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Crustal structure from the Lützow-Holm Bay to the inland plateau of East Antarctica, based on onshore gravity surveys and broadband seismic deployments

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ABSTRACT

Onshore gravity measurements were conducted over the inland traverse routes from Syowa Station (69.0S, 39.6E; SYO) to Dome-F (77.4°S, 39.6°E) by the Japanese Antarctic Research Expeditions (1992, 1997, and 1998). The crustal density structure between Lützow-Holm Bay (LHB) and the inland plateau was inferred from Bouguer gravity anomalies along these routes by assuming an initial model of the P-wave structure derived from seismic refraction surveys in the LHB. A decrease in the Bouguer anomalies down to −200 mGal toward the inland plateau indicate that the crust thickens from 38–40 km at LHB to 48–50 km beneath Dome-F. During the International Polar Year in 2007–2008, the GAmburtsev Mountain SEISmic experiment (GAMSEIS) deployed many broadband stations over the large highland on the ice sheet from the crest of the Gamburtsev Subglacial Mountains (GSM) to the vicinity of Dome-F. S-wave receiver functions and Rayleigh phase velocities determined using the GAMSEIS data indicate that the cratonic crust surrounding the GSM is 40–47 km thick. These thickness estimates agree with those beneath Dome-F from the JARE's gravity surveys and are also consistent with average Pre-Cambrian terrains. Beneath the GSM, the crustal thickness increases to 55–58 km and has been interpreted as providing isostatic compensation for the high mountain elevations. Accordingly, a long-distance crustal model extending over 3000 km from LHB to Dome-F and to the GSM was compiled, providing significant information for future studies of the tectonic evolution of the Antarctic craton.

1. Introduction

An examination of the high interior of East Antarctica has significant implications for numerous scientific issues, including the inner structure of the ice sheet, the sub-glacial environment, bedrock topography, and the structure and dynamics of the underlying crystalline crust and upper mantle. Geophysical and glaciological data collected across the high plateau of the East Antarctic ice sheet highlight the unique aspects of the continent and surrounding ocean, as well as general issues that are important to global Earth science. The East Antarctic continent could be a keystone of the supercontinent formation and break-up process in Earth's history. Additionally, the tectonic and thermal structure of the Antarctic lithosphere, which affect current ice sheet dynamics, are the most predominant problems to be solved by the research community. The locations of geophysical investigations focused on investigating the crustal structure of the East Antarctic continent are summarized in Fig. 1.

In order to constrain the structure and dynamics of the ice sheet and the underlying crust, several types of geophysical and glaciological surveys have been carried out close to Syowa Station (69.0S, 39.6E; SYO), in the Lützow-Holm Bay (LHB) — Mizuho Plateau area of Eastern Dronning Maud Land, East Antarctica (Fig. 2). As part of the geophysical measurements, onshore gravity surveys were periodically carried out since 1961 on the plateau in this region by the Japanese Antarctic Research Expedition (JARE; Yanai and Kakinuma, 1971). The previously acquired data on the LHB — Mizuho Plateau was compiled as 3D maps of gravity anomalies (Nagao et al., 1991). Onshore gravity surveys along traverse routes between SYO and the inland plateau station at Dome-Fuji (77.4°S, 39.6°E, altitude 3800 m; Dome-F) were not conducted until the 1990s because of the huge logistical requirements needed to accomplish the field operation on the

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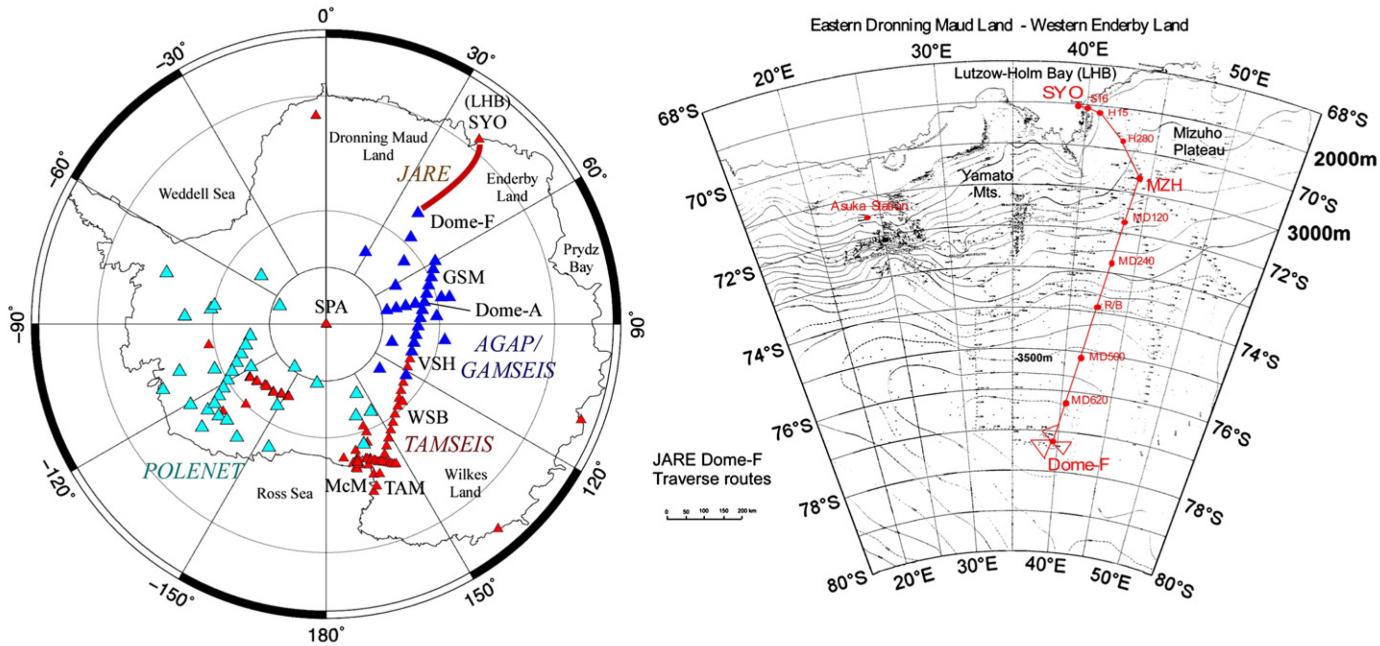

Fig. 1. (left) General location of inland traverse area for gravity surveys conducted by JARE (red thick line; from SYO to Dome-F), along with broadband seismic stations deployed by AGAP/GAMSEIS (dark blue triangles) and TAMSEIS (red triangles; extending from McM to near Dome-A). Some of the seismic stations deployed before the IPY 2007–2008 are also indicated by the other red triangles (mainly located in West Antarctica and near McM). The POLENET stations in West Antarctica deployed during the IPY are also indicated by the light blue triangles. Major location names are provided. Abbreviations are as follows. McM; McMurdo Station, TAM; Trans-Antarctic Mountains, WSB; Wilkes Subglacial Basin, VSH; Vostok Subglacial Highlands, SPA; South Pole Station, GSM; Gambursev Subglacial Mountains; SYO; Syowa Station, LHB; Lützow-Holm Bay. (right) Map showing the onshore traverse routes from Syowa Station (SYO) in the Lützow-Holm Bay region (LHB) to Dome-F station in East Antarctica. The same traverse routes were utilized by the three gravity surveys conducted by JARE-33, -38 and -39. Labeled red dots show the locations of the main observation points along the routes (MZH; Mizuho Station). Several other traverse routes conducted by JAREs are plotted on the map (black points). Contours indicate surface elevation.

highest portion of the plateau. The land traverse parties to Dome-F were primarily aimed at conducting deep ice core drilling operations almost down to the bedrock surface, over 3000 m in depth. However, the traverses were not conducted every year because of limited funding and logistic difficulties. Initially, the gravity surveys conducted along the traverses to Dome-F were focused on mapping the detailed bedrock topography to guide ice core drilling. In order to drill the ice core holes over the smallest flow point of the ice sheet, their locations were determined above the valleys before starting the construction of the main buildings.

Three onshore gravity surveys from LHB to Dome-F, covering a distance of about 900 km, were conducted by JARE (Fig. 1). The first traverse was performed by the 33rd JARE (JARE-33) from September 21 to December 29, 1992 (Kamiyama et al., 1994a,b). The second survey along the Dome-F inland route was conducted by the JARE-38 from October 7 to November 19, 1997 (Higashi et al., 2001). The third survey was conducted by the JARE-39 from December 22, 1997, to February 8, 1998 (Kanao and Higashi, 1999). For each ground traverse, Free-air and simple Bouguer gravity anomalies along the routes were obtained using the surface altitude of the ice sheet and the bedrock elevation from radio-echo sounding. Details about the methodology used to computed the gravity anomalies are summarized in Section 2.

During the International Polar Year (IPY 2007–2008; http://classic.ipy.org/), the 'Antarctica's GAmburtsev Province/GAmburtsev Mountain SEISmic' experiment (AGAP/GAMSEIS; http://epsc.wustl.edu/seismology/GAMSEIS/; Wiens, 2007) was conducted as an internationally coordinated broadband seismic deployment in the middle of the East Antarctic plateau (Fig. 1). More than 26 broadband seismometers were deployed over a huge highland on the ice sheet, from the crest of the Gambursev Subglacial Mountains (GSM; surrounding of the Chinese Plateau station of Dome-A — 79.6S, 77.4E; altitude 4000 m) to Dome-F at the western end of study area, making it possible to study the broad area between Dome-F and the GSM.

Broadband seismic studies using teleseismic events recorded by GAMSEIS have provided new information on the fine crustal structure in this region. These studies have also helped to constrain the origin of GSM and, more broadly, the structure and evolution of the Pre-Cambrian craton in East Antarctica. The GSM is one of the most enigmatic tectonic features on Earth, but until GAMSEIS was conducted, only limited constraints were available on the deep structure beneath the mountain range (Cogley, 1984; DeConto and Pollard, 2003). Buried beneath the thick ice sheet, the mountain is characterized by peak elevations reaching ~3000 m above sea level (Bell et al., 2011). New seismic data from GAMSEIS has allowed for more detailed investigations of the crustal structure beneath the GSM and the surrounding region (Hansen et al., 2010).

In this paper, a crustal density model is presented, which has been derived from the gravity anomalies obtained by the JARE surveys conducted in the 1990s. The Bouguer gravity anomalies were calculated by assuming a crustal density with an initial model of the P-wave velocity structure derived from seismic refraction studies within the LHB. After combining the crustal structure determined from broadband seismic studies that used AGAP/GAMSEIS data with those from the gravity surveys, a crustal model from LHB to Dome-F and the GSM is discussed. The crustal cross-section covers a distance over 3000 km in length across the middle of East Antarctica and provides unique information on the tectonic evolution of the Gondwana super-continent in Earth's early history.

2. Gravity anomalies from onshore traverses

2.1. Gravity measurements

Onshore gravity surveys from LHB to Dome-F were conducted by JARE in 1992, 1997, and 1998 (Table 1). During the first survey, gravity measurements along the traverse routes were conducted using a LaCoste & Romberg gravity meter (No., G-515) with a ~10 km interval

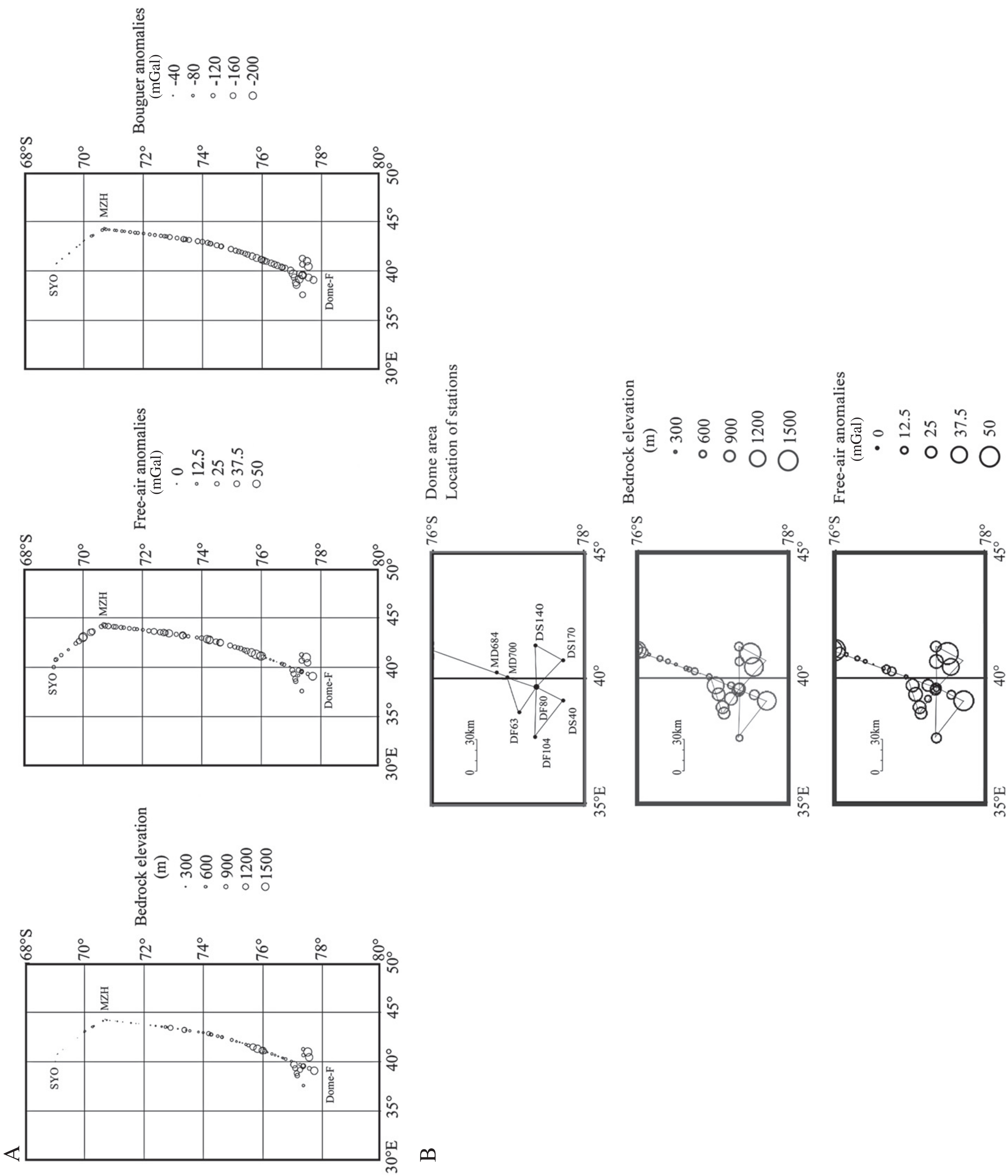
A
B
Bedrock elevation
(m)
300
600
900
1200
1500
Free-air anomalies
(mGal)
0
12.5
25
37.5
50
Bouguer anomalies
(mGal)
-40
-80
-120
-160
-200
SYO
MZH
Dome-F
68°S
70°
72°
74°
76°
78°
80°
30°E
35°
40°
45°
50°
Dome area
Location of stations
MD684
MD700
DS140
DS170
DF63
DF80
DF104
DS40
0 30km
76°S
78°
35°E
40°
45°

Table 1
Summary of the gravity surveys along the Dome-F traverse routes for JARE-33, -38, and -39.

JARE-33: Kamiyama et al., 1994a,b	
Observer:	Kokichi Kamiyama
Gravity meter:	LaCoste-Romberg G-515
Number of stations:	78 points
Date:	Sep. 21, 1992–Dec. 29, 1992
JARE-38: Higashi et al., 2001	
Observer:	Toshihiro Higashi
Gravity meter:	LaCoste-Romberg G-680
Number of stations:	67 points
Date:	Oct. 7, 1997–Nov. 19, 1997
JARE-39: Kanao and Higashi, 1999	
Observer:	Michio Kawabata (J-38)
Gravity meter:	LaCoste-Romberg G-515
Number of stations:	40 points
Date:	Dec. 22, 1997–Feb. 8, 1998

for 78 points (Fig. 2-A). The main purpose of the first traverse was to examine the bedrock topography under the ice sheet along the traverse routes, particularly in the vicinity of Dome-F, in order to guide where the station buildings should be located before the construction (Fig. 2-B). The station was located over the midpoint of a valley by comparing the gravity values with bedrock elevation data obtained from radio-echo soundings measured by an over-snow vehicle (Maeno et al., 1994). Details of the field gravity operation were summarized by Kamiyama et al. (1994b).

The second survey along the Dome-F inland routes was performed by the JARE-38 in 1997. The main purpose of the second traverse was to supply fuel to the station at Dome-F. Therefore, only limited geophysical investigations were accomplished. During this expedition, the gravity measurements were made using a G-680 type LaCoste & Romberg gravity meter with a ~10 km interval for 67 points (Higashi et al., 2001). The third gravity survey from the LHB to Dome-F was performed by the JARE-39 in 1998. The main purpose of this traverse was to perform glaciological studies of the chemical composition of the surface snow on the ice sheet, but additional gravity measurements were also conducted. The measurements were made using a G-515 type LaCoste & Romberg gravity meter at a total of 40 points (Kanao and Higashi, 1999). A detailed description of the measurements can be found in the corresponding initial reports (Higashi et al., 2001; Kanao and Higashi, 1999).

For the individual ground traverses, both the Free-air and simple Bouguer gravity anomalies along the inland routes were calculated using the surface altitude of the ice sheet and the bedrock elevation from radio-echo soundings. The approach taken for the anomaly calculations are described in the following section.

2.2. Calculation of anomalies

A reference point for the field measurements was taken at an absolute gravity point (IAGBN-A; g = 982,524.327 mGal; Kaminuma et al., 1997) at the gravimetric hut at SYO. Gravity anomalies were calculated after correction for Earth tides and gravity meter drift using the following formula. The Free-air anomaly (Δg) is given by:

$$\Delta g = g - \gamma + 0.3086H + 0.87 - 0.0000965H$$

where g and γ are the measured gravity value and the 'normal gravity' defined on the reference ellipsoid 1967 in mGal, respectively. H is the elevation of the gravity station in meters. The latter two terms in the above formula are atmospheric corrections. The simple Bouguer anomaly ($\Delta g'$) was calculated by:

$$\Delta g' = \Delta g - 2\pi G \rho_1 H$$

for an ice-free area, where G is the Earth's gravity constant and ρ_1 is the density of the bedrock (assumed to be 2.67 g/cm^3). For measurements on ice sheets, this equation becomes:

$$\begin{aligned}\Delta g' &= \Delta g - 2\pi G \rho_1 (H - I) - 2\pi G \rho_2 I \\ &= \Delta g - 2\pi G \rho_1 H + 2\pi G(\rho_1 - \rho_2) I\end{aligned}$$

where ρ_2 is the density of ice (assumed to be 0.90 g/cm^3) and I is the ice sheet thickness in meters. The bedrock density (2.67 g/cm^3) used to calculate the Bouguer reduction was adopted from the values for representative metamorphic rocks from the drilling cores at SYO. The rock velocities of these samples were measured by laboratory experiments (Yukutake and Ito, 1984). Table 2 summarizes the density, porosity, and major mineralogical constituents for several samples retrieved at SYO. There is a short segment of the traverse between SYO and Mizuho Station (70.1°S, 44.3°E, altitude 2245 m; MZH) where the bedrock is below the zero level. However, for simplicity, this can be ignored when calculating the gravity anomalies since it is small in scale compared to the large continental-scale structure discussed in this paper.

Table 2
The density, porosity, and major mineralogical constituents of the sample rocks from the drilled core at SYO (modified from Yukutake and Ito, 1984).

Sample no.	Depths (m)	Bulk density (g/cm3)	Apparent porosity (%)	Major mineralogical constituents
SP-B	2.1	2.686	0.3	Quartz (rich), feldspar, biotite
SP-C	4.4	2.612	1.0	Quartz (rich), feldspar, biotite
SP-D	9.1	2.822	0.6	Quartz, feldspar, magnetite, hornblende (rich)
SP-E	14.6	2.775	0.6	Quartz, feldspar (rich), biotite
SP-G	23.8	2.618	0.7	Quartz, feldspar, hornblende
SP-H	29.1	2.668	0.8	Quartz, feldspar, biotite, hornblende
SP-V	2.0	2.664	1.8	Quartz, feldspar, garnet

The gravity anomalies were calculated based on the concept of the 'gravity disturbance' by using surface elevation (altitude) from differential GPS positioning. The 'gravity disturbance' is a type of anomaly that has been reduced to the surface of the normal earth ellipsoid (Segawa, 1984). For the Bouguer reduction procedure in this study, the 'Bouguer disturbance' is based on the height measured above the GPS-defined ellipsoid, and we refer to this disturbance as the 'Bouguer anomaly'. In practice, this helps to avoid confusion when making interpretations of the results obtained from high elevation on the plateau.

Some gravity anomaly values with bedrock elevation along the traverse routes were already reported in JARE Data Reports and other Japanese publications (Higashi et al., 2001; Kamiyama et al., 1994a; Kanao and Higashi, 1999). However, all the acquired data were compiled in this paper using the same procedures described in the previous section. The obtained gravity anomalies were presented with the ice-sheet and bedrock elevation shown in Fig. 3. The upper part of Fig. 3 shows the surface elevation of the ice sheet and the

Fig. 2. *A.* Example of bedrock topography (left), Free-air gravity anomalies (middle), and simple Bouguer gravity anomalies (right) along the traverse routes from SYO to Dome-F from the JARE-33 survey. The measurement points are represented by the center of the circles and the diameter is proportional to the distinct values. *B.* Extended map around Dome-F for key station locations (upper), bedrock topography (middle), and Free-air gravity anomalies (lower) from the JARE-33 survey. The measurement points are represented by the center of the circles and the diameter is proportional to the distinct values.
a, b: modified from Kamiyama et al., 1994b.

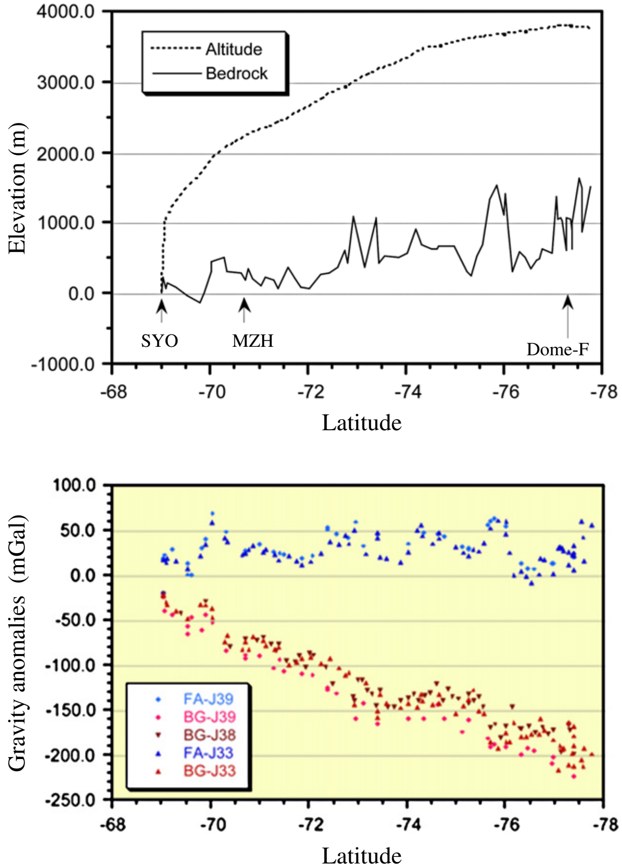

Fig. 3. (top) Surface elevation of the continental ice sheet (broken line) and bedrock topography (solid line) along the traverse routes projected onto the 40°E cross-section from LHB (SYO) to Dome-F. (bottom) Free-air gravity anomalies (blue symbols) and simple Bouguer gravity anomalies (red symbols) along the Dome-F routes projected to the 40°E cross-section for all the JARE traverses.

bedrock topography along the traverse routes. These traces are projected onto the 40°E cross-section from SYO to the Dome-F area. The lower part of Fig. 3 displays the Free-air and simple Bouguer gravity anomalies, assuming a reduction bedrock density of 2.67 g/cm^3, along the traverse routes, also projected onto the 40°E cross-section.

The obtained gravity anomalies include a ~3 mGal error estimate for the Free-air anomalies due to an uncertainty of ~10 m in the elevations determined by the differential GPS positioning between the measurement stations on the ice sheet and the fixed point at SYO. In contrast, the accumulated errors associated with the Bouguer anomalies are up to 10 mGal. These include the uncertainty from the GPS height measurements as well as those from the bedrock elevation determined by radio-echo sounding. The maximum estimated errors in the bedrock elevation were at most 100 m (Maeno et al., 1994), which leads to a ~7 mGal uncertainty in the Bouguer anomalies.

2.3. Correlation with bedrock elevation

The bedrock elevation was a significant parameter when deriving the crustal structure from gravity data because even small errors in the bedrock elevation might cause fluctuations in the gravity anomalies. The Free-air anomalies and bedrock elevation show a strong positive correlation, suggesting that the large density contrast at the bedrock-ice contact is the predominant influence on the Free-air anomalies. The relationship between bedrock elevation and Free-air anomalies just beneath Dome-F represent better correlation when compared with that for the whole traverse route from LHB to the dome region (Kamiyama et al., 1994b, by Japanese with English abstract). Dense measurement points in gravity surveys near Dome-F can support the evidence of better correlation between bedrock elevation and Free-air anomalies.

Additionally, a relationship between the bedrock elevation and the Bouguer anomalies also exists. It is useful to adopt a filtering technique in order to divide the gravity anomalies into the short and long periods of fluctuation. The feasible cut-off wavelength was considered to be about 100 km (Daggett et al., 1986) to identify the structural difference between regional and local scales. Fig. 4 shows the relationship between the bedrock elevation and the observed Bouguer anomalies along the Dome-F traverse routes. The obtained gravity anomalies correlate well (R = 0.8023) with that of the bedrock elevation from radio-echo soundings. It is recognized that there are correlations, to some extent, between these two parameters not only for long but also for short wavelengths. In deriving the density model, the short-wavelength variation in Bouguer anomalies was considered to be caused by changes in thickness of the shallow part of the structure, such as the depth of the first layer in the crust, while the long-wavelength variation of more than 100 km was attributed to the change in the density structure in the deeper part of the crust and the upper mantle (i.e. the depth change of the Moho and the Conrad discontinuities).

Based on the long-wavelength variation shown in Fig. 3, the Bouguer anomalies decrease gradually from SYO, reaching about −200 mGal beneath Dome-F, regardless of the high elevation of the bedrock topography in the 1000–1500 m beneath the dome area. This inclination of the long-wavelength trend suggests that the topographic variations of the crust–mantle boundary (Moho discontinuity) deepens toward the inland plateau of Dome-F. Another plausible reason could be that the assumed rock density (2.67 g/cm^3) for the Bouguer anomaly calculation might have lateral heterogeneities along the traverse routes, particularly in the deeper part of the crust. These findings imply that the density of the deep crust, as well as the crust–mantle boundary, gradually changes along the traverse routes towards the inland plateau.

3. Crustal structure modeling

3.1. Initial velocity model by seismic refractions

The first Deep Seismic Surveys (DSS) on the Mizuho Plateau were conducted by JARE-20, -21, and -22 (1979–1981; Ikami et al., 1984). The velocity structure of the crust and uppermost mantle along the

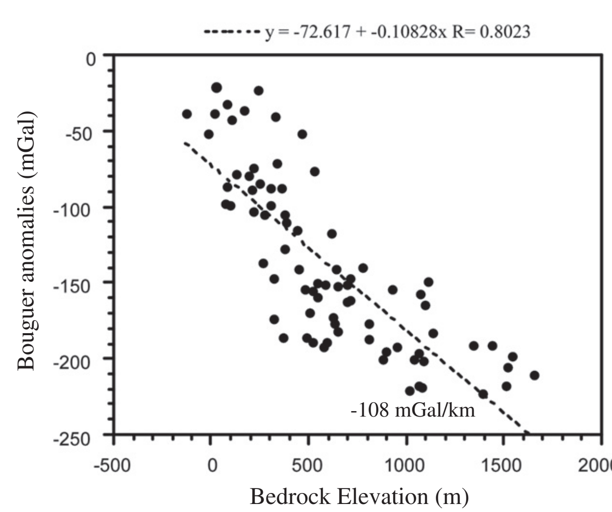

Fig. 4. Relationship between the bedrock elevation and the Bouguer anomalies for a reduction density of 2.67 g/cm^3. The regressive curve (broken line) indicates a gradient of −108 mGal/km.

seismic profile between SYO and MZH was achieved from wide-angle reflection/refraction analyses (Ikami and Ito, 1986; Ito and Ikami, 1984). In addition to the DSS, geophysical measurements by gravity and aeromagnetic surveys were conducted along the same routes (Kaminuma and Nagao, 1984; Shibuya et al., 1984).

The crustal density structure between SYO and MZH was estimated using the gravity data acquired over the ice sheet in order to fit the seismic P-wave velocity model from the DSS (Ito and Ikami, 1986). The Bouguer gravity anomalies were calculated using the method of Talwani et al. (1959) by assuming a crustal density model with a layered structure which fit the observed gravity anomalies. We referred to the previous density model of the northern Mizuho plateau (Ito and Ikami, 1986) to construct the new crustal structure toward Dome-F in this study.

In the austral summer 2000, another DSS with denser station distribution and new instruments was conducted on the continental ice sheet of the Mizuho Plateau by JARE-41 (Fig. 5). The seismic deployment (SEAL-2000 DSS) was conducted as one of the multidisciplinary geoscience projects called the "Structure and Evolution of the East Antarctic Lithosphere (SEAL)" (Kanao et al., 2004, 2011). A total of 3300 kg of dynamite charges were detonated at seven shot points along the traverse routes, generating enough seismic energy to acquire information about the regional velocity models and reflectors in the crust. Details of the survey specifications and operation procedures are given by Miyamachi et al. (2001, in Japanese with English abstract).

The SEAL-2000 DSS was carried out along the same traverse route across the Mizuho Plateau that was covered by the Pre-SEAL exploration in 1980. However, more accurate velocity models by wide-angle reflection/refraction analyses (Yoshii et al., 2004) and reflection cross-sections (Kanao et al., 2011) were obtained by the SEAL-2000 DSS. Wide-angle reflection/refraction travel-time analyses applied to the SEAL-2000 data (Yoshii et al., 2004) revealed that the Moho depth varies from 38 to 42 km along the seismic profile, with velocities in the upper, middle, and lower crust and the uppermost mantle, as 6.08–6.26, 6.45, 6.56 and 8.03 km/s, respectively (Fig. 5).

The density model of the crustal structure between SYO and the inland plateau at Dome-F developed in the current study was derived based on the P-wave velocity model of the Mizuho plateau from SEAL-2000 DSS. The crustal structure of the southern part of the inland plateau, from MZH to Dome-F, was derived only from the gravity data measured by the inland traverses in 1992, 1997, and 1998 (for JARE-33, -38 and -39) as mentioned in Section 2. Although a density model for the 1992 survey was proposed by Kanao et al. (1994) using a similar method, the current study determines the crustal structure by combining all the gravity data retrieved by the onshore traverses to Dome-F.

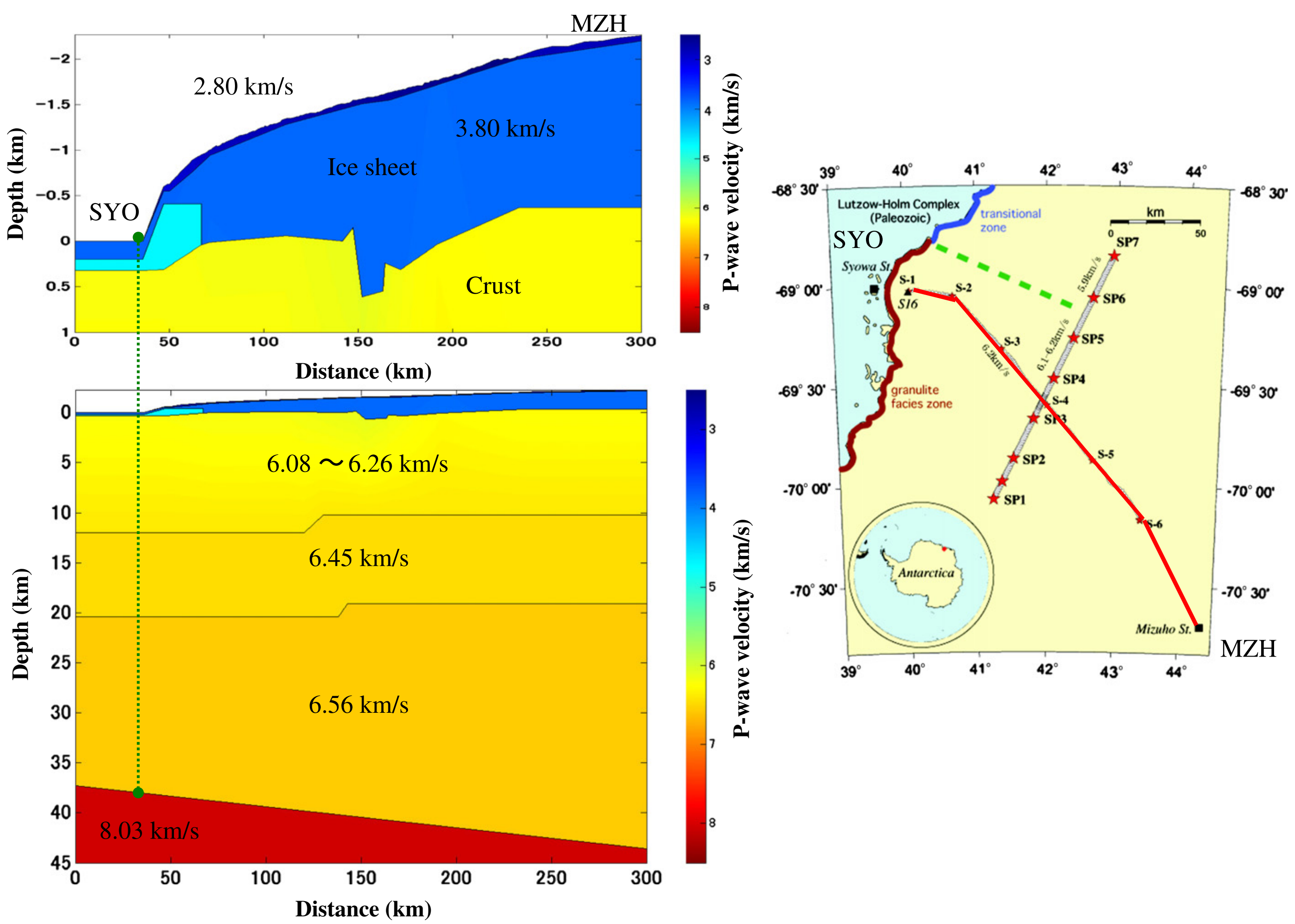

Fig. 5. (left) P-wave velocity model of the crust, uppermost mantle, and the overlying ice sheet on the Mizuho Plateau, derived from refraction experiments by the SEAL-2000 DSS (modified from Yoshii et al., 2004). The average depths of the discontinuities between the upper and lower crust (Conrad) and between the lower crust and upper mantle (Moho) are 20 km and 40 km, respectively. (right) Location of the DSS profiles on the Mizuho Plateau. The SEAL-2000 profile is indicated by the red line, which was the same location as the Pre-SEAL DSS in 1980. The NE–SW seismic profile (SP1–SP7) indicates the DSS conducted by SEAL-2002.

3.2. Modeling of the density structure

The density model of crustal structure extends from LHB (SYO) to MHZ and towards the southern continental area around Dome-F, 900 km inland from the coast. The structure between SYO and MHZ was initially presented by Ito and Ikami (1986), followed by a modification to fit the seismic velocity model by Yoshii et al. (2004). The rest of the southern structure between MHZ and Dome-F was derived by fitting the calculated gravity anomalies to the observed ones. The simple Bouguer anomalies were calculated by assuming a layered structure as an approximation to n-sided polygons (Talwani et al., 1959) to fit the observed gravity anomalies.

The depths of discontinuities for each layer were assumed using the P-wave velocity structure between SYO and MZH derived from the SEAL-2000 DSS (Yoshii et al., 2004). In the Yoshii et al. (2004) model, the average depths of the discontinuities between the upper and lower crust (Conrad) and the lower crust and upper mantle (Moho) were 20 km and 40 km, respectively. The upper crust has a layered structure where the uppermost, surface layer has a P-wave velocity of 6.08–6.26 km/s, and the layer above the Conrad discontinuity has a velocity of 6.45 km/s. The final density model (Fig. 6-A) was assumed to have four layers in order to constrain the structure,

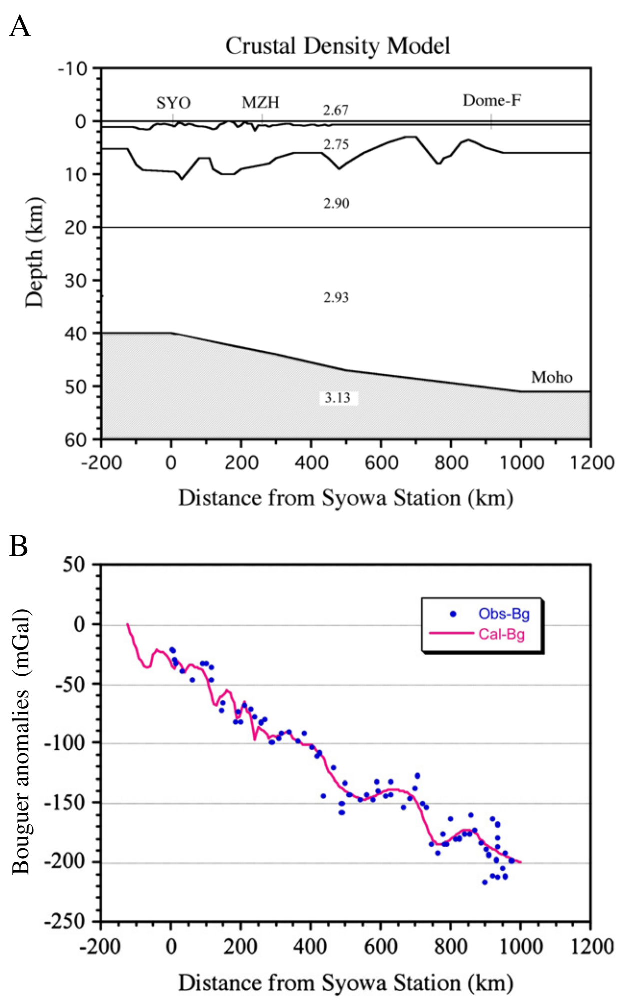

Fig. 6. *A.* The estimated crustal density model from LHB (SYO) to Dome-F. The structure between SYO and MZH was derived from Bouguer anomalies on the basis of the P-wave velocity structure shown in Fig. 5. The rest of the structure between MZH and Dome-F was modeled from only gravity data. The density of each layer in g/cm^3 is indicated. *B.* Observed Bouguer anomalies (blue circles: Obs-Bg) and calculated anomalies from the density model in Fig. 6-a (pink line: Cal-Bg). The horizontal axis indicates the southward distance from SYO.

and considers both short and long period variations in the Bouguer anomalies, as discussed in previous section. The short-wavelength variation in Bouguer anomalies was considered to be caused by the undulation of the discontinuity between the first and second layers, and this boundary was arbitrarily adjusted to achieve a fit between the observed and calculated gravity anomalies.

We assumed the short period fluctuations could be caused by lateral heterogeneity at upper crustal depths. The densities of the first three layers in the upper crust were assumed to be 2.67, 2.75, and 2.90 g/cm^3, respectively, based on the surface metamorphic rock velocities at SYO (Yukutake and Ito, 1984; Table 2) and the velocity model from SEAL-2000 DSS (Yoshii et al., 2004; Fig. 5). To produce the density model, the same layer boundaries were taken from the DSS P-wave velocity model by Yoshii et al. (2004). The densities for each layer were converted from P-wave velocities on the basis of laboratory measurements by Barton (1986). Densities of the first and the second layers, in particular, are taken from the actual retrieved rock samples by Yukutake and Ito (1984). On the other hand, long-wavelength variation over 100 km may be affected by the depth changes in deeper discontinuities, specifically the Moho (between the lower crust, with a density of 2.93 g/cm^3, and the uppermost mantle, with a density of 3.13 g/cm^3). It was expected that the effect from the Moho variations might be larger than those from the Conrad discontinuity because the density difference between the two adjacent layers was larger across the crust–upper mantle boundary.

The calculated anomalies were fit to the observed ones, and the final crustal density model obtained is shown in Fig. 6-A. The Moho discontinuity deepens slightly from SYO (40 km) toward Dome-F (48–50 km), while the Conrad discontinuity is at an almost constant depth of 20 km. The discontinuity depths between the surface and the second layer in the upper crust were varied between 6.5 km to 14 km in the modeling to fit the observed anomalies. Fig. 6-b shows the observed and calculated Bouguer anomalies from the assumed density model. In our analysis, we assumed the isostatic compensated model of the crust so there is no difference between the modeled crustal thickness and a Moho defined by isostasy.

4. Discussion

4.1. Accuracy of gravity anomalies and structure

The accumulated errors in calculating Bouguer anomalies are up to 10 mGal, which include the uncertainty from GPS height measurements and those from bedrock elevation by radio-echo sounding. Since the radio-echo soundings were accompanied by multiple echoes, the interpretation was not simple. The uncertainty of bedrock elevation determined by the radio-echo soundings was up to 100 m (Maeno et al., 1994), which leads to an uncertainty in the Bouguer anomalies of ~7 mGal. This results in an estimated uncertainty in the Moho depth of about 0.3 km. This seems to be the most predominant cause of the estimated errors.

Additionally, if the actual bedrock elevation is deeper than that estimated, the depth of the Moho could be somewhat shallower and still fit the Bouguer anomalies based on the crustal density model. Nagao (1984) noted a discrepancy in bedrock elevation estimates between those determined by radio-echo soundings and those from gravity anomalies along the traverse routes, and Kudo and Nagao (1994) pointed out that the gradient in Fig. 4 contradicts the theoretical one. For the simple case of a layered half space model, the gradient is $2\pi G(\rho_1-\rho_2)$I, where G is the gravity constant, and ρ_1 and ρ_2 are the densities of bedrock and the ice, respectively. If we assume $\rho_1=2.67$ and $\rho_2=0.90$ g/cm^3, the gradient is about -80 mGal/km. However, the observed gradient is steeper, approximately -108 mGal/km. In order to explain this difference, it was suggested that the average density of the crust is much larger than 2.67 g/cm^3. Another possible explanation were errors in the ice thickness estimation. If this is the case, the ice sheet thickness should be larger than that observed. Additionally, the shallower bedrock elevation determined by radio-echo soundings could also be explained by the existence of a mixed layer of ice and moraine over the bedrock.

Regarding the density assumption used in the calculation of the Bouguer anomalies, we adopted a value of 2.67 g/cm^3 for the surface bedrock density (for the first layer of the density model). To test the dependence on this value, the reduction density was varied from 2.60 to 2.80 g/cm^3. Fig. 7 shows the Bouguer anomalies along the whole route to Dome-F obtained with the different reduction density values. An offset of 13 mGal in Bouguer anomalies beneath the dome area corresponds to the density variations in 0.1 g/cm^3. Therefore, the estimated change of the Bouguer anomalies was at most ~30 mGal between the assumed value of 2.67 g/cm^3 for the surface layer in this study, and 2.90 g/cm^3 in the middle–lower crustal depths of the Mizuho Plateau. As the average density of the whole crust is about 2.83 g/cm^3 (Christensen and Mooney, 1995), an assumption of the density in the deeper part of the crust along the routes could be in acceptable range of the continental crust.

4.2. Crustal thickness from LHB to Dome-F

The Bouguer anomalies observed in this study are about −200 mGal beneath Dome-F, 900 km inland from the continental margin. By fitting the calculated Bouguer anomalies to the observed ones, the depth of the Moho discontinuity was determined to increase from 38–40 km at SYO to approximately 48–50 km beneath the Dome-F (Fig. 6-A). Over the wider area of the LHB, Nagao and Kaminuma (1984) used Bouguer anomalies to show a ~4 km variation in the Moho depth between the central Lützow-Holm Bay and Yamato Mountain area (Fig. 1), 300 km southwest from SYO. Since we cannot obtain any information about the crustal structure beneath the high latitude plateau southward from MZH, the combined information derived from the gravity anomalies and DSS surveys at LHB are essential for understanding the interior of the East Antarctic continent.

A similar analysis of the crustal structure using a density model was performed in the Ross ice stream, West Antarctica, using seismic refraction/wide-angle reflection and gravity profiles (Munson and Bentley, 1992). These results indicate that the crustal thickness is about 15 km in that area. The mean depth of the Moho discontinuity is about 35–45 km in East Antarctica and 30 km in West Antarctica, which includes the results from several DSS (Bentley, 1991; Kadmina et al., 1983; Reading, 2006; Ritzwoller et al., 2001). The crustal thickness determined using gravity data along 76°S latitude in East Antarctica was also previously obtained by Groushinsky and Sazhina (1982). In their study, the Moho discontinuity is shown to deepen from beneath the continental margins to an area 600–700 km inland, which agrees with our results.

Comparing our results with those from the JARE's South Pole Traverse in 1968 (Yanai and Kakinuma, 1971), the thickening crust toward Dome-F from SYO shows almost the same characteristics in both studies. Although the Moho depth in their study along the 75°S latitude was almost constant, at about 46–48 km, the bedrock elevation could not be obtained accurately, particularly for the inland plateau region along the South Pole traverse. The advantage of our study is that the continuous bedrock topography profiles were obtained with high resolution by the precise ground-based radio-echo soundings. Therefore, the results for the Bouguer anomalies and the obtained crustal structure are more accurate than those of the 1968 study. The Moho depth estimates we obtain beneath Dome-F are deeper than those of other studies in the marginal region of East Antarctica (Kanao et al., 2011; Reading, 2004; 2006; Yoshii et al., 2004).

4.3. Continuation to the GSM structure

The GSM are one of the most enigmatic tectonic features on Earth, but until recently only limited constraints were available on the deep structure beneath the mountain range (Cogley, 1984; DeConto and Pollard, 2003). Buried beneath the thick ice sheet, the mountains are characterized by peak elevations reaching ~3000 m above sea level (Bell et al., 2011). A crustal thickness of over 40 km beneath the GSM and the surrounding region was estimated using satellite and air-borne data (von Frese et al., 1999). As part of the AGAP/GAMSEIS project during IPY 2007–2008, more than 26 broadband seismometers were deployed over a large area in the middle of East Antarctica, extending from the crest of the GSM to Dome-F at the western end of the target area. Teleseismic events recorded by GAMSEIS provide new information on the fine crustal structure and help to constrain the origin of the GSM as well as the structure and evolution of the Pre-Cambrian craton in East Antarctica.

Using the GAMSEIS data, analysis of S-wave receiver functions and Rayleigh wave phase velocities has provided significant estimates of the crustal thickness beneath the GSM and the surrounding region (Hansen et al., 2010). The cratonic crust surrounding the GSM was found to be 40–45 km thick, consistent with average Pre-Cambrian crustal thickness observed globally (Mooney et al., 1998; Rudnick and Gao, 2003). The crustal thickness at the western end of the GAMSEIS *EW* profile (Fig. 8b, station GM05, crustal thickness of 47 km) is in good agreement with the crustal thickness beneath Dome-F (GM07 in Fig. 8-A) obtained from the JARE's gravity anomalies in this study. The eastern end of GAMSEIS *NS* profile (station N132-N124 in Fig. 8-B, crustal thickness of 42–45 km) also agrees well with the crustal thickness estimates obtained beneath the 'TransAntarctic Mountain SEISmic experiment (TAMSEIS)' profile from the Wilkes Land (Hansen et al., 2009; Lawrence et al., 2006).

Beneath the GSM, Hansen et al. (2010) observed thicker crust, with the Moho at a depth of 55–58 km. It was suggested that the thicker crust beneath the GSM provides isostatic support for the high mountain elevations (Fig. 8-a). Three-dimensional surface wave tomography models, also developed using the GAMSEIS data, also indicate a thickened crust around the GSM (An et al., 2010). Additionally, high velocity anomalies beneath the GSM determined by Rayleigh wave phase analyses (Heeszel et al., 2010) support the idea of crustal root beneath this region and do not show any

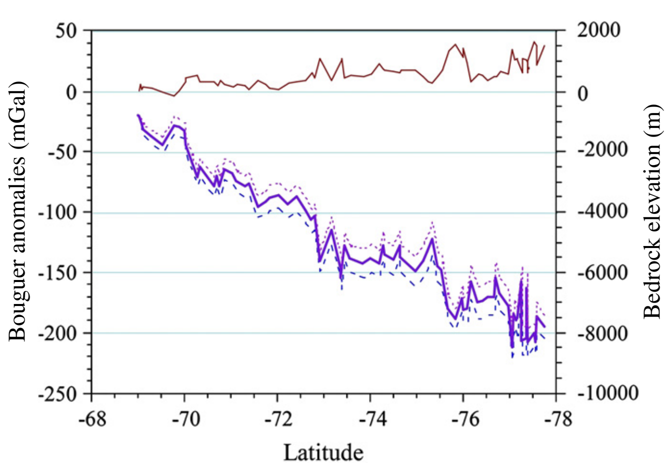

Fig. 7. Observed simple Bouguer anomalies for different reduction densities. The three violet-colored traces correspond to densities of 2.60 g/cm^3 (upper), 2.70 g/cm^3 (middle), and 2.80 g/cm^3 (lower), respectively. The bedrock elevation is also shown (brown line) for comparison.

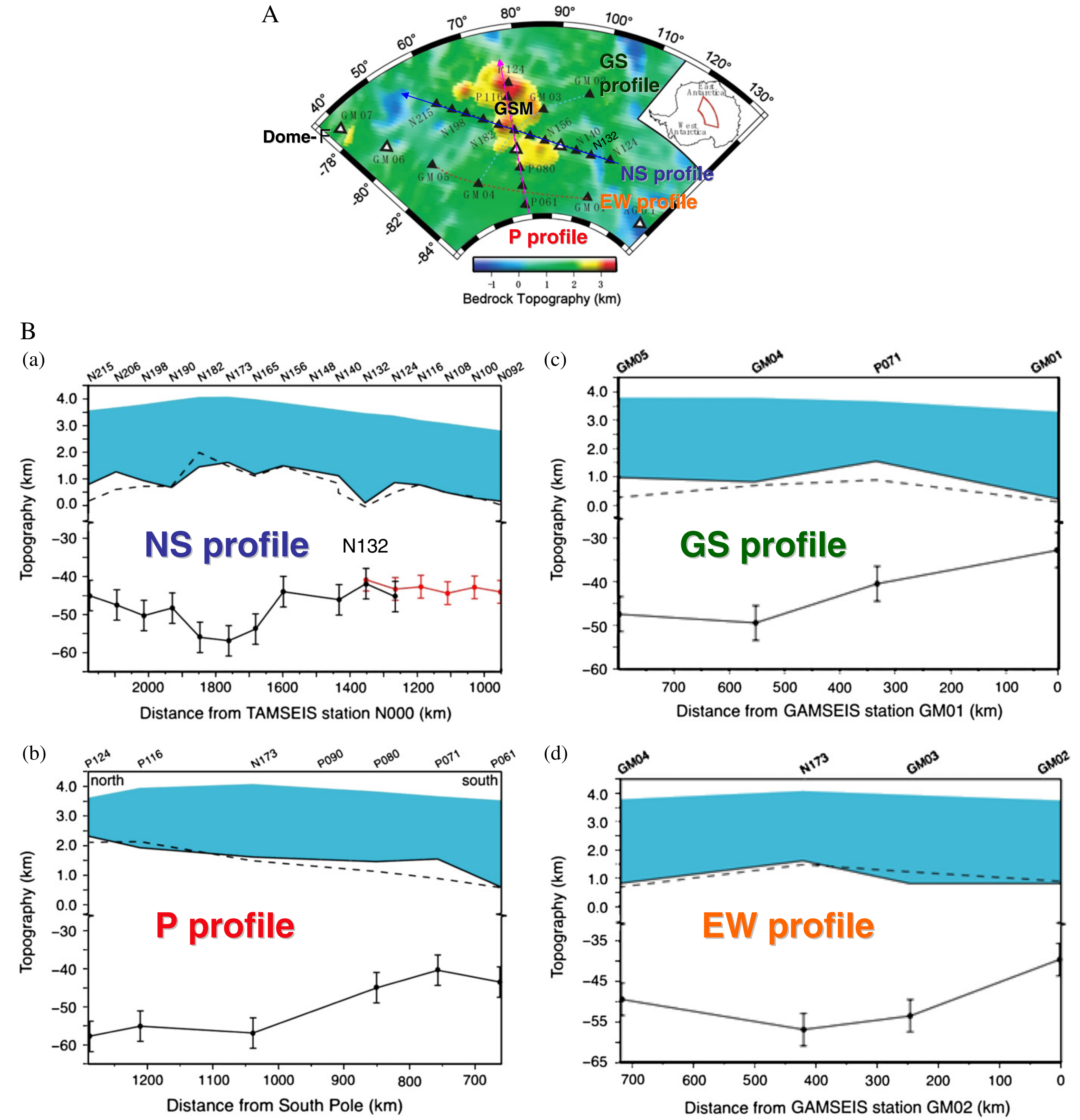

Fig. 8. *A*. Distribution map of the GAMSEIS seismic stations with four different profiles across the GSM indicated (NS, GS, P, and EW profiles). Background colors indicate bedrock topography from BEDMAP (Lythe et al., 2001). *B*. Crustal structure (Moho topography) along the four different profiles shown in Fig. 8-a, determined by S-wave receiver function analysis. In all plots, the light blue area and the solid black line indicate ice thickness and surface topography, respectively, which were determined using teleseismic P-wave receiver function stacks. For comparison, the surface topography from BEDMAP (Lythe et al., 2001) is shown by the dashed black line. The Moho depth estimates from Hansen et al. (2010) are shown by black dots with ± 4 km error bars. Along the NS profile, the estimated Moho discontinuity was extended further eastward, connecting to the results from TAMSEIS (Hansen et al., 2009; shown by red dots).
a, b: modified from Hansen et al. (2010).

evidence of a mantle plume (seismic low velocities) in the upper mantle beneath GSM. The thicker crust beneath the GSM might reflect ancient continental tectonic features associated with Proterozoic–Paleozoic orogenic events in East Antarctica. Pan-African or Grenville orogenic/metamorphic events may be the most plausible candidates to explain the formation of the GSM (Fitzsimons, 2000; 2003; Liu et al., 2006).

In order to conduct more detailed investigations of the crustal structure beneath wide areas of East Antarctica, denser geophysical surveys with a significant number of onshore stations are required, particularly in the LHB, Dome-F, and GSM regions. The IPY 2007–2008 provided a good opportunity to conduct such studies. The compilation of the gravity and geomagnetic data help constrain the formation process of the Gondwana continent in Earth's

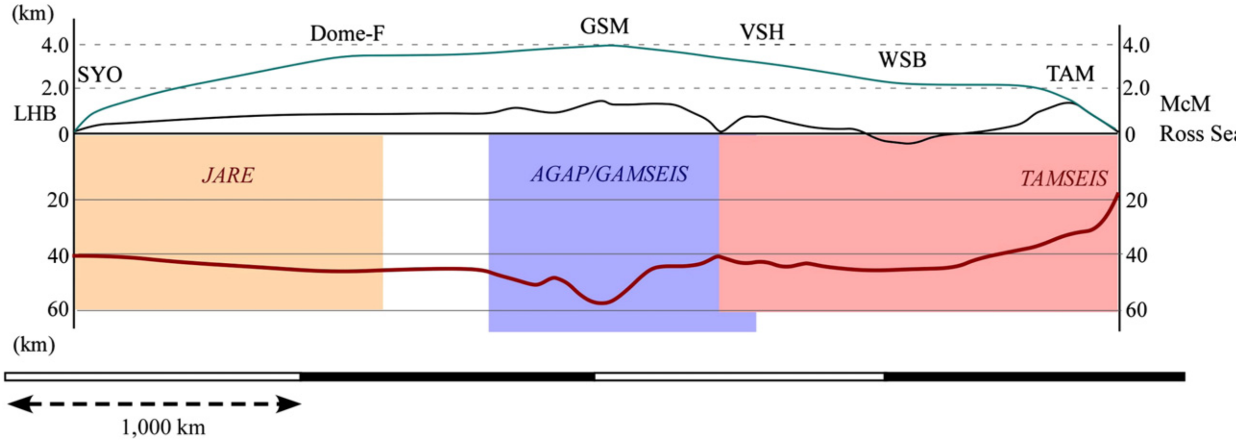

Fig. 9. An illustration of the combined crustal cross-section across East Antarctica from LHB (SYO) to Dome-F, the GSM, and the TAM (McM). The Moho topography (thick solid red line), bedrock topography (black solid line), and the surface elevation of the ice sheet (light blue solid line) are shown. Abbreviations of the region names as the same as in Fig. 1. The JARE structure is obtained from the current study while the GAMSEIS and TAMSEIS structure are from Hansen et al. (2010; NS profile in Fig. 8-b), and Hansen et al. (2009), respectively.

history. A combination of the seismic refraction and reflection experiments, the broadband seismic deployments, the geo-potential air-borne surveys, and the precise radio-echo soundings would provide the most useful information for understanding the interior of the Antarctic continent and the boundary conditions controlling the overlying ice sheet. Recent progress of ice-core drilling technology also makes it possible to measure the actual ice thickness as well as to investigate the actual geothermal conditions at the bottom of ice sheet. Determination of the bedrock elevations could constrain the uppermost crustal structure and should be directed by strong cooperation between glaciology and geophysics involving Antarctic research.

5. Conclusion

Onshore gravity measurements were conducted by three JAREs (1992, 1997, and 1998) over inland traverse routes from LHB to Dome-F in East Antarctica. Free-air and Bouguer anomalies based on gravity disturbances along these routes were obtained using both surface and bedrock elevation measurements from radio-echo soundings. The crustal density structure between the LHB and the inland plateau was derived from the P-wave velocity model determined by seismic refraction surveys. A decrease in the Bouguer anomalies down to −200 mGal from LHB toward the inland plateau indicates that the crust thickens from 38–40 km at LHB to 48–50 km beneath Dome-F. The GAMSEIS project deployed a significant number of broadband stations over the large highland on the ice sheet from the crest of the GSM to the vicinity of Dome-F during the IPY. The acquired seismic data helped to refine crustal structure estimates beneath the GSM and to constrain the origin of the mountain range as well as the broad structure and the evolution of the East Antarctic craton. S-wave receiver functions and Rayleigh phase velocities determined using the GAMSEIS data (Hansen et al., 2010) indicate that the cratonic crust surrounding the GSM is 40–47 km thick. These values agree with those from the JARE's gravity surveys around Dome-F and are consistent with average Pre-Cambrian crustal thicknesses. Beneath the GSM, Hansen et al. (2010) have found that the crust thickens to 55–58 km, and it was interpreted that this thicker crust provides isostatic support for the high mountain elevations. Thicker crust beneath the GSM might reflect the ancient continental formation history associated with Proterozoic or Paleozoic orogenic events. Accordingly, a long-distance crustal model from LHB to Dome-F and to the GSM were combined using both seismic and gravity studies. The crustal cross-section extends over 3000 km across the central part of East Antarctica (Fig. 9) and could provide significant information on the tectonic evolution of the Gondwana continent.

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