J Geod (2008) 82:403–414
DOI 10.1007/s00190-007-0189-2

ORIGINAL ARTICLE

Regional geoid determination in Antarctica utilizing airborne gravity and topography data

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Received: 9 January 2007 / Accepted: 25 September 2007 / Published online: 3 November 2007

Abstract Antarctica is the only continent that suffers major gaps in terrestrial gravity data coverage. To overcome this problem and to close these gaps as well as to densify the global satellite gravity field solutions, the International Association of Geodesy (IAG) Commission Project 2.4 "Antarctic Geoid" was set into action. This paper reviews the current situation concerning the gravity field in Antarctica. It is shown that airborne geophysical surveys are the most promising tools to gain new gravity data in Antarctica. In this context, a number of projects to be carried out during the International Polar Year 2007/2008 will contribute to this goal. To demonstrate the feasibility of the regional geoid improvement in Antarctica, we present a case study using gravity and topography data of the southern Prince Charles Mountains, East Antarctica. During the processing, the remove–compute–restore (RCR) technique and least-squares collocation (LSC) were applied. Adding signal parts of up to 6 m to the global gravity field model that was used as a basis, the calculated regional quasigeoid reveals the dominant features of bedrock topography in that region, namely the graben structure of the Lambert glacier system. The accuracy of the improved regional quasigeoid is estimated to be at the level of 15 cm.

Keywords Antarctica · Regional geoid modelling · Remove–compute–restore (RCR) technique · Aerogravimetry

1 Introduction

The determination of the gravity field of the Earth and its temporal changes is one of the central tasks of geodesy. In the framework of a "Global Geodetic Observation System" (GGOS) it aims at reaching a relative accuracy of 10^{-9} for the (static) gravity field (Drewes 2007). In recent years, remarkable progress has been made in fulfilling this goal due to the availability of the new data from the satellite missions "Challenging Minisatellite Payload" (CHAMP, launched 2000) (Reigber et al. 2002) and "Gravity Recovery and Climate Experiment" (GRACE, launched 2002) (Tapley et al. 2004). A third dedicated gravity mission, "Gravity Field and Steady-State Ocean Circulation Explorer" (GOCE), will be launched in 2008[1] (Rummel et al. 2002; Drinkwater et al. 2003).

Nevertheless, we face the polar gap problem when dealing with satellite missions in Antarctica (Baur and Sneeuw 2006; Sneeuw and van Gelderen 1997). Even more, Antarctica remains the continent with the largest gaps of terrestrial gravity data coverage. To overcome these problems, the Antarctic Geoid Project (AntGP) of the International Association of Geodesy (IAG) was set into action. At first, this paper briefly reviews the situation of the observation and determination of the gravity field in Antarctica. It will be shown that airborne gravimetry provides the most promising tool to decisively

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[1] www.esa.int/esaLP/LPgoce.html, last accessed 23 October 2007.

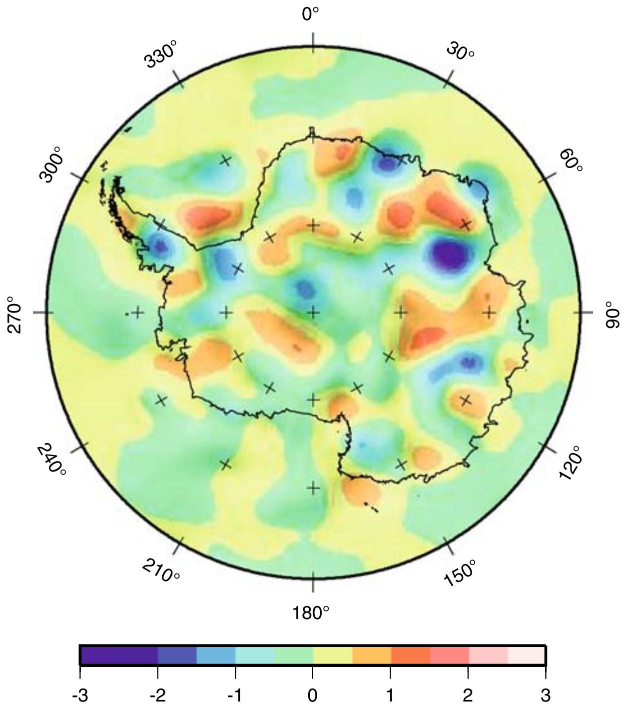

Fig. 1 Geoid height difference between EIGEN-GL04C and EGM96 for Antarctica up to degree and order 360 (units: m)

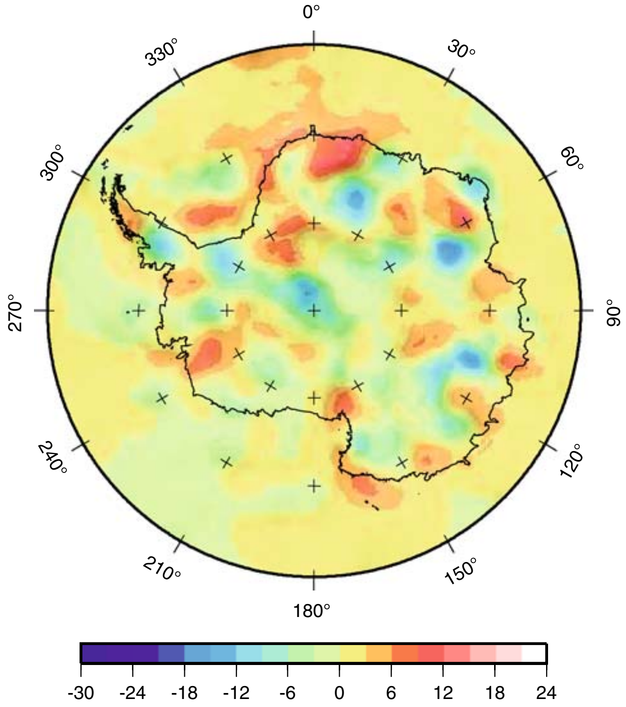

Fig. 2 Free-air anomaly difference between EIGEN-GL04C and EGM96 for Antarctica up to degree and order 360 (units: mGal)

extend the gravity data coverage in Antarctica. Using consistent data from a joint German-Australian expedition to East Antarctica we will present a case study of regional quasigeoid modelling in Antarctica.

2 Determination of the gravity field in Antarctica

2.1 The global gravity field as seen in Antarctica

Using CHAMP and GRACE data in combination with terrestrial gravity data, new high-resolution global models of the gravity field of the Earth have been computed. Latest examples of these combination models are EIGEN-CG03C and EIGEN-GL04C (Förste et al. 2005, 2006) and GGM02C (Tapley et al. 2005). Comparing EIGEN-GL04C to the pre-CHAMP/GRACE model EGM96 (Lemoine et al. 1998) the current level of knowledge of the Antarctic gravity field can be assessed by plotting the differences of geoid heights and of gravity anomalies, respectively (Figs. 1, 2).

The difference of both models reaches a level of several meters for quasigeoid heights and a range of about 50 mGal (and peak values of up to 100 mGal) for gravity anomalies. These differences mainly illustrate the lack of observed terrestrial data that have to be included in the combined global gravity field models. Also, geophysically extrapolated gravity anomalies represent the actual gravity field in Antarctica only insufficiently since, especially, models on bedrock topography (Lythe et al. 2000) lack resolution and/or accuracy. Nevertheless, these extrapolated gravity anomalies are inevitable to provide the globally complete data coverage needed to determine the global combination models. Hence, what can be seen when comparing the aforementioned models is the poor coverage of terrestrial gravity data for the Antarctic continent. Only for a few smaller regions ground-based or airborne measured gravity was included into the combination, originating from the data holdings of the National Imagery and Mapping Agency [NIMA, now National Geospatial-Intelligence Agency (NGA)] and additionally from the West Antarctic Ice Sheet campaign (Bell et al. 1999) and the Weddel Sea region (Studinger 1998). Over the oceans, the situation is much better due to the inclusion of satellite altimetry, see e.g. (Schöne 1997).

Furthermore, the new satellite data show gaps in the polar regions. On one hand, this is caused by the deviation of the orbit inclination from 90°. The polar gap, covering an area with a diameter of about 700 km for CHAMP (orbit inclination 87°), 100 km for GRACE (89.5°) and 1,300 km for GOCE (95.5°), influences the determination of the zonal harmonics and deteriorates long-wave length information for regional improvements in that region, see e.g. (Rudolph et al. 2002; Sneeuw and van Gelderen 1997). On the other hand, the resolution is limited to about 300 km (GRACE) and 200 km (GOCE) due to the standard analysis procedures that use spherical harmonics. Alternative approaches can be applied to gain a higher resolution from satellite data only, see e.g. (Mayer-Gürr et al. 2006).

2.2 IAG Commission Project 2.4 “Antarctic Geoid”

At its General Assembly in Birmingham, 1999, the IAG recognized the need for terrestrial and airborne gravity measurements in the polar regions in order to cover the polar gaps and to improve the geoid in these regions, and adopted a respective resolution. For the Arctic, the IAG SSG 3.178 “Arctic Gravity Project” undertook a major international collaboration effort to close the polar gap in the northern hemisphere (Kenyon and Forsberg 2002). Grids of free-air anomalies and geoid undulations north of 64° have been released.[2]

In order to enforce international cooperation and to improve the situation of the terrestrial gravity coverage in Antarctica, AntGP was adopted as the IAG Commission Project 2.4 “Antarctic Geoid” within Commission 2 “Gravity Field”, Sub-Commission 2.4 “Regional Geoid Determination” at the IAG General Assembly in Sapporo, 2003. The status and progress of AntGP were reported several times by its chair (Scheinert 2005a; Scheinert et al. 2006a, b). A concise overview was given in (Scheinert 2005b) and a website[3] has been set up.

The main goal of AntGP is to compile a completed gravity data set for Antarctica by collecting already existing gravity data as well as by performing new surveys. The rationale of AntGP is closely linked to similar initiatives, especially to the Scientific Committee on Antarctic Research (SCAR), Standing Scientific Group on Geosciences (SSG-GS) and Group of Specialists on “Geodetic Infrastructure in Antarctica” (GIANT). Within the GIANT programme, which was newly adopted at the SCAR XXIX Conference in Hobart, July 2006, the project 3 “Physical Geodesy” is chaired by the first author and co-chaired by A. Capra (Italy).

Moreover, the “International Polar Year” (IPY), which lasts from 1 March 2007 to 28 February 2009 and thus covers two Arctic and two Antarctic seasons, offers promising chances to take a great step forward in closing the polar gap in gravity through intensified international cooperation efforts. Aerogeophysical methods provide the most powerful tool to carry out gravity surveys in Antarctica, since large areas can be surveyed in an economic way. Different IPY projects are already under detailed discussion and hence we can expect plenty of new gravity data in the next few years [e.g. IPY project 67 “Origin, evolution and setting of the Gamburtsev subglacial highlands (AGAP)” and IPY project 97 “Investigating the Cryospheric Evolution of the Central Antarctic Plate (ICECAP)”[4]].

Besides the inherent task of geodesy to determine the (external) gravity field and the geoid, an Antarctic terrestrial gravity database should be established for densification and validation purposes. A regional, hence high-resolution geoid model in Antarctica can be used for further studies in geodesy, geophysics, glaciology and oceanography. For example, applying a regional geoid model for the computation of free-board heights of ice-shelves from ellipsoidal heights, an ice-thickness model can be inferred using equations of the hydrostatic balance (Horwath et al. 2006).

2.3 Regional gravity surveys on the Antarctic continent

Ship-based gravity measurements provide a major source of data for the Antarctic ocean region and thus complement and densify gravity information yielded by satellite altimetry (Schöne 1997). In this paper, we will concentrate on the situation on the Antarctic continent.

Pointwise gravity measurements have been reported, e.g., for North Victoria Land (Reitmayr 1997, 2003) and for central Dronning Maud Land (Reitmayr 2005; Fritzsche 2005). Gravity observations were made with the help of a helicopter or going by snow vehicles along traverses. Using this information, Korth et al. (1998) calculated an improved regional geoid model for the region of the Schirmacher Oasis by means of a point-mass model (Barthelmes 1986; Barthelmes and Dietrich 1991). Other geoid computations of regional extent were reported by Coren et al. (1997), Karner et al. (2005) and Coren et al. (2004) for the Ross Sea region and for North Victoria Land. Further point gravity observations were carried out along snow-vehicle traverses dating back to the International Geophysical Year 1956/1957 and by pointwise measurements on the Antarctic Peninsula (Ferraccioli et al. 2006), the Weddell Sea region (Aleshkova et al. 2000) and some other Antarctic regions like Low Dome or the Ross Sea embayment.

Due to the hostile environment and vast extension of Antarctica, ground-based gravity surveys are restricted to regions near the coast and are of very limited coverage. Long traverses provide a gravity tie between different regions but do not cover a larger area. Nevertheless, terrestrial observations may help to identify problems in the gravity datum, e.g. for airborne surveys, if they are connected to absolute gravity stations, which have been established in research stations at the Antarctic coast (Mäkinen et al. 2007).

A much more powerful tool to measure gravity at larger areas is provided by airborne geophysics. Different observation techniques are usually combined aboard an aircraft. These comprise a gravimeter (mostly an adapted sea gravimeter like LaCoste-Romberg S-type or Bell BGM-3); Global Navigation Satellite System (GNSS) and Inertial Navigation System (INS) for a precise determination of the flight trajectory and of the accelerations and attitude angles needed for the correction and reduction of the raw gravity observations; ground penetrating radar for the determination of ice thickness and thus—by a combination with GNSS—of ice

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[2] earth-info.nga.mil/GandG/wgs84/agp.

[3] www.tu-dresden.de/ipg/antgp.

[4] IPY project database at www.ipy.org.

surface heights and bedrock topography. A laser altimeter (or scanner) and magnetic field sensors may complement the geophysical instrumentation aboard a research aircraft.

For the Arctic, the power of airborne gravimetry and geophysics has been impressively demonstrated by the Greenland survey in the early 1990s (Forsberg and Brozena 1993; Forsberg 1993) and by the Arctic Gravity Project in the subsequent years (Kenyon and Forsberg 2002). In the Antarctic, logistic efforts are much higher. Despite sparse landing strips on gravel, airfields have to be maintained on ice (or compacted firn), and for remote operations extensive field camps have to be erected for the time of the campaign.

Compared to the total area of the Antarctic continent, only a small part could be surveyed by airborne gravimetry, although the individual campaigns often comprise several thousands to tens of thousands of kilometers of flight tracks. Airborne gravimetric surveys have been carried out mostly by larger agencies, like US polar program (NSF), Russian institutions (RAE, VNIIOkeangeologia/PMGRE), British Antarctic Survey (BAS) or Alfred Wegener Institute for Polar and Marine Research (AWI) in Germany.

They include e.g. the Lake Vostok region (Studinger et al. 2003; Holt et al. 2006), a transect of the Transantarctic Mountains (Studinger et al. 2004), the West Antarctic Ice Sheet (Bell et al. 1999; Studinger et al. 2002), the Weddell Sea region (Aleshkova et al. 2000), the Antarctic Peninsula and adjacent regions (Jones et al. 2002), (Ferraccioli et al. 2006), western and central Dronning Maud Land (Nixdorf et al. (2004)) and the Jutulstraumen in western Dronning Maud Land (Ferraccioli et al. 2005). A high-resolution airborne gravity and geophysics survey was conducted in the Prince Charles Mountains area, East Antarctica (Damaske and McLean 2005) during the “Prince Charles Mountains Expedition of Germany and Australia” (PCMEGA) 2002/2003. This PCMEGA survey provides an excellent data set for a case study of a regional geoid improvement, which shall be discussed in the following section.

Summarizing, one has to state that the huge effort of any airborne survey in Antarctica can be realized only in close interdisciplinary cooperation. Utilizing gravity data, the geophysical and the geodetic applications are quite complementary, striving for the inner structure of the Earth and of the ice sheet on one hand, and for determining the outer gravity field on the other hand.

3 Regional geoid determination in the Prince Charles Mountains area: a case study

3.1 The PCMEGA airborne survey

During the austral summer 2002/2003 Germany and Australia carried out a joint interdisciplinary project in the region of the Lambert Glacier and Prince Charles Mountains, East Antarctica. This project comprised geophysical, geological, geodetic and glaciological sub-programs and was called “Prince Charles Mountains Expedition of Germany and Australia” (PCMEGA).

One of the major goals of PCMEGA was to gain a better understanding of crustal structure and tectonic regime in that region. The Amery Ice Shelf—Lambert Glacier region with its north–northeast trending graben structure is believed to give evidence for an Early Cambrian orogenesis and thus showing fundamental structures of Gondwana (Boger and Wilson 2005). An extensive airborne geophysical survey was part of PCMEGA, covering the southern Prince Charles Mountains and further south, from 72°15′S to 77°30′S latitude and from 62°E to 72°E longitude. For details on this survey see Damaske and McLean (2005).

The instrumentation was installed aboard a De Havilland DHC-6 Twin Otter contracted by the Australian Antarctic Division (AAD) and comprised a LaCoste-Romberg S gravity meter, a Scintrex CS-2 magnetometer, both provided by Fugro (Baron-Hay et al. 2003), and an ice penetrating radar, provided by the Federal Institute for Geosciences and Resources (BGR) (Damaske and McLean 2005). BGR was also responsible for the planning and execution of the entire airborne survey. The layout of the survey can be seen in Fig. 3. Since the ice surface heights increase southwards, and the gravity meter has to be flown at constant altitude, the gravity data were collected at three flight levels (2,160, 2,760 and 3,360 m), which caused some data gaps (Fig. 3). The flight profiles went approximately in north–south direction with a line spacing of 5 km. Additional control lines were flown perpendicular to this main flight direction at a spacing of 25 km.

3.2 Available data sets

Processing of the raw gravity data, including all corrections and reductions, was carried out by Fugro (Baron-Hay et al. 2003). Especially, the line-oriented gravity data were adjusted using a proprietary software developed by Fugro. The gridding was carried out applying a “moving weighted average Gaussian distribution gridding routine (MAVG)” (Baron-Hay et al. 2003), thus also closing remaining data gaps. Finally, Fugro supplied gridded free-air gravity anomalies with a grid cell size of 1,000 m in three subsets according to the three different flight levels. It is this gridded dataset which served as original input data for our regional geoid improvment. McLean and Reitmayr (2005) performed a first detailed analysis of the gravity data, based on complete Bouguer anomalies and an isostatic modelling analogous to the Airy model, both computed by a 2D fast fourier transform (FFT).

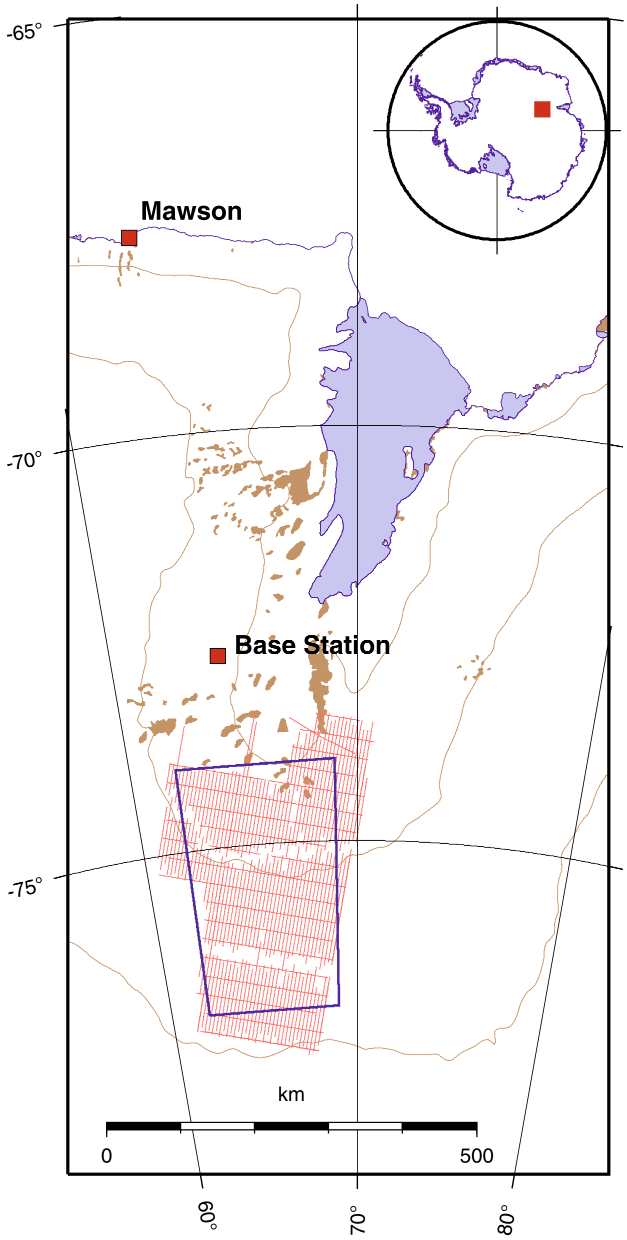

Fig. 3 Overview of the area of investigation. The Amery Ice Shelf is shown in the North, the Lambert Glacier extends southwards. Mountainous regions and outcrops are visible as smaller dark patches. Height contours are given at an interval of 1000 m. The Australian Antarctic station Mawson served as a logistic base to supply the PCMEGA base station, which was erected at Mt. Cresswell in the southern Prince Charles Mountains. All flight lines are plotted, which included gravity observations. Locations where no data were aquired are not plotted, so that the subdivision into the three different flight levels (2,160 m, 2,760 m and 3,360 m) is visible. The area of computation is outlined by a box. (Map source: Antarctic Digital Database (ADD) v4.0 (ADD Consortium 2000), polar stereographic projection.)

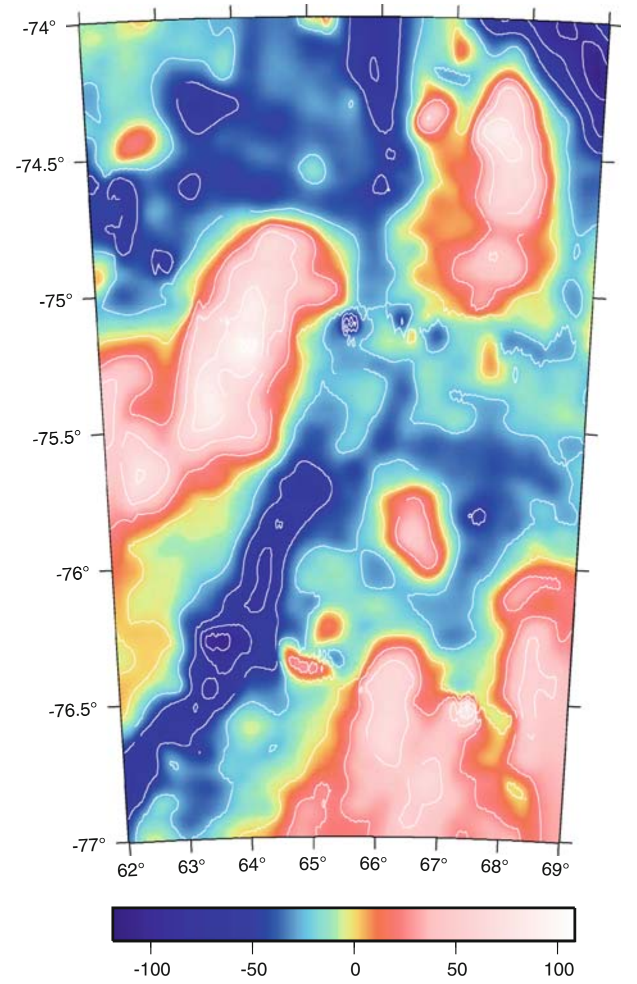

Fig. 4 Gravity disturbances at flight levels 2,160, 2,760 and 3,360 m (from north to south) as supplied by Fugro (Baron-Hay et al. 2003) (units: mGal; polar stereographic projection)

For our case study, a regular subgrid of the original gridded gravity data between 74°S and 77°S and 62°E and 69°E was chosen in such a way that the area of computation was completely covered by PCMEGA flight lines and that no data extrapolation was necessary. Differently to the Fugro report (Baron-Hay et al. 2003), the provided gravity data are gravity disturbances rather than gravity anomalies, cf. (Hackney and Featherstone 2003). While processing the raw gravity data, the normal gravity reduction was computed using the Gravity Formula 1967 (Baron-Hay et al. 2003). In order to be consistent with the geodetic datum used for the aircraft positioning, an additional correction was applied to refer to the Gravity Formula 1980 (Moritz 1984). This correction term (Fuchs and Soffel 1984) reaches an almost constant value of 0.9 mGal for the entire area of computation. The grid of the gravity disturbances at flight level as shown in Fig. 4 resembles the graben structure of the Lambert rift system.

From the simultaneous radar observations, maps of the ice thickness and—combined with a precise GPS positioning of the aircraft trajectory—of ice surface height and bedrock topography were obtained (Damm 2007). In order to apply the RCR technique the effect of topography in terms

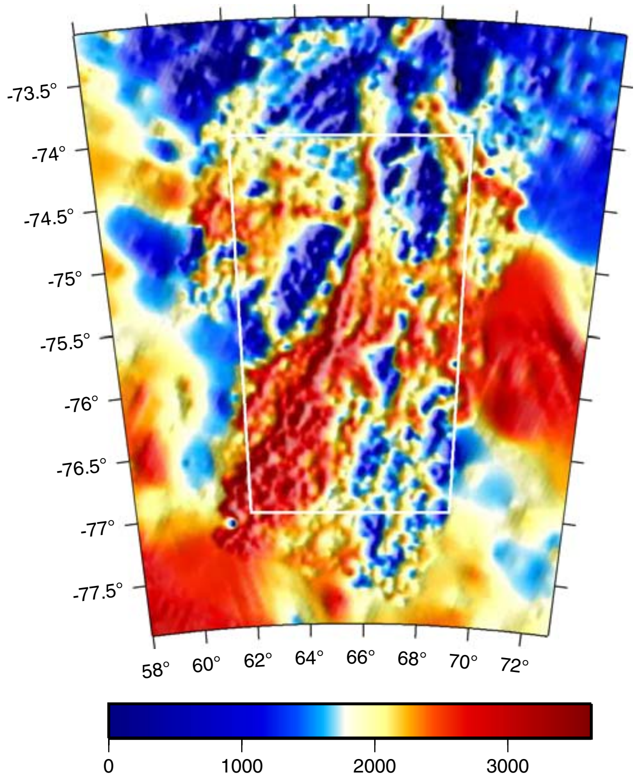

Fig. 5 Ice thickness (units: m). The PCMEGA dataset has been patched by data of the BEDMAP project (Lythe et al. 2000) (polar stereographic projection)

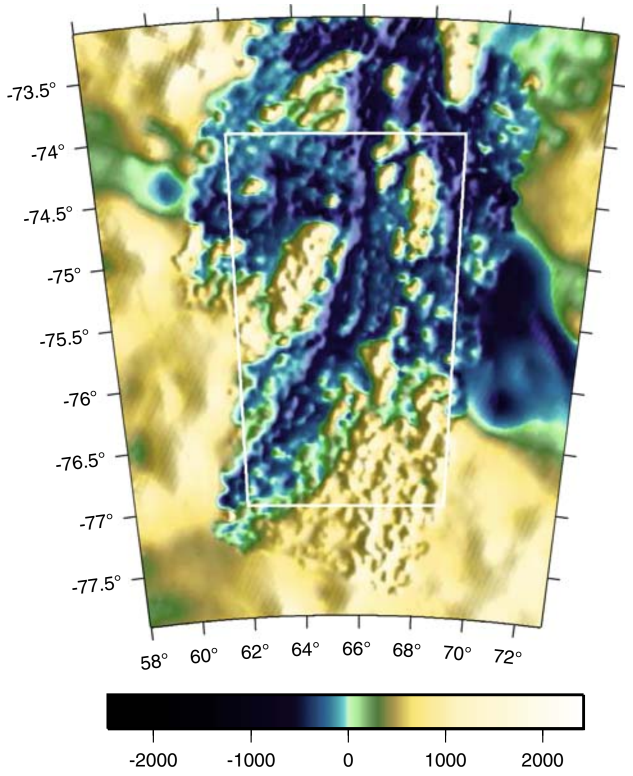

Fig. 6 Bedrock topography (units: m). The PCMEGA dataset has been patched by data of the BEDMAP project (Lythe et al. 2000) (polar stereographic projection)

of gravity and of geoid have to be computed. In order to avoid edge effects resulting from missing topography data at the edge of the computation area, a border region of about 100 km was patched by data from the BEDMAP project (Lythe et al. 2000). Although the BEDMAP data are of lower resolution (and accuracy), they fulfill the requirements, considering that the effect of topographic masses attenuates with increasing distance.

The resulting data sets are shown in Fig. 5 (ice thickness) and Fig. 6 (bedrock topography). In these figures, the area of computation is marked by a white box. Comparing the gravity disturbances (Fig. 4) with the bedrock topography data (Fig. 6), a strong correlation can be seen (correlation coefficient 0.78). The PCMEGA bedrock topography reveals the above mentioned, strongly pronounced rift structure and a couple of rocky outcrops. The thickness of the ice layer reaches values of up to 3,500 m, nevertheless some ice-free areas still exist.

3.3 Processing steps for the regional geoid improvement

For the calculation of the regional geoid the RCR technique was applied, discussed in detail e.g. by Forsberg and Tscherning (1997) and Sjöberg (2005). In the remove step, a long-wavelength part (predicted by a global gravity field model) and a short-wavelength part (predicted by topography) are removed from the original gravity data. In the compute step, the obtained band-pass filtered gravity disturbances are transformed into quasigeoid heights, using least-squares collocation in this study. Least-squares collocation offers the advantage of providing error estimates for the resulting geoid. After having carried out the compute step, the long-wavelength part and the short-wavelength part are restored to the quasigeoid. In the following, we will concentrate on the practical application of the RCR technique considering the special conditions of Antarctica. For the computations, we could make use of the program package GRAVSOFT (Forsberg et al. 2003; Tscherning 1974), which offers a variety of programs for the geodetic gravity field modelling. Statistical parameters for the individual processing steps are given in Table 1.

First, the boundary surface has to be defined, which serves as the surface where the data should be given according to the concept of the geodetic boundary value problem in Molodensky's formulation of the quasigeoid determination. In our case, dealing with the Antarctic situation, the bedrock topography can be chosen as this boundary surface. There is only one exception: in case the ellipsoidal height of the bedrock topography is negative, the ellipsoidal height of the boundary surface is set to zero. To carry out the downward

Table 1 Statistical parameters of the different data sets described in Sects. 3.2–3.4

Processing step	Figure	Minimum (mGal)	Maximum (mGal)	Mean (mGal)	St. deviation (mGal)
Gravity disturbances at flight level 2,160 m		−127.12	92.92	−13.22	36.31
Gravity disturbances at flight level 2,760 m	(4)	−120.78	110.85	−6.69	33.64
Gravity disturbances at flightlevel 3,360 m		−83.69	114.23	0.82	35.95
Downward continued gravity disturbances	(7)	−191.00	268.39	−9.90	40.54
Band-pass filtered gravity disturbances	(8)	−236.68	184.46	−58.99	34.10
Processing step	**Figure**	**Minimum (m)**	**Maximum (m)**	**Mean (m)**	**St. deviation (m)**
Residual height anomalies	(10)	−5.65	0.91	−2.61	1.54
Improved regional quasigeoid	(11)	3.22	13.06	6.65	2.02

continuation of the gravity disturbances from the flight level to the boundary surface, a 2D FFT was applied, which uses the second partial derivative of the disturbing potential T_{zz} (where z denotes the vertical).

Furthermore, since the area of computation is situated completely in East Antarctica, which hosts the Earth's largest ice sheet, one additionally has to deal with the ice layer, which reaches a thickness of more than 3,000 m in the area. To take the gravitational effect of the ice layer into account, a reduction was applied which equals a complete Bouguer reduction using the density value of ice (917 kg/m^3). The height differences or thickness values which are introduced into the computation of this complete Bouguer reduction had to be chosen according to the definition of the boundary surface. This first processing step, combining downward continuation and removal of the gravitational effect of the ice layer, resulted in gravity disturbances given at the boundary surface as shown in Fig. 7. The graben structure of the Lambert rift system is even more pronounced compared to the data at flight level, which is also confirmed by the range of variation (Table 1).

Secondly, the remove step itself was carried out. The goal of this remove step is to yield residual gravity values, which reduces possible numerical errors introduced by the following Stokes integration (Sjöberg 2005). Furthermore, it can provide an idea to what extent the global gravity field and the regional topography, respectively, dominate the gravity signal. For the removal of the long-wavelength part, EIGEN-CG03C, one of the latest global gravity models was used (Förste et al. 2005). According to the concise examination of the global, models described in Sect. 2.1 it was decided to cut off the model at degree and order 120. This upper limit of the spherical frequency was chosen because in East Antarctica the global model almost does not contain any observed terrestrial data (Förste et al. 2005). Moreover, GRACE data as the most relevant satellite input to the EIGEN-CG models is significant up to degree 120 (Tapley et al. 2004). This is especially true for polar regions where due to the denser orbit sampling GRACE solutions are more accurate than on global average.

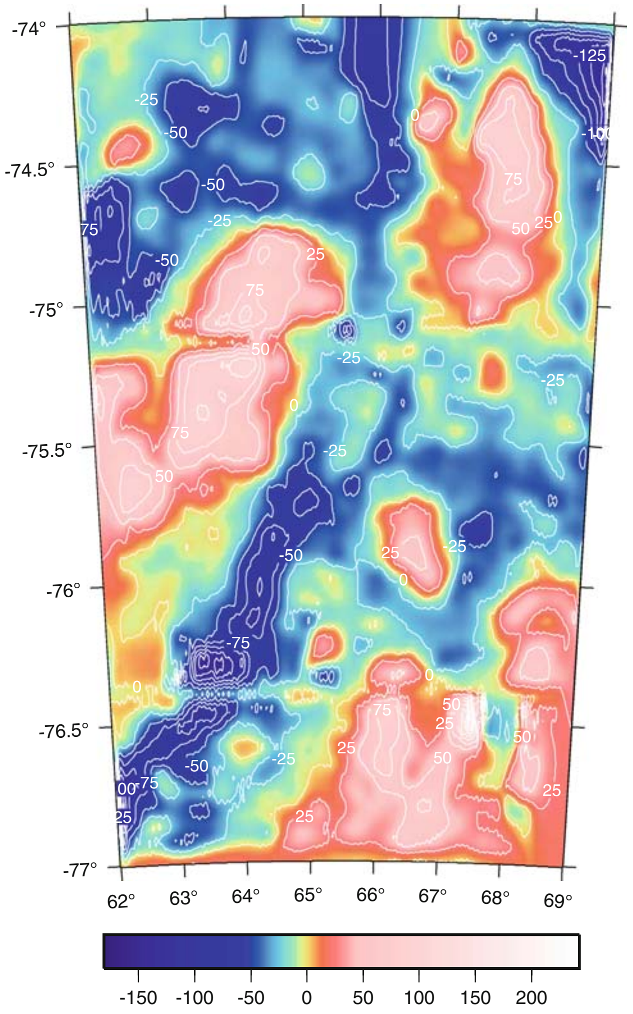

Fig. 7 Downward continued gravity disturbances at the boundary surface (units: mGal; polar stereographic projection)

The gravity disturbances calculated from the EIGEN-CG03C model reach a minimum of −30 mGal in the northern part and a maximum of up to 20 mGal in the southern part of the area of computation. For the removal of the short-wavelength part, the bedrock topography model was used (Fig. 6). In order to avoid a double consideration of signal parts at wavelengths already removed by application of the global model, the topographic effect has to be high-pass filtered at a cut-off wavelength of 300 km, which corresponds to the upper spherical harmonic degree 120 used with the global model. To realize this high-pass filter, the residual terrain model method (RTM) was applied (Forsberg 1984). A reference topography surface was obtained from the original bedrock topography by low-pass filtering (realized by a moving average operator with a length of 300 km).

To calculate the short-wavelength gravitational effect to be removed from the data, the residual terrain height was used originating from the difference between the actual bedrock topography and the reference topography surface. The cases when the ellipsoidal heights of the boundary surface were set to zero—hence there still exists an ice layer of a certain thickness (i.e. the actual ellipsoidal height of bedrock topography is negative)—were treated analogously to the case of ocean areas taking the bathymetry into account. Instead the density of sea water the density of ice was used.

Subtracting the effects of the global model and of bedrock topography according to the RTM calculation, residual or band-pass filtered gravity disturbances were obtained. This residual signal is shown in Fig. 8. It is slightly smoother than the original (Fig. 4) or the downward-continued signal (Fig. 7), which is also confirmed when calculating the correlation coefficient with bedrock topography to be now -0.32 (compared to 0.78 for the original signal, see Sect. 3.2).

Thirdly, the compute step could be carried out. The Stokes integration was realized by LSC as elaborated by Tscherning and Rapp (1974) and Knudsen (1987), and implemented in the GRAVSOFT package (Tscherning 1974). In preparation of this collocation step, an empirical covariance function was estimated from the residual gravity data, which then was fitted by an analytical covariance function using a Tscherning-Rapp degree-variance model (Tscherning and Rapp 1974; Knudsen 1987). The analytical covariance function was chosen in such a way that it optimally predicts the decay behaviour of the empirical covariance function (Fig. 9). The predicted analytical covariance function was introduced into the LSC. Additionally, the a priori standard deviation of the gravity data was set to 5 mGal, which represents a more conservative estimate of the error measure. The collocation yielded residual or band-pass filtered quasigeoid heights which are plotted in Fig. 10. This residual height anomaly signal is further smoothed in comparison to the residual gravity signal and reaches minimum values of about -5 m (which correspond to the graben structure to be seen in the bedrock topography, Fig. 6) and maximum values of about $+1$ m in the northern part of the area of computation.

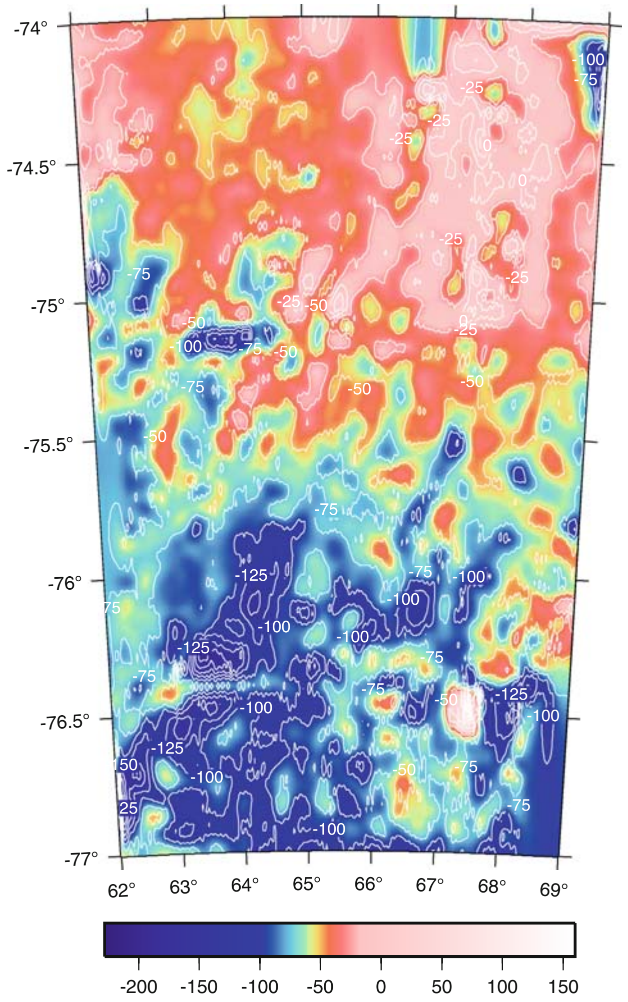

Fig. 8 Band-pass filtered gravity disturbances at the boundary surface (units: mGal; polar stereographic projection)

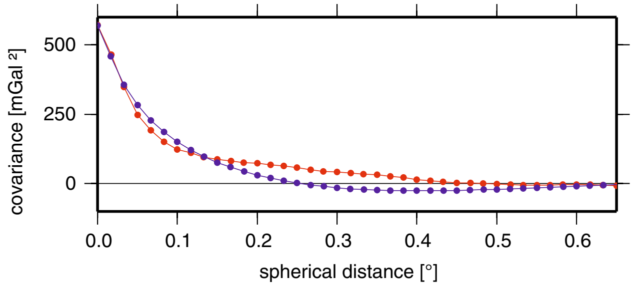

Fig. 9 Empirical (*upper curve*) and analytical (*lower curve*, crossing the abscissa at 0.25°) covariance functions of the band-pass filtered gravity disturbances

3.4 Final regional quasigeoid

Having completed the remove and computation steps, the restore step was carried out. The quasigeoid effects coming from the used global gravity model EIGEN-CG03C and from the bedrock topography (Fig. 6), respectively, were computed similarly as described in Sect. 3.3 and restored to the

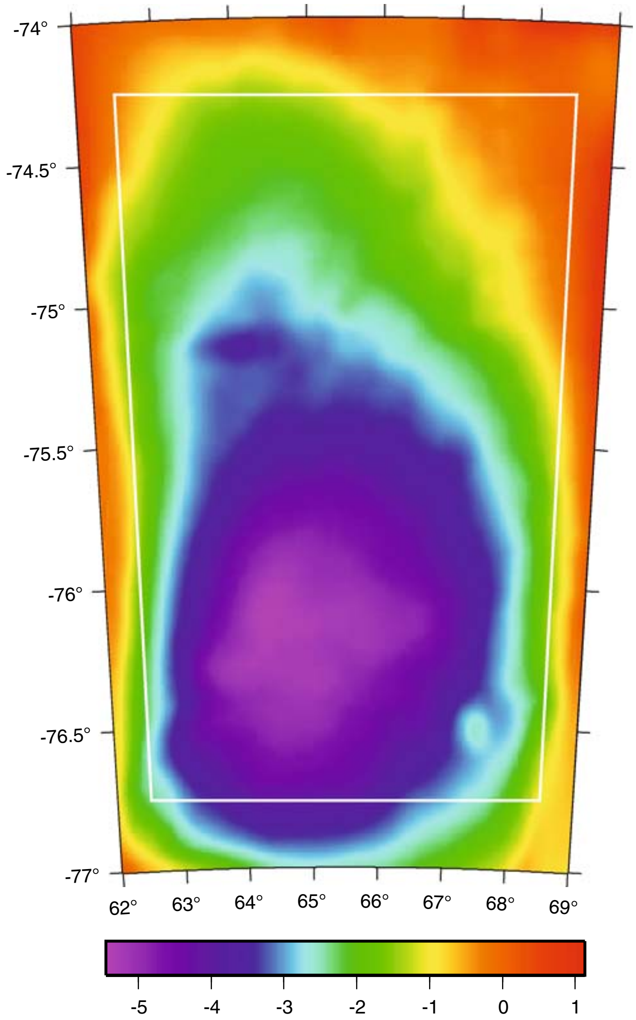

Fig. 10 Residual height anomalies (units: m; polar stereographic projection). The *white box* shows the area where the final regional quasigeoid is provided (cf. Sect. 3.4 and Fig. 11)

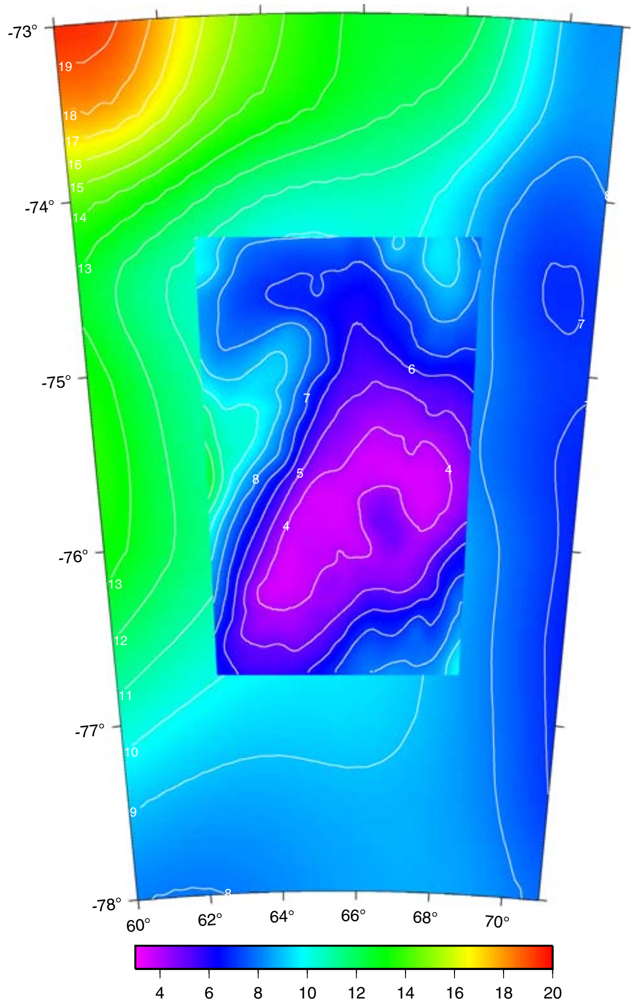

Fig. 11 Improved regional quasigeoid (height anomalies, units: m; polar stereographic projection). In the background, the global model EIGEN-CG03C (Förste et al. 2005), which was applied in the remove step, is plotted. The improved regional geoid provides a finer resolution compared to the global model

residual quasigeoid signal. The resulting final regional quasigeoid is shown in Fig. 11. In the background the global model EIGEN-CG03C (up to degree and order 360) was plotted at a larger area to visualize the improvement of the regional quasigeoid computation. Compared to the global model, which covers long-wavelength variations of about 3 m in the area of interest, a signal in the range of up to 6 m has been added (cf. also Fig. 10). The improved regional quasigeoid reveals a much finer resolution which is due to the observed airborne gravity and topography data. The pattern repeats the features visible in the bedrock topography data (Fig. 6), especially the central graben structure of the Lambert rift system, and the topography peaks (i.e. lower ice thicknesses), which are more dominant in the north of the area.

Discussing the accuracy of the obtained regional geoid model, one has to take into account two types of error sources—erroneous data on one hand, and errors originating in the model approximations and computation steps, on the other hand. The airborne gravity data observed by the PCMEGA survey is at the typical accuracy level of airborne gravimetry, which was considered by introducing the a priori standard deviation of 5 mGal (see Sect. 3.3). Furthermore, the gravity data are limited to a bounded region, a fact which will inherently induce edge effects.

The topographic data, which were patched by less accurate data in the outer regions (see discussion in Sect. 3.2, Figs. 5, 6), are not that important with respect to the error budget because they were used only for the remove step to produce band-pass filtered gravity disturbances and then to restore the contributions in terms of quasigeoid heights. The global gravity field model EIGEN-CG03C has a cumulative error of about 10 cm at degree 120 (Förste et al. 2005). Nevertheless, based on CHAMP and GRACE data, the global model provides highly accurate information for the long

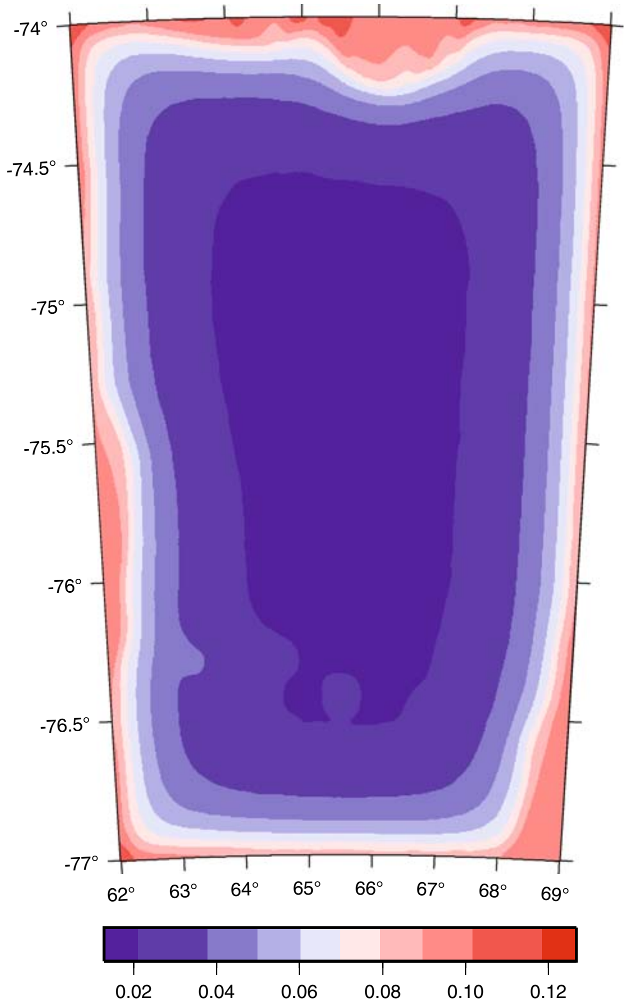

Fig. 12 Standard deviation of the computation step, resulting from LSC (units: m; polar stereographic projection)

wavelengths (larger than 300 km) which helps to minimize possible biases. The LSC provides a formal standard deviation which is plotted in Fig. 12 and can be taken as a measure for the inner accuracy of the computation step. Mostly, this standard deviation is less than 7 cm. Clearly, the aforementioned edge effects become visible by larger errors (more than 9–10 cm) at the border of the area of computation. Consequently, the area where the final regional quasigeoid is provided (Fig. 11) has been cut at the edges by about 25 km compared to the area of computation in order to count for the edge effects. Summarizing all error sources discussed above, the accuracy of the final regional quasigeoid can be estimated to be at the level of 15 cm.

4 Conclusions

The PCMEGA airborne survey provided valuable and consistent data sets for gravity, ice thickness and bedrock topography in the southern Prince Charles Mountains area, East Antarctica. Using these data the feasibility of a regional geoid improvement could be demonstrated, applying the remove–compute–restore technique in connection with the residual terrain method and least-squares collocation. The final regional geoid resembles the dominant features of the bedrock topography, especially the graben structure of the Lambert glacier system. The accuracy of the regional geoid is estimated to be at the level of 15 cm. To strive for an accuracy at the 1 cm level and to make full use of the high-quality long-wavelength information of the global gravity field models resulting from CHAMP, GRACE and future GOCE data, one has to take into account adapted computation strategies (modified Stokes kernels, refined topographic corrections) as discussed, e.g. by Sjöberg (2005). Considering the current data situation in Antarctica, the accuracy level of 1 dm is a realistic and appropriate goal in that area of the world. The data coverage in Antarctica will most likely be subject to major improvements when further airborne surveys will be carried out. The International Polar Year 2007/ 2008 provides a reasonable framework for international and interdisciplinary cooperation in that field. With regard to IAG, the Commission Project 2.4 “Antarctic Geoid” works towards the goal of closing the gaps in the gravity data coverage and improving the geoid in Antarctica.

Acknowledgments The “Prince Charles Mountains Expedition of Germany and Australia” (PCMEGA) was funded by the Federal Institute for Geosciences and Resources (BGR), Germany, and the Australian Antarctic Division. We like to thank all colleagues who contributed to the great success of PCMEGA. Parts of this work were funded by the German Research Foundation (DFG) under Grant DI-473/17. M. Scheinert likes to acknowledge the cooperation and fruitful discussions with all members of the International Association of Geodesy Commission Project 2.4 “Antarctic Geoid”. Finally, we like to acknowledge the work of the editor-in-chief, R. Klees, and of L. Sjöberg, H. Denker and two further anonymous reviewers whose remarks helped to improve the manuscript.

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