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Wide-angle seismic imaging of a Mesoproterozoic anorthosite complex: The Nain Plutonic Suite in Labrador, Canada

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Abstract. The Mesoproterozoic Nain Plutonic Suite (NPS) of Labrador (Canada), one of the largest anorogenic plutonic terranes, was studied by a refraction/wide-angle seismic experiment. Four ocean bottom seismometers and 18 land stations were deployed along a 330-km profile and recorded air gun shots from the easternmost 160 km with the NPS located in the center of the line at the suture of the Nain and Churchill Provinces. *P* and *S* wave velocity models were developed by forward modeling of travel times and amplitudes. Upper and middle crustal *P* wave velocities outside and beneath the NPS range from 5.9 to 6.5 km/s, lower crustal *P* wave velocities range from 6.55 to 7.0 km/s. Within the anorthositic rocks, velocities are as high as 6.8 km/s, and reflections define the base of the NPS to be 8 km deep in the SE Churchill Province and 11 km in the Nain Province, a variation that may be the result of lateral density changes within the country rocks or the anorthosites. The total crustal thickness is 39 km west of the NPS but is only 32-34 km beneath the NPS, some 5 km less than Nain Province crust distal from the NPS. The inferred crustal thinning is possibly related to anatexis of the lowermost crust by a thermal plume that generated the plutonism. The Poisson's ratios are 0.275 within the anorthosite plutons, 0.27 in the upper and middle crust, and 0.285 in the lower crust. These values are some 0.03 higher than in the Archean Nain crust distal to the NPS, indicating a higher plagioclase content at all crustal levels as result of the plutonism. We postulate that a crustal root, similar to the root observed farther north in the Torngat Orogen, was completely removed by anatexis and the silicic and basic magmas probably ascended to midcrustal levels along preexisting zones of weakness at the Nain-Churchill boundary.

1. Introduction

The Nain Plutonic Suite (NPS) in Labrador (northeastern Canada) comprises a diverse assemblage of 1350-1290 Ma anorogenic igneous rocks collectively referred to as an anorthosite-mangerite-charnockite-granite (AMCG) suite or complex [*Emslie*, 1978; *Emslie et al.*, 1994]. The anorthositic part of the NPS is among the largest areas of such rocks in the world. Large volumes of anorthosite within continental crustal plutonic settings such as the NPS are generally thought to result from the accumulation of plagioclase associated with mantle-driven magmatic processes, but the actual mechanism of formation, the tectonic setting, and the apparent limiting of such rocks to the Proterozoic are still matters of much debate [*Emslie*, 1978; *Ashwal*, 1993; *Emslie et al.*, 1994; *Longhi et al.*, 1999].

Seismic studies of large anorthosite terranes are few [e.g., *Musacchio et al.*, 1997; *Martignole and Calvert*, 1996]. Lithoprobe's Eastern Canadian Shield Onshore-Offshore Transect (ECSOOT) offered an opportunity to conduct a refraction/wide-angle reflection (R/WAR) seismic profile across the Nain Plutonic Suite, covering the central part of the NPS and its enveloping rocks (Figures 1 and 2). The objectives of the survey were (1) to determine the structure and characteristics of the Archean Nain Province crust and the Archean to Paleoproterozoic Churchill Province crust that bounds the NPS, (2) to examine the circa 1860 Ma continental collisional suture (Torngat Orogen) that welds together the Nain and Churchill provinces, and (3) to evaluate the setting and characteristics of the NPS in relation to its position as a suture-stitching anorogenic batholith emplaced some 500 Myr after terminal collision between the Nain and Churchill Provinces. *P* wave velocities and Poisson's ratios derived from the data gathered during the survey are used to determine the internal characteristics of the metamorphic and igneous rocks in the Nain transect and allow us to advance a reasonable geological picture of the crust in this part of Labrador.

2. Geological Setting

The Nain Plutonic Suite straddles the boundary zone between the Archean and Early Proterozoic SE Churchill Province and the Archean Nain Province (Figure 1). These two crustal blocks were sutured during development of the Torngat Orogen, the thermotectonics associated with the continental collision lasting from 1.86 to 1.74 Ga [*Bertrand et al.*, 1993; *Scott and Machado*, 1995]. The orogen is

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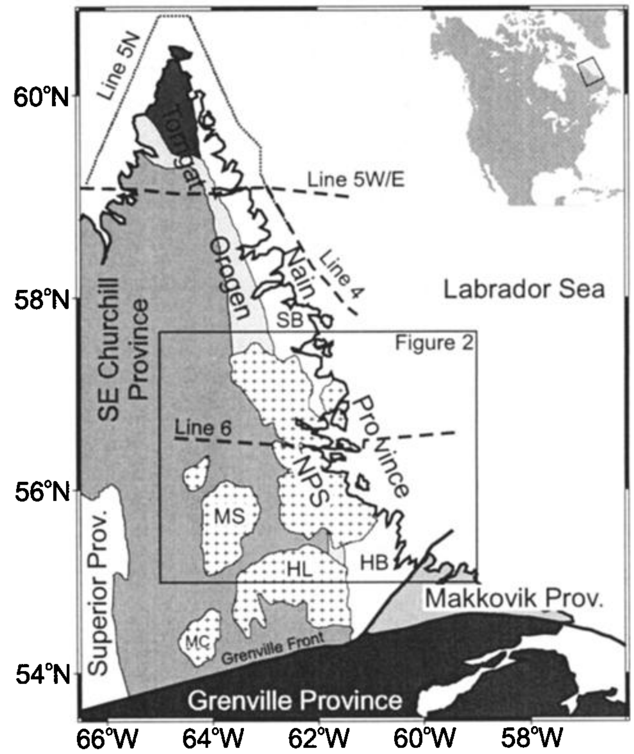

Figure 1. Tectonic provinces in NE Canada. The inset shows the location of the map relative to North America. Dashed lines show the location of ECSOOT 1996 R/WAR seismic lines relevant to this study, and the dotted line (line 5N) marks the outline of a seismic tomography experiment [*Funck et al.*, 2000]. AMCG complexes identified on the map are the Nain Plutonic Suite (NPS), the Harp Lake Intrusive Suite (HL), the Mistastin batholith (MS), and the Michikamau intrusion (MC). The Nain Province consists of the Saglek block (SB) and the Hopedale block (HB). The Tasiuyak gneisses (light shading) represent the axial zone of the Torngat Orogen along the western boundary of the Nain Province. A detailed map of the study area is shown in Figure 2. Prov. is Province.

characterized by a crustal root (total crustal thickness of >50 km) that is best preserved in the central Torngat Orogen but seems to disappear southward at the NPS [*Funck and Louden*, 1999; *Funck et al.*, 2000]. The metasedimentary Tasiuyak gneiss (Figure 2), which is the envelope to the western half of the NPS, probably represents an accretionary prism formed at 1.94-1.895 Ga [*Scott and Gauthier*, 1996] during eastward dipping subduction of SE Churchill crust beneath the Nain craton [*Wardle and Van Kranendonk*, 1996]. The western boundary of the Nain Plutonic Suite correlates closely with the Falcoz shear zone, which is one of the large shear zones associated with the oblique convergence of the Nain and Superior Provinces (trapping the SE Churchill Province between them) and the formation of the Torngat Orogen.

The Nain Plutonic Suite is exposed over an area of some 19,000 km^2 [*Hill*, 1988] and its tectonic setting is considered to be anorogenic [*Emslie et al.*, 1994]. It is one of several large AMCG complexes north of the Grenville front (Figure 1), which were formed during Mesoproterozoic Elsonian magmatism lasting from 1.45 to 1.29 Ga [*Emslie et al.*, 1994]. The NPS was emplaced between 1.35 and 1.29 Ga [*Connelly and Ryan*, 1994; *Hamilton et al.*, 1994; 1998] during the second major phase of Elsonian magmatism, and it consists of roughly equal areas of anorthositic and granitoid rocks (Figure 2). Ferrodiorites and other mafic rocks comprise <10% of the area [*Emslie et al.*, 1994]. Nd isotopic data in the Nain Plutonic Suite show a fundamental east-west subdivision with lower initial Nd values (ε_{Nd}) in the east, interpreted to delineate the approximate boundary at depth between the Nain and Churchill Provinces [*Emslie et al.*, 1994]. However, the surface trace of the Nain-Churchill boundary as delineated by gneissic enclaves in the NPS lies farther to the west (Figure 2). Recently, some anorthositic rocks previously thought to be part of the Mesoproterozoic NPS were dated as old as 2.1 Ga [*Connelly and Ryan*, 1999; *Hamilton et al.*, 1998]. These rocks were formed in the northeast of the NPS, but their full extent is unclear.

The Grenvillian orogeny (Figure 1) at about 1.1 Ga did not produce identifiable effects in the region of the Nain Plutonic Suite. However, the eastern part of the study area was affected by Mesozoic-Tertiary rifting [*Srivastava*, 1978; *Roest and Srivastava*, 1989] that separated the Nain Province from its Archean equivalent in Greenland during the opening of Labrador Sea. No AMCG complexes are known in the Archean block of Greenland.

3. Wide-Angle Seismic Experiment

3.1. Data Acquisition and Processing

Line 6 of the ECSOOT 1996 R/WAR seismic experiment is a 330-km-long E-W transect ranging from Labrador Sea, across the Nain Plutonic Suite, and into the SE Churchill Province (Figure 2). The seismic source was an array of six 1000 inch^3 (16.4 L) air guns towed by CSS *Hudson*. This source was used on the 161-km-long shot line in Labrador Sea and the fjord system north of Voisey's Bay, including a 8-km N-S offset of the line that was required for safe navigation. Along the line 1174 shots were fired at an interval of 1 min, resulting in an average shot spacing of 137 m. Water depth along line 6 varies between 18 and 487 m with an average of 139 m. Shots were recorded by 18 three-component 4.5-Hz geophone Reftek (Refraction Technology, Inc.) stations and by four analog ocean bottom seismographs (OBS) from the Geological Survey of Canada, equipped with a hydrophone and a 4.5-Hz vertical geophone. Station spacing was 13 km for the land stations and 23 km for OBS. Data from station 34 could not be retrieved. Air gun shots were located using smoothed GPS navigation, and the positions of land stations were determined by differential corrections to raw GPS locations.

Seismic data were converted to SEG-Y format (Society of Exploration Geophysicists) and coherency mixed across five traces using the method of *Chian and Louden* [1992] that accounts for time domain dip, followed by a band-pass filter (4 to 10 Hz). Record sections (Figures 3 and 4) for *P* and *S* waves are displayed with a reduction velocity of 7.5 and 4.0 km/s, respectively, and amplitudes are scaled by range.

3.2. Methodology

The data were modeled using a two-dimensional velocity model. For the setup of the model, receivers were projected onto a straight line extending from station 44 to the easternmost shot. To minimize the errors associated with the

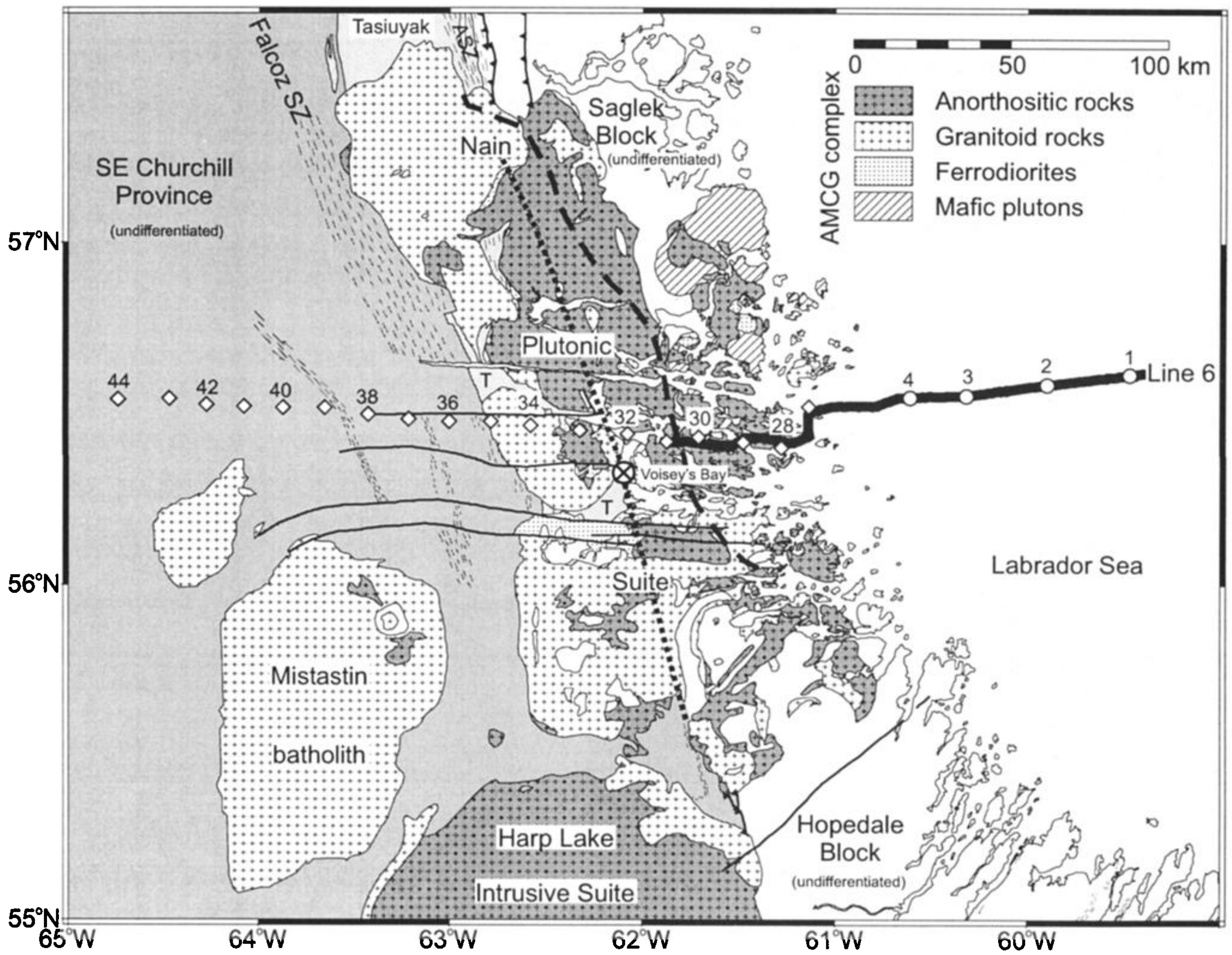

Figure 2. Geological map of the study area (simplified from *Wardle et al.* [1997]), showing the location of the seismic experiment with OBS and land stations marked by circles and diamonds, respectively. The solid bold line shows the shot line. The Voisey's Bay ore body is marked by a circle filled with a cross. Shear zones (SZ) are indicated by thin dashed lines, faults are indicated by solid lines (sense of displacement unknown), and thrust faults are indicated by solid lines with teeth representing the dip direction. The bold dashed line shows the inferred boundary at depth between the SE Churchill Province (shaded) and the Nain Province (white) based on isotopic data [*Emslie et al.*, 1994]. The dotted line is the projected surface trace of this boundary. ASZ is the Abloviak shear zone, and T's are Tasiuyak gneisses.

offset in the shot line, stations 44 to 28 were positioned by keeping their distance to station 44, and station 27 and the OBS were positioned using their distance to the eastern end of the line. This projection results in maximum errors of 2.5 km for the relative position of shots along the line. Land stations 27 to 31 were deployed close to the shot line at elevations ranging from 10 to 50 m. For the modeling, these stations were projected to the appropriate depth level in the water column assuming a water velocity of 1.5 km/s and a basement velocity of 6.0 km/s.

To determine the velocity structure of the crust and upper mantle along line 6, the programs RAYINVR and TRAMP [*Zelt and Ellis,* 1988; *Zelt and Smith,* 1992; *Zelt and Forsyth,* 1994] were used. In a first step, the observed P wave travel times were simulated by forward modeling. Later, amplitude information of the records was incorporated in the modeling procedure to match both observed travel times and amplitudes. The amplitude modeling was based on the computation of ray-theoretical synthetic seismograms and a qualitative comparison with the field records. The source signal used for the synthetic seismograms was extracted from stacking 100 coherent traces of a field record. Gaussian random noise has been added to the synthetic seismogram in Figure 3e to match the signal-to-noise ratio of the original record section. After developing a P wave velocity model, travel times of S waves were considered by using the P wave velocity model and assigning Poisson's ratios to crustal blocks from which the S wave velocities were calculated.

3.3. Seismic Data

P wave records (Figures 3b-d and 4) contain refracted phases through shallow sediment layers (P_{S1}, P_{S2}, and P_{S3}) the upper and middle crust (P_g), the lower crust (P_c), and the

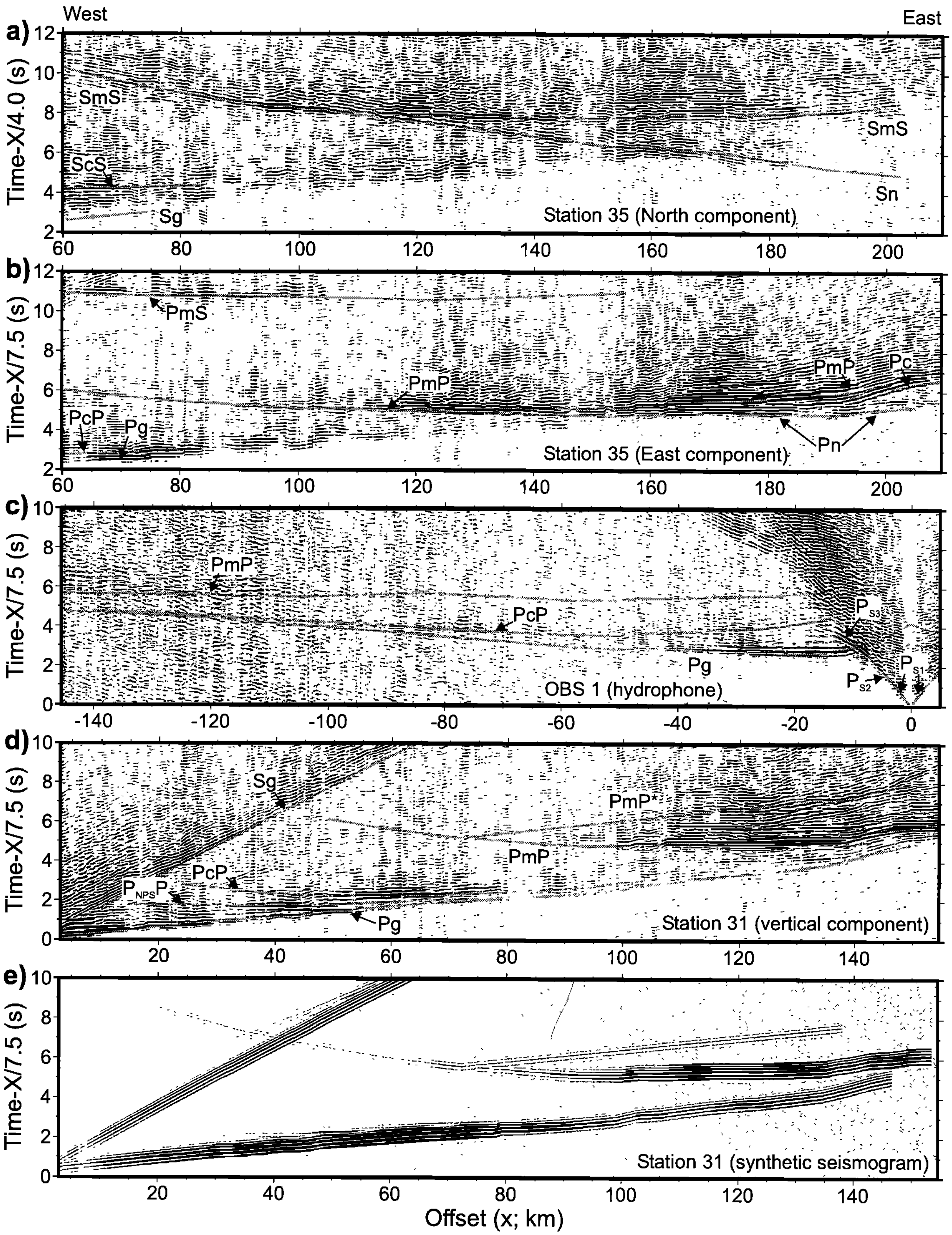

Figure 3. Record sections of (a) the north component of station 35, (b) the east component of station 35, (c) the hydrophone component of OBS 1, and (d) the vertical component of station 31 with computed travel times shown as gray lines. Processing includes coherency mix (five traces) and band-pass filter (4-10 Hz). Amplitudes are scaled by range. See text for description of phases. (e) Display of synthetic seismogram for the record shown in Figure 3d. Horizontal scale is shot-receiver distance, and vertical scale is the travel time using a reduction velocity of 4.0 km/s (Figure 3a) and 7.5 km/s (Figures 3b-3e).

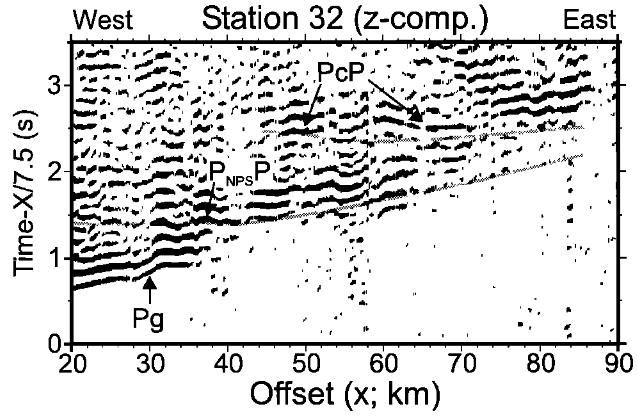

Figure 4. Detail of vertical component of land station 32 with reflections marked by gray lines. Horizontal scale is shot-receiver distance, and vertical scale is the travel time using a reduction velocity of 7.5 km/s. Processing includes coherency mix (five traces, 5.0-8.5 km/s) and band-pass filter (4-10 Hz). Amplitudes are scaled by range. The record section shows a strong reflection interpreted as from the base of the anorthositic plutons ($P_{NPS}P$) and from a midcrustal boundary (P_cP).

upper mantle (P_n). For error analysis the P_g phase was subdivided into P_{g1} and P_{g2}, corresponding to refractions in the batholith/upper crust and the middle crust, respectively. Reflections can be identified from the middle to lower crustal boundary (P_cP) and the Moho (P_mP). A medium-amplitude reflection in the area of the Nain Plutonic Suite precedes the P_cP phase and is interpreted as reflection from the base of the anorthositic portion of the NPS ($P_{NPS}P$).

OBS record sections show delays of the crustal phases due to thick sediment sequences at the continental margin. On OBS 1 (Figure 3c), three sediment layers can be identified based on refractions (P_{S1}, P_{S2}, and P_{S3}) with phase velocities of 2.1, 2.3, and 3.6 km/s, respectively. The signal-to-noise ratio for the OBSs is generally lower than for most land stations, as can be seen by comparison of the P_mP phase of a good OBS record section (OBS 1, Figure 3c) with a typical land station (station 31, Figure 3d). A number of land stations show arrivals behind the P_mP phase (Figure 3d). Some of these later arrivals are internal multiples associated with the high-velocity igneous rocks in the batholith, and others are related to diffractions or complex ray paths (named P_mP^*) caused by the geometry of the NPS. Similar complexities are also observed for some of the P_n phases (named P_n^*). There is also the possibility that some of the arrivals following the initial P_mP phase might be secondary PmP branches resulting from a complex Moho geometry (curvature) at the continental margin. However, the margin is too close to the eastern limit of the line to allow for a direct image of the Moho geometry in that area.

P_g phase velocities within the Nain Plutonic Suite range from 6.2 to 6.65 km/s, which is higher than velocities in the surrounding crust and, in particular, higher than in the underlying crust. This results in a local low-velocity zone at the base of the NPS. The impedance contrast between the high-velocity batholith and the underlying crust is high enough to generate a medium-amplitude reflection from the base of the anorthositic part of the NPS (Figures 3d and 4).

Shear wave phases observed on the record sections (Figures 3a and 3d) could be modeled with the P-to-S wave conversion occurring at the basement beneath the air gun shots. Upper crustal and midcrustal refractions (S_g) and Moho reflections (S_mS) are the most common shear wave arrivals, but there are also S wave reflections from the base of the plutonic complex ($S_{NPS}S$) and the midcrustal boundary (S_cS). One station also recorded a mantle phase (S_n, Figure 3a). P_mS reflections with mode conversion at the Moho are recorded on a number of horizontal components, although their amplitudes are generally weak and do not allow reliable picking of the arrival times. However, the east-component of station 35 (Figure 3b) did record an extraordinary P_mS phase with high amplitudes for shot-receiver offsets between 60 and 106 km.

4. Results

Examples of ray diagrams and the corresponding observed and calculated travel times are shown in Figure 5 for a selection of nine P wave and three S wave records along the line, out of a total of 21 recording stations. The layout of the experiment with shots restricted to the eastern half of the line results in reduced resolution in the west, in particular in the upper crust. With these limitations in mind, the velocity structure was kept as simple as possible to obtain a minimum structure model. Details of the resolution are addressed in the error analysis. Positions along the 330-km-wide line are given by their distance to the western border of the model.

4.1. *P* Wave Velocity Model

The crust in the SE Churchill Province to the west of the Nain Plutonic Suite is 39.5 km thick with a lower crustal thickness of 23 km (Figure 6a). Velocities in the upper and middle crust range from 5.9 to 6.4 km/s with a decrease of the velocity gradient at a depth of 5 km. The position at which the velocity gradient changes is defined as the boundary between the upper and middle crust, although no velocity discontinuity is associated with this boundary. Lower crustal velocities range from 6.55 to 7.0 km/s.

In the area of the Nain Plutonic Suite an 8 to 11.5-km-thick high-velocity body with a width of ~100 km is embedded in the upper and middle crust. This seismically well-defined body is interpreted as the anorthositic part of the NPS (consisting of anorthosite, leuconorite, and leucotroctolite) based on its physical properties (P and S wave velocities, density) and correlation with the surface geology. P wave velocities within the high-velocity body range from 6.12 km/s at the surface to 6.85 km/s at the base. The total crustal thickness within the NPS is reduced to 34 km with the midcrustal boundary located at a depth of 14-15 km. Velocities range from 6.3 to 6.5 km/s in the middle crust beneath the NPS and from 6.55 to 7.0 km/s in the lower crust.

At 240 km the anorthositic plutons either thin or disappear. Uppermost crustal velocities of 6.05 to 6.3 km/s are similar to those west of 240 km at the top of the batholith. Hence, on the basis of the P wave velocities the anorthositic part of the NPS may continue eastward as a 3-km-thick layer. However, the steep flank of the anorthositic complex at 240 km is reasonably constrained by P_g phases (e.g., stations 1, 27, and 30 in Figure 5) and complex P_mP^* and P_n^* phases (e.g., stations 32 and 44 in Figure 5). Midcrustal velocities are as high as 6.5 km/s, and the depth of the midcrustal boundary remains at 14 km east of 240 km. Moho depth decreases from 33 to <30 km at the eastern end of the line. At the easternmost

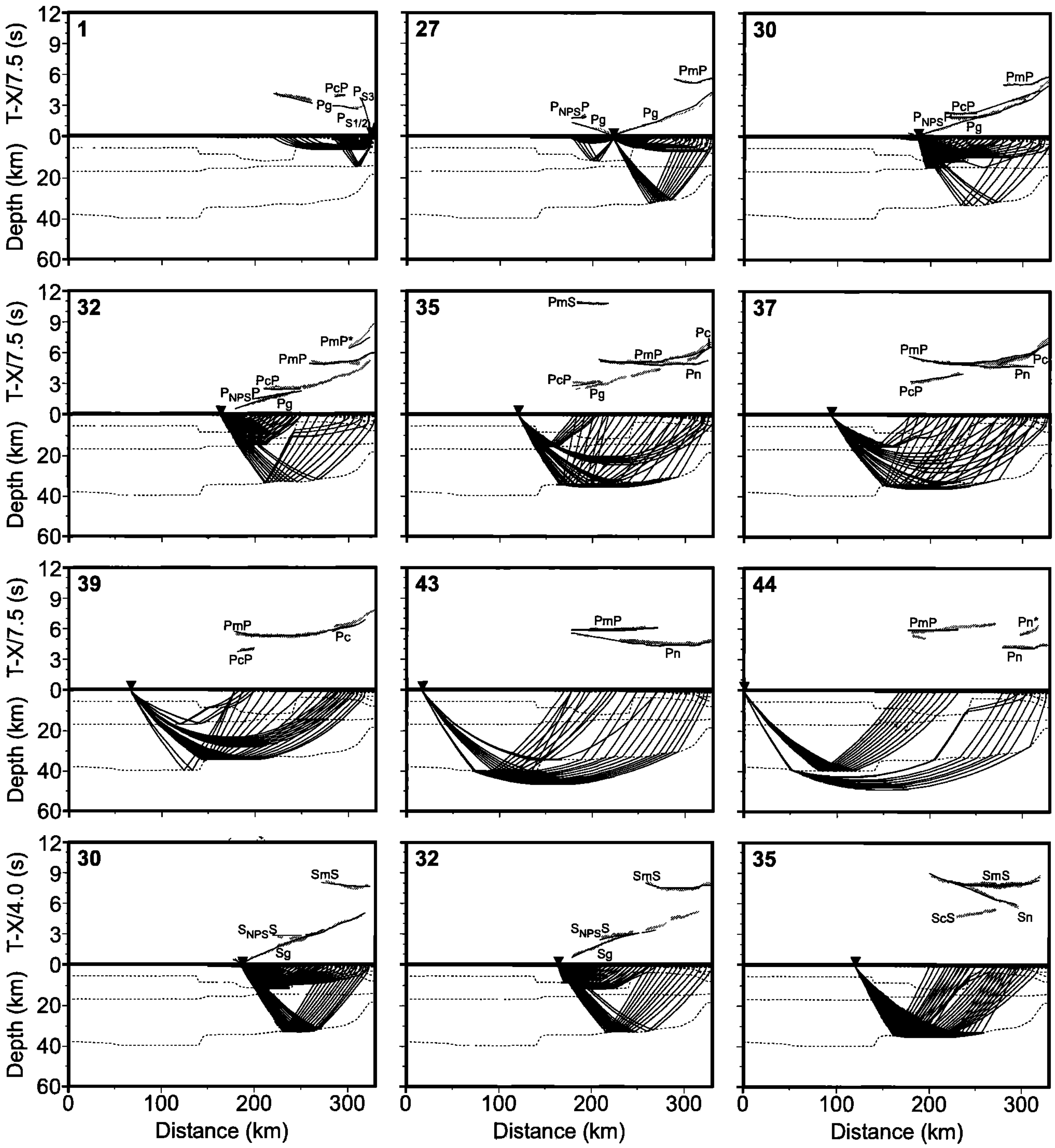

Figure 5. Comparison of observed and calculated travel times for a selection of nine stations (from a total of 21 stations), shown together with the corresponding ray paths. The upper nine sections show a compilation for *P* waves and the lowermost three sections display *S* wave phases. Observed data are indicated by gray lines with the line thickness representing pick uncertainty; calculated data are indicated by solid lines. Triangles mark the receiver location. Horizontal scale is the model position; a reduction velocity of 7.5 and 4.0 km/s has been applied for the travel times of the *P* waves and *S* waves, respectively.

end of the profile, there is an abrupt increase in sediment thickness. The uppermost sediment layer is <400 m thick with velocities of ~2.0 km/s. The second sediment layer continues down to a depth of 2.5 km with velocities of 2.1 to 2.9 km/s, and the lowermost sediment unit is up to 2 km thick with velocities of ~3.3 km/s. The total maximum sediment thickness is 4.5 km.

4.2. *S* Wave Velocity Model

The *S* wave velocity model is shown together with the Poisson's ratios in Figure 6b. West of ~130 km, no Poisson's ratios were computed because the quality of *S* wave arrivals decreases to the west. Together with the already lower resolution of *P* wave velocities in the west, the errors of the

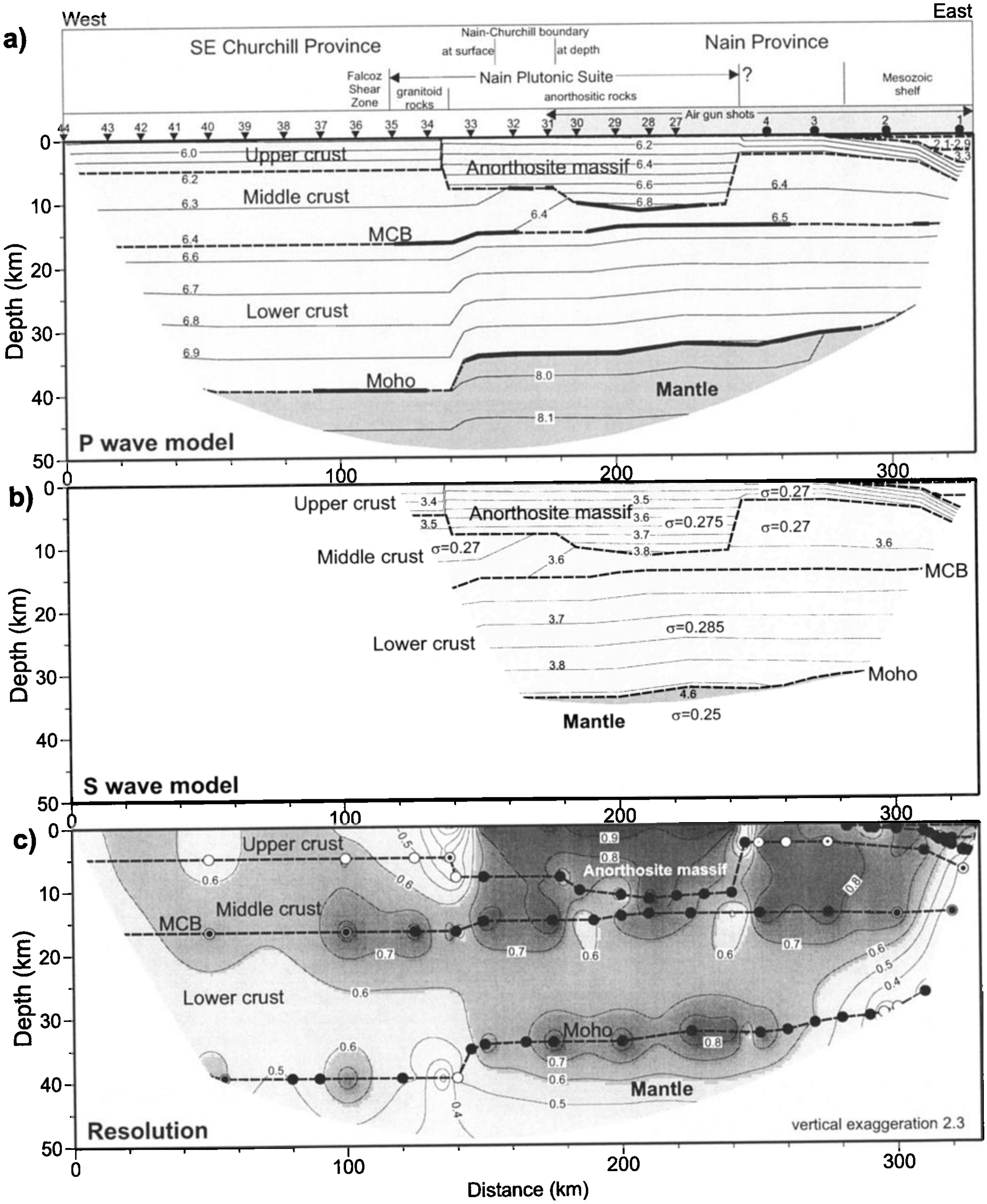

Figure 6. (a) P wave velocity model with a contour interval (thin solid lines) of 0.1 km/s in the crystalline crust and mantle. Numbers indicate velocity in km/s. The outer perimeter of the model with no ray coverage is omitted. Layer boundaries are drawn with dashed lines where they are not constrained by reflections and with bold solid lines elsewhere. MCB is the midcrustal boundary. The gray box above the model shows the segment of the line with air gun shots; circles and triangles mark the location of OBS and land stations, respectively. The simplified surface geology is indicated in the uppermost box. Vertical exaggeration is 2.3. (b) S wave velocity model with a contour interval of 0.05 km/s. The σ is the Poisson's ratio. (c) Diagonal values of resolution matrix for P wave velocity model displayed in contour format (0.1 contour interval) with shading for velocity nodes (darkness increases with diagonal value) and solid circles for interface nodes (radius of inner solid circle increases with diagonal value).

Poisson's ratios would be too large to allow any useful interpretation. In addition, no Poisson's ratios could be determined for the sediment column because the observed S phases travel through the sediments as P waves with the P-to-S conversion occurring at the basement.

The anorthositic intrusions of the Nain Plutonic Suite are characterized by a Poisson's ratio of 0.275±0.01, which is similar to the surrounding upper and middle crust with a ratio of 0.27±0.01. The lower crust was best modeled with a Poisson's ratio of 0.285±0.01, and for the mantle a value of 0.25±0.01 was obtained. Constraints on the shear wave velocities in the mantle are rather poor because only one land station (station 35) recorded a clear S_n phase.

4.3. Model Resolution and Uncertainty

The travel time residuals and the number of travel time picks for individual P and S phases are summarized in Tables 1 and 2. The root-mean-square (rms) travel time residual for P waves is higher by 13 ms than that for the S phases, which is related to the neglect of the westernmost stations (stations 36-44) for the S wave modeling as explained in section 4.2. Omission of these stations would result in a residual of 134 ms for the P waves, which is compatible with the 131 ms for the S waves.

In addition to limitations associated with the lack of air gun shots in the west, some ray-tracing difficulties occurred at the high-velocity rocks of the NPS embedded in low-velocity country rocks. Ray paths are very sensitive to the shape of the batholith: for instance, the P_mP* and P_n* phases that get deflected at the eastern border of the anorthositic rocks (stations 32 and 44 in Figure 5). Other phases affected by the geometry of the NPS are refracted phases through the plutonic mass and immediately below it. In some cases, these crustal refractions cannot be modeled, although they are observed on the record sections, and in other cases these phases should be visible according to the ray-tracing but cannot be identified on the records. This indicates that the structure of the NPS is probably more complex than shown in the velocity model (Figure 6a). Three-dimensional effects certainly contribute to some of the mismatches, because (1) the batholith is not strictly a two-dimensional feature, (2) the line is not exactly perpendicular to the strike direction of the NPS, and (3) there is an 8-km north-south offset in the line.

Table 1. Modeling Results for P Waves

Phase	Number of Observations	Average Pick Uncertainty, ms	RMS Misfit, ms
P_{S1}	16	77	68
P_{S2}	31	35	118
P_{S3}	11	35	48
P_{g1}	876	65	86
$P_{NPS}P$	295	69	101
P_{g2}	610	82	172
P_cP	383	93	136
P_c	448	99	131
P_mP	2349	107	154
P_n	618	123	168
All P phases	5637	95	144

Table 2. Modeling Results for S Waves

Phase	Number of Observations	Average Pick Uncertainty, ms	RMS Misfit, ms
S_{g1}	639	62	130
$S_{NPS}S$	140	104	165
S_{g2}	486	99	127
S_mS	663	136	126
S_n	96	117	140
All S phases	2024	101	131

The velocity model is determined with minimum structure avoiding large lateral variations. There are no P_g phases west of 120 km, preventing a direct determination of upper crustal and midcrustal velocities in that area. For this reason, velocities east of 120 km were extrapolated to the west. This did not result in large errors, and some reflections from the midcrustal boundary (e.g., station 39 in Figure 5) indicate a reasonable fit of the travel times. In the lower crust west of 130 km, the ray coverage is better than in the upper crust, and the data did not require lateral velocity changes. P_mP and P_n phases also indicate a constant Moho depth. In summary, the data though limited are consistent with a relatively homogenous crust west of the Nain Plutonic Suite.

East of 130 km, the ray coverage is improved with refractions through all crustal layers and the mantle, and with crossing ray paths in the east. Only the middle crust beneath the NPS is less defined because of the lack of refractions through that area. However, P_cP reflection travel times and the amplitudes of the $P_{NPS}P$ and P_cP phases provide partial constraints.

To determine the uncertainty of the velocities and boundary layers, single nodes of the model were perturbed to check for the sensitivity of the travel times on these changes. The velocity uncertainty in the crust is about ±0.15 km/s west of 130 km and ±0.10 km/s to the east. Sediment velocities have an uncertainty of ±0.15 km/s and mantle velocities of ±0.10 km/s. The outer perimeter of the ray coverage is included in Figure 6a, and layer boundaries defined by reflections are marked by bold lines. The base of the NPS is defined within ±1.0 km, but there are no $P_{NPS}P$ reflections defining it west of 160 km. The midcrustal boundary has an uncertainty of ±1.5 km and the Moho of ±1.0 km with a slightly higher value of ±1.5 km west of 140 km.

With respect to the lateral resolution, there is no exact image of the crustal thinning from 39 to 34 km around 140 km. The Moho step shown in Figure 6a may be shallower or steeper and up to 10 km farther to the west. While the eastern limit of the thick high-velocity anorthosite body is well defined at around 240 km, there is less resolution on its western border. With the failure of station 34 the seismic data only define the boundary to the west of land station 33 but to the east of station 35. For the velocity model (Figure 6a) the contact between granitoid and anorthositic rocks in the surface geology was chosen to define the western boundary of the anorthositic plutons. However, it is possible that the granitoid rocks are a veneer over the anorthositic rocks as in the Mistastin batholith (Figure 2).

The values of the diagonal of the resolution matrix for the velocity and depth nodes are shown in Figure 6c. Ideally, these values are 1, with values <1 indicating a spatial averaging of the true Earth structure by a linear combination of model parameters [*Zelt*, 1999]. Resolution matrix diagonals of > 0.5 to 0.7 indicate reasonably well-resolved model parameters [*Lutter and Nowack*, 1990]. Figure 6c shows that most parts of the velocity model are well resolved with exception of the upper crustal velocities around 140 km. Most depth nodes are well resolved with exception of the boundary between the upper and middle crust west of 130 km and east of 245 km. In the west this is related to the low coverage with P_g phases. In the east the interface is less resolved because it does not represent a first-order velocity discontinuity but a change of velocity gradient.

To estimate the error of the Poisson's ratios, their sensitivity upon perturbation was checked as well. Deviations of ±0.01 result in acceptable fits, whereas larger variations show misfits of travel times and phase velocities.

4.4. Gravity Modeling

To check the velocity model of line 6 for consistency with the gravity data, two-dimensional gravity modeling was performed using the Talwani algorithm [*Talwani et al.*, 1959]. Gravity data were provided by the Geological Survey of Canada. Bouguer anomalies are used for land areas, and free-air anomalies are used for offshore areas. The initial gravity model (Figure 7a) was obtained from conversion of *P* wave velocities to density using the empirical formula of *Ludwig et al.* [1970], which is given by

$$\rho=-0.00283v^4+0.0704v^3-0.598v^2+2.23v-0.7,$$

where v is *P* wave velocity in km/s and ρ is density in g/cm^3. The seismically unconstrained Moho depth at the Labrador continental margin in the east was modeled in relation to the increasing sediment thickness at the shelf. The model was continued to infinity at both the eastern and western border. The misfit between the observed and calculated data is up to 90 mGal. While the observed gravity decreases continuously between 0 and 280 km (from -40 to -80 mGal), there is an opposite trend in the calculated gravity. This trend is related to the eastward shallowing of the Moho and to the assumed high densities for the high-velocity anorthosite body of the NPS.

An anorthosite sample collected from the Nain Plutonic Suite close to land station 30 has a density of 2.73 g/cm^3 [*Muzzatti and Salisbury*, 1998], and other studies have shown that the density of anorthosite can be as low as 2.67 g/cm^3 [*McCaffree Pellerin and Christensen*, 1998]. Thus the density of the anorthosites is too high when using their velocities in the formula of *Ludwig et al.* [1970]. Using a density of 2.70 g/cm^3 for the anorthosite massif reduces the misfit to the

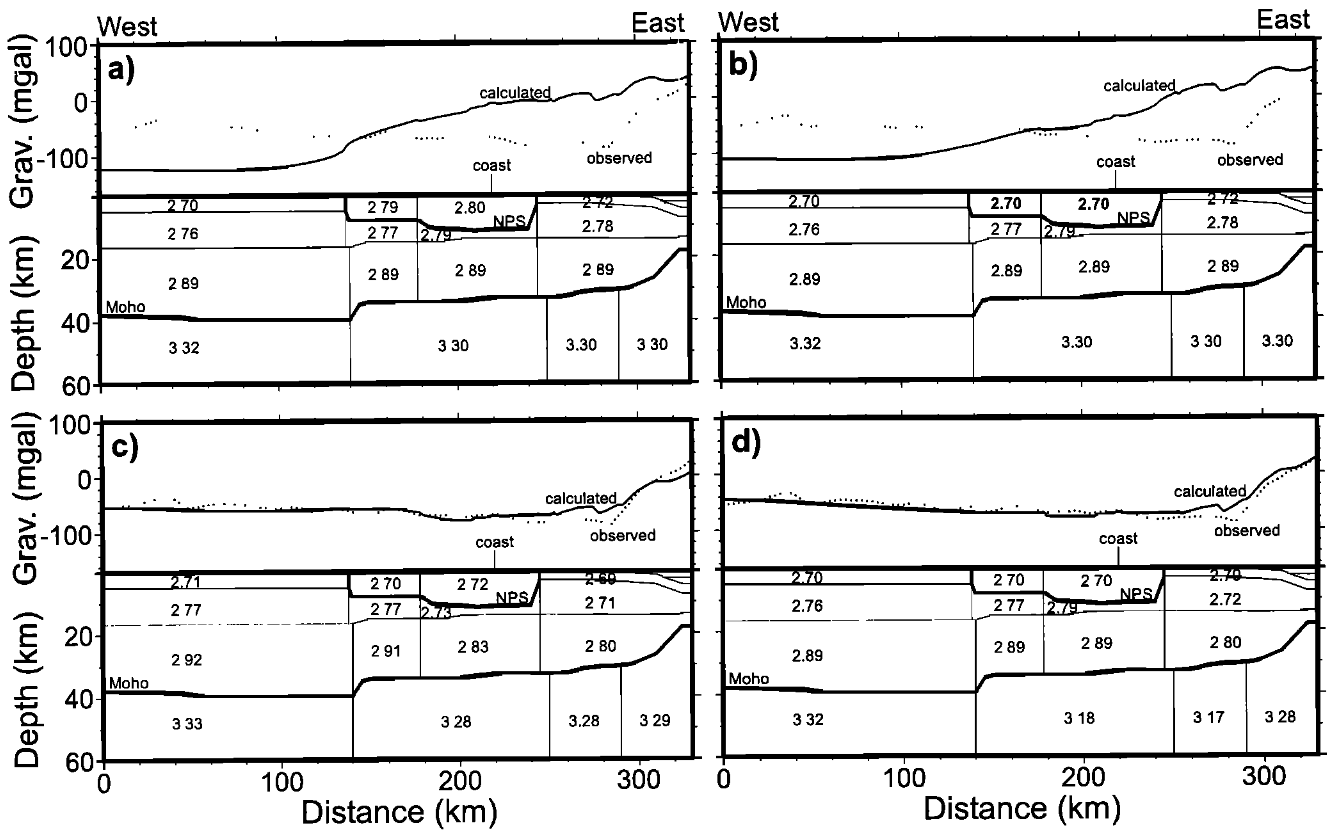

Figure 7. Two-dimensional gravity modeling for line 6. The observed gravity is shown by a dotted line and the calculated gravity by a solid line. Densities are given in g/cm^3. (a) Densities converted from *P* wave velocities using the formula of *Ludwig et al.* [1970]. (b) Same densities as in Figure 7a apart from reduced values for the anorthosites that do not follow the empirical relationship of *Ludwig et al.* [1970]. (c) Additional lateral density variations introduced to the crust with increased densities west of 175 km and reduced values to the east. (d) Lateral density variations within the mantle to fit the observed gravity but crustal densities east of 245 km still need to be reduced.

observed data (Figure 7b) but cannot account for the general trend. Lateral density changes within the crust or mantle are necessary to fit the data. Figure 7c shows a model where these lateral variations are primarily restricted to the crust. Densities within the eastern Churchill Province were increased by 0.01 to 0.03 g/cm^3 compared to the formula of *Ludwig et al.* [1970], while densities in the Nain crust (east of 180 km) were reduced by 0.03 to 0.09 g/cm^3. This model shows a good fit with the largest misfit occurring in the east at the continental margin. Since the Moho depth is not seismically constrained at the margin, these misfits can be accounted for by changing the Moho geometry.

In another model (Figure 7d) the necessary lateral density changes were applied to the mantle rather than to the crust. The mantle density between 140 and 290 km was reduced to 3.17-3.18 g/cm^3. However, a reasonable fit to the observed data could only be achieved by also reducing the crustal densities east of 245 km by up to 0.09 g/cm^3. With this adjustment a good agreement between observed and calculated gravity is achieved. On the basis of the gravity data alone, there are no strong arguments in favor of either of the end-member models in Figures 7c and 7d. Rather they show the possible range of gravity models that can explain the data.

5. Discussion

5.1. Crustal Composition

Seismic velocities can be analyzed in terms of crustal composition, in particular if both *P* and *S* wave velocities are known as for the eastern part of line 6. Seismic properties for the NPS between 140 and 240 km are similar to laboratory measurements from an anorthosite sample collected in the study area [*Muzzatti and Salisbury*, 1998]. This rock sample has a *P* wave velocity of 6.70 km/s at a pressure of 200 MPa (corresponding to a depth of 7.3 km) and of 6.89 km/s at 600 MPa. The Poisson's ratio is 0.280 and the density is 2.73 g/cm^3. *P* wave velocities obtained from the seismic experiment range from 6.1 to 6.8 km/s with a Poisson's ratio of 0.275, and the gravity modeling (Figure 7) yields the best results for densities between 2.70 and 2.72 g/cm^3. The combination of high *P* wave velocities, a high Poisson's ratio, and relatively low densities are characteristic features of these anorthosites.

The lower crust in the Nain Province has *P* wave velocities (v_p) between 6.55 and 7.00 km/s and a Poisson's ratio (σ) of 0.285. *Holbrook et al.* [1992] specify four lower crustal rock types that reasonably match these properties. These are anorthosite (v_p=6.81±0.31 km/s, σ=0.30±0.02), mafic granulite (v_p=6.86±0.27 km/s, σ=0.31±0.02), amphibolite (v_p=7.00±0.24 km/s, σ=0.29±0.02), and metapelite (v_p=7.17±0.36 km/s, σ=0.27±0.01). The densities of these rocks are 2.80±0.10 g/cm^3 for the anorthosite and 3.03 to 3.10 g/cm^3 for the other three rock types. Gravity models for line 6 (Figure 7) suggest rather low densities for the lower crust in the Nain Province (2.80-2.91 g/cm^3) in favor of an anorthositic composition. Other studies within the Nain Province [*Funck and Louden*, 1998; *Reid*, 1996] and in the conjugate Archean block of Greenland [*Chian and Louden*, 1992] indicate a fairly homogeneous velocity structure within the North Atlantic craton (Nain Province and Archean block of Greenland). However, there are differences in velocity between line 6 and other studies. *P* wave velocities on line 4 (Figure 1) are between 6.8 and 6.9 km/s in the lowermost crust, which is slightly lower than the 7.0 km/s found for line 6. However, this difference is close to the limit of resolution given the velocity uncertainties of ±0.1 km/s. The most striking difference is the Poisson's ratio for the lower crust, which is 0.26 on line 4 and 0.285 on line 6. Assuming that the original composition of the Nain crust in the area of the NPS was similar to the remainder of the North Atlantic craton, this would suggest a modification of the lower crust by the plutonism that resulted in an increase of the plagioclase/anorthosite content. This process would explain why the original intermediate composition of the lower Nain crust [*Funck and Louden*, 1998] was modified toward a more anorthositic content with slightly higher *P* wave velocities, a higher Poisson's ratio, and lower densities.

There is also the possibility that some of the assumed anorthosites in the lower crust are associated with a Proterozoic (2.1 Ga) anorthosite suite in that area [*Hamilton et al.*, 1998; *Connelly and Ryan*, 1999]. A Paleoproterozoic anorthosite belt in the northern Torngat Orogen [*Van Kranendonk and Wardle*, 1996] was probably exhumed from deeper levels of Nain crust, which is interpreted as indirect evidence of substantial amounts of anorthosite at middle to lower crustal levels. Rocks similar to this anorthosite belt continue for over 350 km southward, well into the region of the northern NPS [*Van Kranendonk and Wardle*, 1996].

Upper crustal and midcrustal *P* wave velocities within the Nain Province on line 6 (5.9-6.5 km/s) are similar to those on line 4 but again have a higher Poisson's ratio (σ=0.27) than on line 4 (σ=0.24). None of the rock samples collected in the Nain Province [*Muzzatti*, 1998] can explain both the *P* wave velocities and the Poisson's ratios. While *P* wave velocities on line 6 fit those of felsic gneiss, there is a disagreement between the Poisson's ratios, which are 0.27 on line 6 and around 0.24 for the felsic gneisses in the northern Nain Province [*Funck et al.*, 2000]. These high Poisson's ratios are consistent with addition of anorthosite at upper crustal and midcrustal levels, similar to the lower crust.

The lack of Poisson's ratios for the western part of line 6 prevents a detailed analysis of the crustal composition for the SE Churchill Province west of 140 km since *P* wave velocities on their own are rather ambiguous. The gravity model (Figure 7) uses higher densities in the SE Churchill Province than in the Nain Province, which may be an indication of a lower anorthosite content in the Churchill crust. However, this interpretation depends on the original crustal lithology and densities. *P* wave velocities in the SE Churchill Province on line 6 are very similar to those found one line 5W/E (Figure 1) some 300 km farther north [*Funck and Louden*, 1999]. In the north there is a good agreement between upper and mid crustal P-wave velocities and the results obtained from felsic gneiss samples [*Funck et al.*, submitted] which have densities between 2.66 and 2.89 g/cm^3. On the basis of this wide range of densities (bracketing the anorthosite densities of ~2.72 g/cm^3), there is no clear answer about possible anorthosite content in the upper 15 to 20 km of the crust in the SE Churchill Province.

5.2. Nain Plutonic Suite

5.2.1. Dimensions. Although the anorthositic portion of the Nain Plutonic Suite is seismically well defined between 160 and 240 km (Figure 6a) by reflections at its base, the

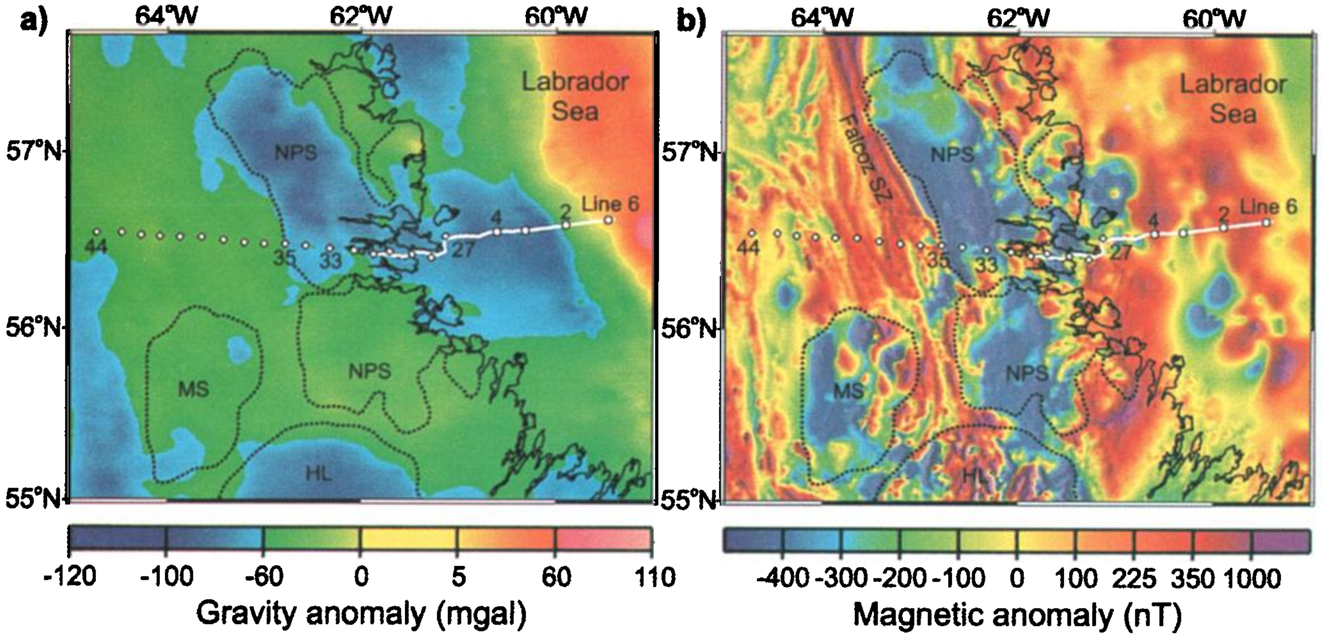

Plate 1. (a) Gravity map and (b) magnetic anomaly map of the study area. Gravity values on land are Bouguer anomalies and in offshore areas are free-air anomalies. Shading of the magnetic anomalies is derived from illumination from the east. Data courtesy of Geological Survey of Canada. Solid white lines show the shot line; circles and diamonds indicate the location of OBS and seismic land stations, respectively, with numbers representing the station number. The coast line is shown by solid black lines, and the outline of the AMCG complexes onshore is indicated by dotted lines. Abbreviations are NPS, Nain Plutonic Suite; HL, Harp Lake Intrusive Suite; MS, Mistastin batholith; and SZ, shear zone.

lateral extension is less well defined. In the west the boundary of the anorthosites was taken from the surface geology. In the east the boundary is in the offshore area where geology is largely unknown. The seismic data indicate a steep flank of the anorthositic complex between 240 and 245 km, but it remains unclear if the anorthosites possibly continue eastward as a shallow body since there is no significant change in velocity between the anorthosites and the upper crust east of 245 km.

The northern Nain Plutonic Suite is characterized by a roughly 20-mGal gravity low (Plate 1a) with anomalies between -60 and -80 mGal. At first glance, the continuation of the gravity low east of 245 km (Figure 7) suggests that the NPS may extend up to a distance of 280 km or close to OBS 2 (Plate 1a). However, an anorthositic intrusion east of 245 km cannot be thicker than ~3 km. Below that depth the velocities are too low to be associated with an anorthositic composition. It remains a possibility that the NPS continues eastward as low-density granitoid rocks. The argument against granitoid rocks is the relatively high Poisson's ratio of 0.27 in that zone [cf. *Holbrook et al.*, 1992]. In summary, the seismic data do not support a thick pluton east of 245 km. To explain the eastward continuation of the gravity low, we suggest that a substantial amount of anorthosite exists within the entire crustal column. This would explain the low density in the lower crust east of 240 km, which is set to 2.80 g/cm^3 in the gravity model (Figures 7c and 7d).

Some support for the interpretation that the anorthositic plutons of the NPS do not continue east of 245 km comes from the magnetic anomaly map (Plate 1b). The NPS is, in general, characterized by negative anomalies, which extend up to OBS 4 corresponding to the eastern border of the thick anorthosite complex in the seismic model. However, one has to be cautious with the interpretation of the magnetic data, since anorthositic rocks in the Harp Lake Intrusive Suite are characterized by both negative and positive anomalies. The same caution is necessary for the interpretation of the gravity data, as can be seen by the north-south variations within the NPS.

5.2.2. Tectonic model. A genetic model for the Nain Plutonic Suite was developed by *Emslie et al.* [1994] on the basis of geochemical data. They suggest that the generation of AMCG complexes starts with a hotspot or a mantle plume at the base of the crust. The heat associated with the basaltic magma results in anatexis at lower crustal levels with the extraction of granitoid partial melts that form the granitic plutons at higher crustal levels. Removal of the granitic partial melt leaves crustal residues lower in SiO_2 and enriched in plagioclase and pyroxene. These residues were assimilated by mantle derived basaltic magma, thus forming anorthositic magma. The rise of anorthositic magmas followed crustal paths preheated by the earlier passage of granitoid magmas. *Ryan* [1997] agrees with this model but points out that not all the granites in the NPS are early.

An adaptation of the model of *Emslie et al.* [1994] to line 6 is shown in Figure 8. The Paleoproterozoic Torngat Orogen sutures the entire Nain-Churchill boundary. North of the study area, where the orogen was not affected by the later anorthositic magmatism of the NPS and Harp Lake Intrusive Suite, is characterized by a preserved crustal root (Figure 9). The root beneath the central Torngat orogen was detected by a regional tomography study and was formed by transpressional

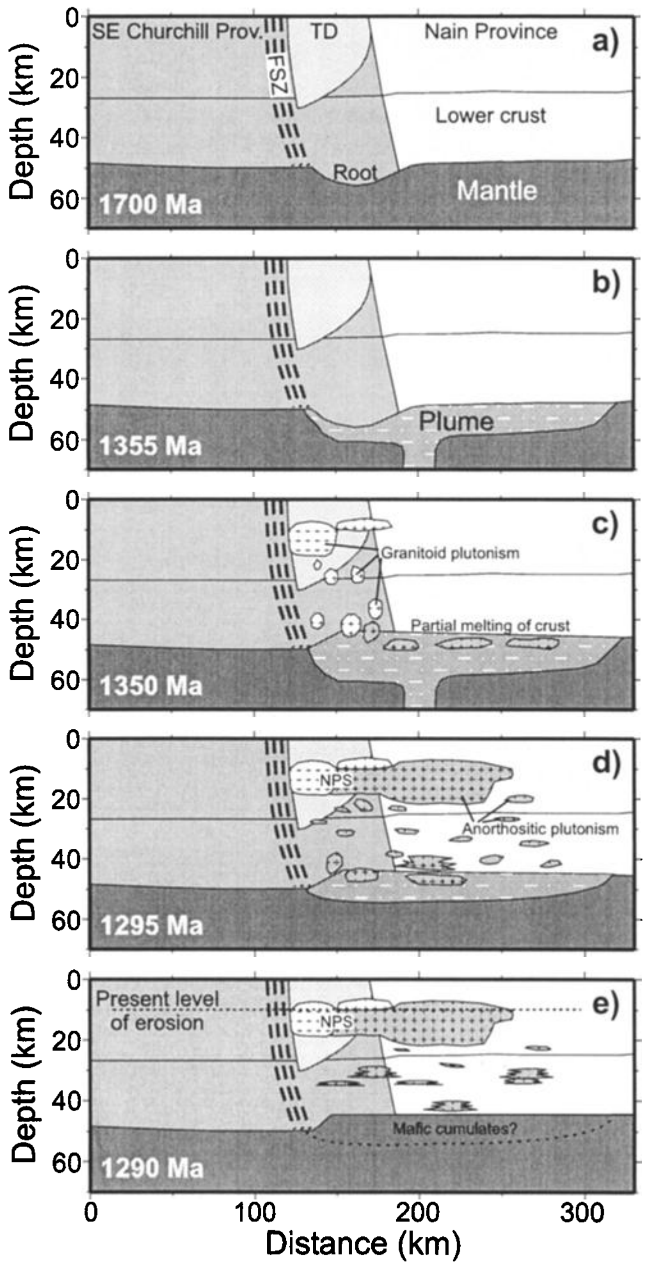

Figure 8. Development of the Nain Plutonic Suite (NPS). (a) A crustal root that formed at the suture between the SE Churchill and Nain Provinces during the Torngat orogeny was preserved until (b) the arrival of a mantle plume at the base of the crust. (c) Anatexis of the lowermost crust produced silicic melts from which the granitoid plutons of the NPS developed. (d) Assimilation of the residual mafic granulites by the plume magmas generated anorthositic magmas. (e) At the end of the anorogenic magmatism the crust beneath the NPS was thinned and modified to a more anorthositic composition at all crustal levels. TD is Tasiuyak domain, and FSZ is Falcoz shear zone.

shearing between the Abloviak and Falcoz shear zones during the orogeny at 1.8 Ga [*Funck et al.*, 2000]. These shear zones extend into the area of the NPS and we postulate in our model (Figure 8a) that a crustal root originally existed in the southern Torngats (future site of NPS). The original crustal thickness in the SE Churchill and Nain Provinces are assumed to be similar as defined by line 5 in the north [*Funck and Louden*, 1999]. There is presently no indication that the crust beneath the Torngat Orogen and Nain Province was already thinned prior to the NPS anorogenic magmatism.

At ~1355 Ma (prior to the onset of the NPS magmatism) a thermal plume of unknown derivation arrived, and the basaltic magmas fanned out laterally to generate a magma pond at the base of the crust (Figure 8b). Shortly afterward, granitoid magmatism was initiated by anatexis of the lower crust (Figure 8c), which resulted in crustal thinning. Large volumes of anorthositic magma developed at a later stage, when the residual mafic granulites were assimilated in quantity by mantle-derived basaltic magmas (Figure 8d). These parental magmas developed large volumes of anorthositic magma in the upper levels of lower crustal to uppermost mantle magma chambers [*Emslie et al.*, 1994]. Buoyant rise of these magmas resulted in the formation of anorthositic plutons at upper to midcrustal levels. At the end of the magmatism (Figure 8e), some remnants of cooled anorthositic magmas remain within

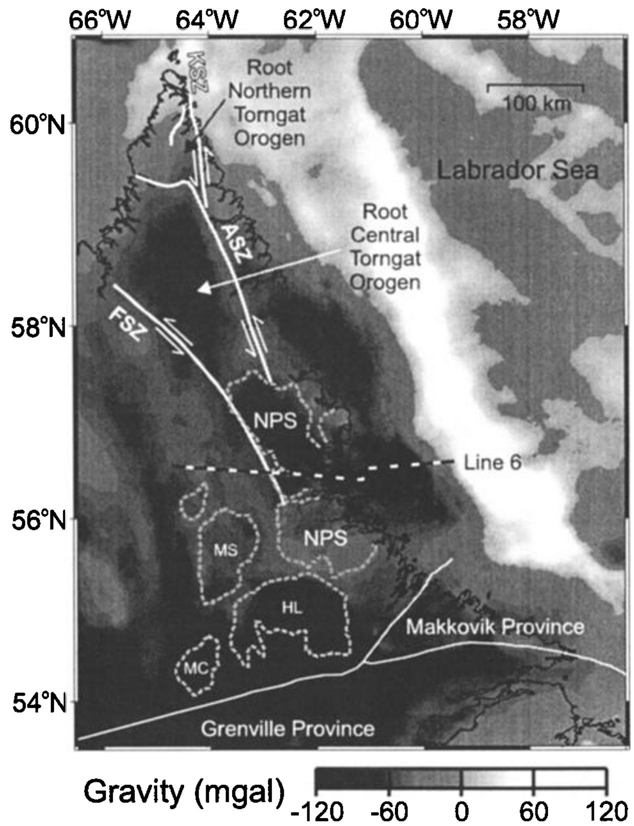

Figure 9. Gravity map of the area shown in Figure 1. Gravity values on land are Bouguer anomalies and in offshore areas are free-air anomalies. Data courtesy of Geological Survey of Canada. Dashed gray lines represent the outline of AMCG complexes (NPS, Nain Plutonic Suite; HL, Harp Lake Intrusive Suite; MS, Mistastin batholith; MC, Michikamau intrusion). The root of the central Torngat Orogen is characterized by a gravity low that is bounded by the Abloviak (ASZ) and Falcoz (FSZ) shear zones (white lines with arrows indicating the sense of displacement). The root beneath the northern Torngat Orogen is located to the west of the Komaktorvik shear zone (KSZ).

the crust beneath the NPS resulting in reduced crustal densities and higher Poisson's ratios compared to Nain crust distal to the AMCG complex (e.g., line 4 [*Funck and Louden*, 1998]). Erosion of some 10 km has stripped off the original roof and subsequently exposed the NPS to its present level.

There is no seismic evidence for remnants of a basaltic underplated layer. P_n phases require velocities of >8.0 km/s beneath the Moho, which are normal mantle velocities. However, this does not exclude the existence of an underplated layer that is thinner than the limit of the vertical resolution of the experiment, which is of the order of 1 km. Gravity modeling (Figure 7) allows some of the lateral density variations in the upper mantle beneath the NPS. Lower mantle densities might be appropriate if they represent mafic cumulates (mostly olivine and pyroxene) that must have been extracted from the initial magmas. These cumulates are seismically more like mantle than basalt and may be the reason for relatively high velocities beneath the Moho (8.0 km/s).

The velocity model (Figure 6a) reveals some other interesting correlations with the surface geology. On the basis of Nd isotopic data, *Emslie et al.* [1994] inferred the boundary between the Nain and SE Churchill Province underlying the NPS, as shown by the dashed line in Figure 2. The boundary occurs at 180 km and correlates with a thickening of the pluton from 8 km in the west to 11 km in the east (Figure 6a). This correlation suggests that the depth of the base of the AMCG complex may be related to the type of crust the magma intruded. One possible reason may be that higher crustal densities west of 180 km (as indicated by the gravity model, Figures 7c and 7d) allowed a rise of the magmas to higher crustal levels within the SE Churchill crust. Alternatively or in addition, the magmas generated with the Churchill Province could have had a lower density (Figure 7c) resulting in more buoyancy.

The Falcoz shear zone predates the NPS, and from the surface geology it is known that the NPS is restricted to the area east of the shear zone (Figure 2). In the study area the Falcoz shear zone subdivides the SE Churchill Province into reworked Archean gneisses to the west and Proterozoic Tasiuyak gneisses to the east, which are associated with the Torngat Orogen [*Van Kranendonk et al.*, 1994]. With the shear zone dividing these two units it seems conceivable that differences in the crustal composition may have prevented the formation of magma west of the shear zone. However, this interpretation would have difficulties in explaining the existence of the Harp Lake Intrusive Suite and Mistastin batholith (Figure 2) to the west of the Falcoz shear zone. In addition, gravity models suggest that the Tasiuyak gneiss does not extend down to the base of the crust [*Feininger and Ermanovics*, 1994]. We therefore assume that the lateral dimensions of the NPS were controlled by preexisting crustal structures rather than by lateral variations in the crustal composition.

The possible extension of the Torngat root into the study area could have limited the lateral spreading of the magma from the upwelling plume, suggesting that the plume head was located east of the root in the Nain Province. When the magma fanned out from the plume head, the crustal root possibly formed a barrier to this magma. Alternatively, the Falcoz shear zone itself could have prevented the spreading of the magma to the west. The synorogenic shear zones of the Torngat Orogen are interpreted as crustal scale features extending down to the Moho [*Funck and Louden*, 1999; *Funck et al.*, 2000]. Weak zones along the suture zone might have provided an early channel for ascending magma. This interpretation is consistent with the occurrence of the granitoid plutons in the western part of the NPS next to the Falcoz shear zone, whereas the anorthositic plutons are farther to the east. Since *Emslie et al.* [1994] generally regard the granites as earlier than the anorthosites, there seems to be support for the usage of the Tasiuyak gneiss as an early path for the rising magma. A preexisting crustal root east of the Falcoz shear zone would also provide large amounts of lower crustal material to form granitic magmas by anatexis.

Several authors have noted that AMCG complexes are localized near interfaces between Archean cratons and Paleoproterozoic orogens [*Gower and Tucker*, 1994; *Scoates and Chamberlain*, 1995]. *Scoates and Chamberlain* [1995] assume that preexisting crustal structures represent zones of lithospheric weakness that facilitated the generation and the ascent of anorthositic magmas. The reactivation of these terrane boundaries may have occurred in response to either regional compression or extension. *Ryan* [1997] assumes that the development of the NPS was possibly aided by extensional tectonics in the middle to upper crust. A series of east-west striking faults in the Voisey's Bay area (Figure 2) cuts all other structures in the Nain area and their age is therefore <1290 Ma. However, these faults may have a history of reactivation, based on older basement features being offset to differing degrees than features in the NPS cut by the same faults (B. Ryan, personal communication, 1999). The development of the faults may therefore bracket the NPS and may be a sign of synplutonic extension.

5.3. Comparison With Other Studies

Few other seismic studies have been reported for AMCG complexes. One detailed R/WAR seismic experiment across the Marcy anorthosite in the Grenville Province was analyzed by *Musacchio et al.* [1997] using both P and S wave velocities. P wave velocities within the anorthosite massif there are ~6.6 km/s and the Poisson's ratio is 0.28±0.01. Because of the high velocities of the surrounding granulite, no reflections from the base of the anorthosites were detected. *Musacchio et al.* [1997] estimate the thickness of the massif to be of the order of 10 km, comparable to the 8-11 km for the NPS on line 6. P wave velocities and Poisson's ratios of the Marcy anorthosite and the NPS are nearly identical within the given uncertainties. Similar to our study, *Musacchio et al.* [1997] found high Poisson's ratios in the lower crust beneath the anorthosites (σ=0.29), which they attribute to a high plagioclase content related to a modification of the lower crust by the emplacement of the anorthosites.

Reflection seismic data collected across the Morin anorthosite complex in the Grenville Province [*Martignole and Calvert*, 1996] reveal a spoon-shaped body extending down to a depth of 7 to 9 km. The crust underneath the anorthosites tends to be thinner than along the rest of the transect as for the NPS, and the basal layer is highly reflective, which is attributed to possible magmatic underplating associated with the anorthosite intrusion.

Ryan [1991] noted that the texture, mineralogy, and geochemistry of the granitic plutons of the NPS are similar to the rapakivi granites of Finland. The rapakivi magmatism postdates the Svecofennian orogeny by ~150-300 Ma and substantially thinned and stabilized the overthickened portions of the Svecofennian crust [*Puura and Flodén*, 1999]. Moho depth beneath the rapakivi intrusions is 41 km or up to

20 km less than in the surrounding crust [*Luosto et al.*, 1990]. *Puura and Flodén* [1999] correlate the initiation of the magmatic processes with the deepest roots of the orogen and therefore interpret the rapakivi magmatism as a process inherited from the Svecofennian orogeny. The Nain Plutonic Suite is located on the axis of the Torngat Orogen, where the crustal thickness can be expected to be larger than in the neighboring SE Churchill and Nain Provinces [*Funck and Louden*, 1999; *Funck et al.*, 2000]. However, the deepest parts of the Torngat root are located farther north in the central Torngat Orogen [*Funck et al.*, 2000], and we therefore do not see strong evidence for a close link between the root and the initiation of the later anorogenic magmatism. Other factors such as crustal extension or compression seem to be necessary in addition to the simple existence of a root. However, as discussed earlier, we see a possible link between a preexisting root and the lateral extension of the magmatic underplating.

In another gravity study, *Thomas* [1990] modeled a line across the northern NPS, and he found a maximum thickness of 19 km for the granitoid rocks whereas the anorthosite body was just over 3 km thick. *Thomas* [1990] used a density of 2.68 g/cm^3 for the granites and 2.72 g/cm^3 for the anorthosites. He noted that the modeled size of the anorthosite body could be increased by decreasing the density difference between granite and anorthosite. Densities of anorthosite complexes are as low as 2.67 g/cm^3 [*McCaffree Pellerin and Christensen*, 1998] and hence are similar to the densities used for the granitoid intrusions. The ambiguities inherent to gravity modeling can not distinguish between granites and anorthosites. *P* wave velocities and Poisson's ratios together allow for a better distinction between these rock types.

6. Conclusions

The R/WAR seismic line across the Nain Plutonic Suite shows a very distinctive high-velocity crustal zone, which is interpreted to be largely NPS anorthositic rocks. Low velocities in the country rocks allow for seismic detection of the base of the anorthosites at a depth of 8 to 11 km. The NPS is one of the few AMCG complexes where the base can be seismically determined. The thickness of the pluton changes across the isotopically inferred boundary between the underlying SE Churchill and Nain Provinces. In the west the pluton is 8 km thick, whereas it is up to 11 km thick in the Nain Province. The lateral change of depth is possibly related to differences in the density structure of the crust into which the anorthositic magmas intruded. Higher densities within the Tasiuyak gneisses could have allowed for a rise of the magmas to higher crustal levels compared to the Nain crust. Alternatively, the density of the magmas generated in the SE Churchill and Nain crust could have differed.

Comparison with other seismic studies within the Nain Province unaffected by the anorogenic plutonism shows that the crust underlying the NPS is some 4 to 6 km thinner in accordance with tectonic models for AMCG complexes. Anatexis of the lowermost crust is related to magmatic underplating as consequence of a plume [*Emslie et al.*, 1994]. There is no seismic evidence for remnants of an underplated layer, but gravity models (Figures 7c and 7d) show a reduced mantle density underneath the NPS which may be an indicator of an earlier plume. Poisson's ratios within the lower crust affected by the crustal thinning are compatible with those in the anorthositic plutons. Comparison with Nain crust outside the NPS shows increased Poisson's ratios at all crustal levels. It is therefore assumed that the AMCG magmatism added anorthosite/plagioclase to the crust. This interpretation is supported by relatively low crustal densities underneath the NPS as required by gravity models.

The proximity of numerous Mid-Proterozoic AMCG complexes to interfaces between Archean cratons and Paleoproterozoic orogens (e.g., NPS, Harp Lake, Mistastin) suggests a link between preexisting crustal structure and the origin and ascent of anorthositic magmas [*Scoates and Chamberlain*, 1995]. The seismic results allow some conclusions to be made concerning the siting of the NPS in combination with other data from the Torngat Orogen. In the central Torngat Orogen, seismic profiles show evidence for a preserved Paleoproterozoic crustal root east of the Falcoz shear zone [*Funck et al.*, 2000]. This root should have continued southward to line 6 but was likely corroded during the Mesoproterozoic anorogenic magmatism. Anatexis of the root by a thermal plume should have produced large amounts of silicic melts that ascended along weak zones at the suture between the Nain and Churchill Provinces and within the Tasiuyak gneiss at the eastern boundary of the Falcoz shear zone where the granitoid plutons of the NPS are concentrated. Large amounts of residual granulite could then be assimilated by the plume magmas resulting in plagioclase enrichment and the generation of anorthositic magmas.

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