, pp. 49-57, Geol. Surv. Of Canada, Ottawa, 1993.

Ansdell, K., and A. R. Norman, U-Pb geochronology and tectonic development of the southern flank of the Kisseynew Domain, Trans-Hudson Orogen, Canada, *Precambrian Res.*, *72*, 147-167, 1995.

Ansdell, K., S. B. Lucas, K. Connors, and R. Stern, Kisseynew metasedimentary gneiss belt, Trans-Hudson orogen (Canada): Back-arc origin and collisional inversion, *Geology*, *23*, 1039-1043, 1995.

Beaumont, C., and G. Quinlan, A geodynamic framework for interpreting crustal-scale seismic reflectivity patterns in compressional orogens, *Geophys. J. Int.*, *116*, 754-783, 1994.

Beaumont, C., S. Ellis, J. Hamilton, and P. Fullsack, A geodynamical model of the transition from subduction to collision tectonics and its application to the eastern Trans-Hudson orogen, in *Trans-Hudson Orogen Transect, Lithoprobe Rep.*, *48*, 231-243, 1995.

Beaumont, C., S. Ellis, J. Hamilton, and P. Fullsack, Mechanical model for subduction-collision tectonics of Alpine-type compressional orogens, *Geology*, *24*, 675-678, 1996.

Bickford, M. E., K. D. Collerson, J. F. Lewry, W. R. Van Schmus, and J. R. Chiarenzelli, Proterozoic collisional tectonism in the Trans-Hudson orogen, Saskatchewan, *Geology*, *18*, 14-18, 1990.

Bleeker, W., New structural-metamorphic constraints on Early Proterozoic oblique collision along the Thompson Nickel Belt, Manitoba, Canada. in *The Early Proterozoic Trans-Hudson Orogen of North America*, edited by J. F. Lewry and M. R. Stauffer, *Geol. Assoc. Can. Spec. Pap.*, *37*, 57-73, 1990a.

Bleeker, W., Evolution of the Thompson Nickel Belt and its nickel deposits, Manitoba, Canada, Ph.D. thesis, 400 pp. and maps, University of New Brunswick, Fredericton, N.B., Canada, 1990b.

Bleeker, W., P. Nägerl, and N. Machado, The Thompson Nickel Belt, Manitoba: Some new U-Pb ages, paper presented at 1995 Geological Association of Canada-Mineralogical Association of Canada Annual Meeting, Victoria, British Columbia, 1995.

Braun, J., and C. Beaumont, Three-dimensional numerical experiments of strain partitioning at oblique plate boundaries: Implications for contrasting tectonic styles in the southern Coast Ranges, California, and central South Island, New Zealand, *J. Geophys. Res.*, *100*, 18,059-18,074, 1995.

Connors, K. A., Unraveling the boundary between turbidites of the Kisseynew belt and volcano-plutonic rocks of the Flin Flon belt, Trans-Hudson Orogen, Canada, *Can. J. Earth Sci.*, *33*, 811-829, 1996.

David, J., and E. C. Syme, U-Pb geochronology of late Neoarchean tonalites in the Flin Flon Belt, Trans-Hudson Orogen: Surprise at the surface, *Can. J. Earth Sci.*, *31*, 1785-1790, 1995.

David, J., A. H. Bailes, and N. Machado, Evolution of the Snow Lake portion of the Paleoproterozoic Flin Flon and Kisseynew belts, Trans-Hudson Orogen, Manitoba, Canada, *Precambrian Res.*, *80*, 107-124, 1996.

Ellis, S., Forces driving continental collision: Reconciling indentation and mantle subduction tectonics, *Geology*, *24*, 699-701, 1996.

Fountain, D. M., and M. H. Salisbury, Exposed cross-sections through the continental crust: implications for crustal structure, petrology and evolution, *Earth Planet. Sci. Lett.*, *56*, 263-277, 1981.

Fueten, F., and P.-Y.F. Robin, Structural petrology along a transect across the Thompson Belt, Manitoba: Dip slip at the western Churchill-Superior boundary, *Can. J. Earth Sci. 26*, 1976-1989, 1989.

Gibb, R. A., A model for suturing of Superior and Churchill plates: An example of double indentation tectonics, *Geology*, *11*, 413-417, 1983.

Gordon, T. M., P. A., Hunt, A. H. Bailes, and E. C. Syme, U-Pb ages from the Flin Flon and Kisseynew Belts, Manitoba: Chronology of crust formation at an Early Proterozoic accretionary margin, in *The Early Proterozoic Trans-Hudson Orogen of North America*, edited by J. F. Lewry and M. R. Stauffer, *Geol. Assoc. Can. Spec. Pap.*, *37*, 177-199, 1990.

Green, A. G., Z. Hajnal, and W. Weber, An evolutionary model of the western Churchill Province and western margin of the Superior Province in Canada and the north-central United States, *Tectonophysics*, *116*, 281-322, 1985.

Hajnal, Z., S. B. Lucas, D. J. White, J. Lewry, S. Bezdan, M.R. Stauffer, and M.D. Thomas, Seismic reflection images of strike-slip faults and linked detachments in the Trans-Hudson Orogen, *Tectonophysics*, *15*, 427-439, 1995.

Heaman, L., N. Machado, T. E. Krogh, and W. Weber, Precise U-Pb zircon ages for the Molson dyke swarm and the Fox River sill: constraints for Early Proterozoic crustal evolution in northeastern Manitoba, Canada, *Contrib. Mineral. Petrol.*, *94*, 82-89, 1986.

Hoffman, P. F., Early Proterozoic foredeeps, foredeep magmatism, and Superior-type iron-formations for the Canadian Shield, in *Proterozoic Lithospheric Evolution, Geodyn. Ser.*, vol. 17, edited by A. Kröner, pp. 85-98, AGU, Washington, D.C., 1987.

Hoffman, P. F., Precambrian geology and tectonic history of North America, in *The Geology of North America - An Overview*, edited by A. W. Bally and A. R. Palmer, pp. 447-512, Geol. Soc. of Am., Boulder, Colo., 1989.

Jones, A. G., and H. Jödicke, Magnetotelluric transfer function estimation improvement by a coherence-based rejection technique, 54th Annual Meeting, Soc. of Explor. Geophys., Atlanta, Ga., December 2-6, 1984.

Jones, A. G., J. A. Craven, G. W. McNeice, I. J. Ferguson, T. T. Boyce, C. Farqharson, and R.

G. Ellis, North American Central Plains conductivity anomaly within the Trans-Hudson orogen in northern Saskatchewan, Canada, *Geology, 21*, 1027-1030, 1993.

Leclair, A. D., S. B. Lucas, H. J. Broome, D. W. Viljoen, and W. Weber, Regional mapping of Precambrian basement beneath Phanerozoic cover in southeastern Trans-Hudson Orogen, Manitoba and Saskatchewan, *Can. J. Earth Sci., 34*, 618-634, 1997.

Lewry, J. F., Lower Proterozoic arc-microcontinent collisional tectonics in the western Churchill Province, *Nature, 294*, 69-72, 1981.

Lewry, J. F., and K. D. Collerson, The Trans-Hudson Orogen: extent, subdivisions and problems, in *The Early Proterozoic Trans-Hudson Orogen of North America*, edited by J. F. Lewry and M. R. Stauffer, *Geol. Assoc. Can. Spec. Pap., 37*, 1-14, 1990.

Lewry, J. F., D. J. Thomas, R. Macdonald, and J. Chiarenzelli, Structural relations in accreted terranes of the Trans-Hudson Orogen, Saskatchewan: Telescoping in a collisional regime?, in *The Early Proterozoic Trans-Hudson Orogen of North America,* edited by J.F. Lewry and M.R.Stauffer, *Geol. Assoc. Can. Spec. Pap., 37*, 75-94, 1990.

Lewry, J. F., Z. Hajnal, A. Green, S. B. Lucas, D. White, M. R. Stauffer, K. E. Ashton, W. Weber, and R. Clowes, Structure of a Paleoproterozoic continent-continent collision zone: A Lithoprobe seismic reflection profile across the Trans-Hudson Orogen, Canada, *Tectonophysics, 232*, 143-160, 1994.

Lucas, S. B., A. Green, Z. Hajnal, D. White, J. Lewry, K. Ashton, W. Weber, and R. Clowes, Deep seismic profile across a Proterozoic collision zone: Surprises at depth, *Nature, 363*, 339-342, 1993.

Lucas, S. et al., Three-dimensional collisional structure of the Trans-Hudson Orogen, Canada, in *Proceedings of the 5th International Conference on Seismic Reflection Probing of Continents and Their Margins*, edited by R. M. Clowes and A. G. Green, *Tectonophysics, 232*, 161-178, 1994.

Lucas, S. B., R. A. Stern, E. C. Syme, B. A. Reilly, and D.J. Thomas, Intraoceanic tectonics and the development of continental crust: 1.92-1.84 Ga evolution of the Flin Flon Belt (Canada), *Geol. Soc. Am. Bull., 108*, 602-629, 1996a.

Lucas, S. B., D. White, W. Bleeker, Z. Hajnal, J. Lewry, and W. Weber, Crustal structure of the Superior Boundary Zone (Thompson Belt) from new Lithoprobe seismic reflection data, in *Trans-Hudson Orogen Transect, Lithoprobe Rep., 55*, 82-94, 1996b.

Machado, N., Timing of collisional events in the Trans-Hudson Orogen: Evidence from U-Pb geochronology for the New Quebece Orogen, the Thompson Belt, and the Reindeer Zone (Manitoba and Saskatchewan), in *The Early Proterozoic Trans-Hudson Orogen of North America,* edited by J. F. Lewry and M. R. Stauffer, *Geol. Assoc. Can. Spec. Pap., 37*, 433-441, 1990.

Machado, N., T. Krogh, and W. Weber, U-Pb geochronology of basement gneisses in the Thompson Belt (Manitoba): Evidence for pre-Kenoran and Pikwitonei-type crust and Early Proterozoic basement reactivation in the western margin of the Archean Superior Province, *Can. J. Earth Sci., 27,* 794-802, 1990.

Manitoba Energy and Mines, Bedrock geology compilation map series, prelim. ed., *NTS-63J*, 1:250,000 scale, Manitoba Energy and Mines, Winnipeg, Manitoba, Canada, 1993.

McBride, J. H., and L. D. Brown, Reanalysis of the COCORP deep seismic reflection profile across the San Andreas Fault, Parkfield, California, *Bull. Seismol. Soc. Am., 76*, 1668-1686, 1986.

McGeary, S., Reflection seismic evidence for a Moho offset beneath the Walls Boundary strike-slip fault, *J. Geol. Soc. London, 146*, 261-269, 1989.

McNeice, G., and A. G. Jones, Multisite, multifrequency tensor decomposition of magnetotelluric data, 66th Annual Meeting, Soc. of Explor. Geophys., Denver, Colo., November 10-15, 1996.

Menard, T., and T. M. Gordon, Heat and fluid flow during deformation of VMS deposits Snow Lake, Manitoba, *Lithoprobe Rep., 55*, 106-112, 1996.

Nicolas, A., A. Hirn, R. Nicolich, R. Polino, and ECORS-CORP Working Group, Lithospheric wedging in the western Alps inferred from the ECORS-CROP traverse, *Geology, 18*, 587-590, 1990.

Schmid, S. M., H. R. Aebli, F. Heller, and A. Zingg, The role of the Periadriatic Line in the tectonic evolution of the Alps, in *Alpine Tectonics*, edited by M.P. Coward, D. Dietrich, and R.G. Park, *Geol. Soc. Spec. Publ., 45*, 153-171, 1989.

Smith, J. T., and J. R. Booker, Rapid inversion of two- and three-dimensional magnetotelluric data, *J. Geophys. Res., 96*, 3905-3922, 1991.

Stauffer, M. R., Manikewan: An Early Proterozoic ocean in central Canada, its igneous history and orogenic closure, *Precambrian Res., 25*, 257-281, 1984.

St. Onge, M. R., and S. B. Lucas, Evolution of the Cape Smith Belt: Early Proterozoic continental underthrusting, ophiolite obduction and thick-skinned folding, in *The Early Proterozoic Trans-Hudson Orogen of North America*, edited by J. F. Lewry and M. R. Stauffer, *Geol. Assoc. Can. Spec. Pap., 37*, 313-351, 1990.

Symons, D. T. A., M. Gala, and H. C. Palmer, Fitting Paleomagnetic data to a plate tectonic model for the Trans-Hudson Orogen with a focus on the Hanson Lake block, in *Trans-Hudson Orogen Transect, Lithoprobe Rep.*, 48, 66-77, 1995.

Weber, W., The Churchill-Superior Boundary Zone, southeast margin of the Trans-Hudson Orogen: A review, in *The Early Proterozoic Trans-Hudson Orogen of North America,* edited by J. F. Lewry and M. R. Stauffer, *Geol. Assoc. Can. Spec. Pap., 37*, 42-55, 1990.

Weber, W., and K. Mezger, An oblique cross section of Archean continental crust at the northwestern margin of the Superior Province, Manitoba, Canada, in *Exposed Cross-Sections of the Continental Crust*, edited by M. H. Salisbury and D. M. Fountain, Kluwer Acad., Norwell, Mass., 1990.

Wernicke, B. P., Uniform-sense normal simple shear of the continental lithosphere, *Can. J. Earth Sci., 22*, 108-125, 1985.

White, D. J., S. B. Lucas, Z. Hajnal, A. G. Green, J. F. Lewry, W. Weber, A. H. Bailes, E. C. Syme, and K. E. Ashton, Early Proterozoic thick-skinned tectonics: Lithoprobe seismic reflection results from the eastern Trans-Hudson Orogen, *Can. J. Earth Sci., 31*, 458-469, 1994.

White, D. J., D. E. Boerner, J. Wu., S. B. Lucas, E. Berrer, J. Hannnila, and R. Somerville, High-resolution seismic and controlled-source EM studies near Thompson, Manitoba, in *Proceedings of Exploration 97: Fourth Decennial International Conference on Mineral Exploration*, edited by A. G. Gubins, pp. 685-694, Prospectors and Developers Association of Canada, Toronto, Ontario, 1997.

Zhai, Y., and H. C. Halls, Multiple episodes of dike emplacement along the northwestern margin of the Superior Province, Manitoba, *J. Geophys. Res., 99*, 21,717-21,732, 1994.

Zwanzig, H. V., Kisseynew Gneiss Belt in Manitoba: Stratigraphy, structure, and tectonic evolution, in *The Early Proterozoic Trans-Hudson Orogen of North America*, edited by J. F. Lewry and M. R. Stauffer, *Geol. Assoc. Can. Spec. Pap., 37*, 95-120, 1990.

Zwanzig, H. V., Alternative structural restorations of the Flin Flon belt - Kisseynew belt boundary zone, in *Trans-Hudson Orogen Transect, Lithoprobe Rep., 48*, 135-142, 1995.

Z. Hajnal, Department of Geological Sciences, University of Saskatchewan, Saskatoon, Saskatchewan, Canada S7N 0W0.

A. G. Jones, S. B. Lucas, and D. J. White, Geological Survey of Canada, 615 Booth Street, Ottawa, Ontario, Canada K1A 0E9. (jones@cg.nrcan.gc.ca; slucas@gsc.nrcan.gc.ca; white@cg.nrcan.gc.ca)

(Received June 4, 1998;
revised December 8, 1998;
accepted January 19, 1999.)