Satellite and airborne gravimetry: their role in geoid determination and some suggestions

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Abstract

This paper will cover a variety of topics. First, it will briefly overview the GRACE (Gravity Recovery and Climate Experiment) and GOCE (Gravity field and steady-state Ocean Circulation Explorer) satellite mission concepts, with a view to the improvements made (and to be made) to the global gravity field. Second, it will summarise some results of the assessment of the recent EGM2008 global gravity field model, which has a spatial resolution of about 10 km. Third, it will describe the computation and evaluation of the AUSGeoid09 model that will be released by Geoscience Australia in the very near future. All three topics will be set in the framework of the restrictions of current data and how airborne gravimetry can contribute. With the increased interest in coastal zone mapping because of threats like sea level change and tsunamis, airborne gravimetry can bridge the gap between land and satellite-altimeter-derived gravity data. As such, a proposal will be made to collect airborne gravimetry in key Australian coastal zones, but preferably along the entire coastline! Another area that lacks gravity data is Antarctica, which can adversely affect global gravity field models (the polar-gap problem). Airborne gravimetry has already been used to survey the gravity field of the Arctic, so another proposal will be made to collect airborne gravity over Antarctica. Of course, both are ambitious and massive projects, but it is important to consider them as valuable applications of airborne gravimetry.

Introduction

Gravity data, in whatever functional form, is the key ingredient in the computation of the geoid, which is the equipotential surface of the Earth's gravity field that corresponds most closely with mean sea level in the open oceans and if there were no other perturbing forces such as currents. A common application of the geoid is for the determination of physically meaningful heights from GPS (e.g., Featherstone, 2008), but it has applications in oceanography and geophysics. In 1849, G.G. Stokes first showed how to solve a free boundary-value problem in potential theory to compute the geoid from gravity anomalies[1]. Other adaptations are also available, such as M. Hotine's integral for the determination of the geoid from gravity disturbances[1] as a solution of a fixed boundary-value problem.

These boundary-value approaches have since been extended to the computation of the geoid from airborne gravimetry and airborne gradiometry (e.g., Schwarz, 1996). This and other variants have now been applied in practice, first using gravimetry from the Greenland airborne geophysics project (Brozena, 1991). Suggested methods and practical geoid computations from airborne gravimetry are described by, e.g., Schwarz and Li (1996), Forsberg et al. (2000), Novàk et al. (2003) and Serpas and Jekeli (2005), among others. This has also spawned airborne geoid mapping systems for geodetic and other applications (e.g., Forsberg et al., 1996; Bastos et al., 1997).

Numerous gravimetric geoid models have been computed all over the world, as well as global models, often in terms of a truncated series expansion in terms of scalar spherical harmonic basis functions. Useful reviews of earlier global gravity field models are given by Lambeck and Coleman (1983), Nerem et al. (1995) and Rapp (1997), whereas more recent models are documented and stored for free download from the International Centre for Global Gravity Field Models (http://icgem.gfz-potsdam.de/ICGEM/ICGEM.html).

Since the, 1960s, global models have relied heavily upon satellite-based geodetic observations to provide the long-wavelength components of the Earth's external gravity field (e.g., Kaula, 1966;

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[1] See Hackney and Featherstone (2003) for a definition and discussion of the subtle differences between gravity 'anomalies' and gravity 'disturbances'.

Reigber, 1989), but some of these global models use additional terrestrial gravimetry from a variety of sensors, including gravity anomalies derived from satellite radar altimetry and a limited amount of airborne gravimetry. Broadly, these can be classified between satellite-only and combined global geopotential models.

More recently, the CHAMP (Challenging Mini Satellite Payload; Reigber et al., 1999), GRACE (Gravity Recovery and Climate Experiment; Tapley et al., 2004) and GOCE (Gravity field and steady-state Ocean Circulation Explorer; Drinkwater et al., 2003; Johannessen et al., 2003) satellite missions have significantly enhanced the long-wavelength determination of the gravity field (cf. Balmino et al., 1999). However, because of attenuation of the gravitational signal with altitude, satellite-only models are always of long wavelength in nature. The short wavelengths can be supplemented by terrestrial gravity data (land, marine, altimeter, airborne), but only where it is available.

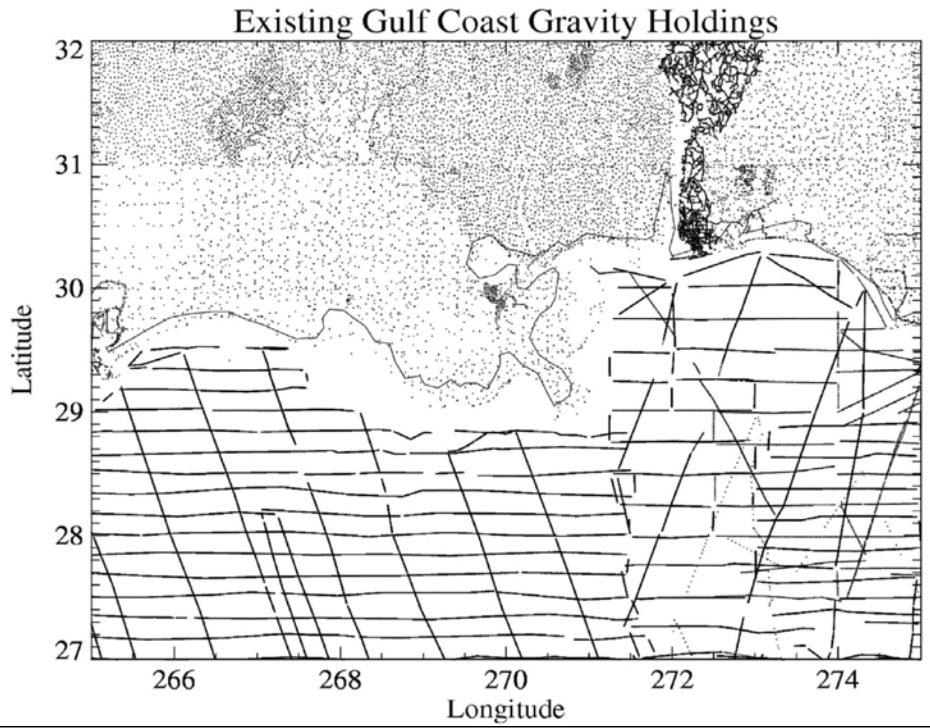

Figure 1. An example of the data gaps of between, 20 km and, 200 km in the coastal zone between land and marine gravity observations (image courtesy of Dru Smith, US NGS).

Notable gravity data gaps in terrestrial gravity data coverage are over Antarctica and almost all coastal and estuarine zones (cf. Figure 1), which is where airborne gravimetry can contribute quite significantly.

- Relatively few gravity observations have been collected in the Antarctic (e.g., Diehl et al., 2008; Scheinert et al., 2008; Jordan et al., 2009; McLean et al., 2009), whereas the Arctic was airborne gravity surveyed relatively recently as part of the Arctic Gravity Project (Kenyon et al., 2008; http://earth-info.nga.mil/GandG/wgs84/agp/). A key benefit of collecting gravity data over Antarctica is that it will help solve the so-called polar gap problem (e.g., Sneeuw and van Gelderen, 1997; Albertella et al., 2001; Rudolph et al., 2002).
- Coastal and estuarine zones lack gravity data because of navigation restrictions for ship-borne gravimetry and because gravity anomalies derived from satellite altimetry, even if re-tracked (cf. Sandwell and Smith, 2009; Andersen et al., 2010), remain contaminated in this region (cf. Deng et al., 2002; Deng and Featherstone, 2006). This is where airborne gravimetry can help (cf. Hwang et al., 2006). Figure 1 shows an example of a coastal gravity data gap centred on New Orleans in the US, which will be filled by airborne gravity as part of the US GRAV-D project (described later).

Modern satellite gravimetry

At the broadest conceptual level, dedicated satellite gravimetry missions observe (either directly or indirectly) the Earth’s external gravitational gradients. This is essentially through differential measurements between two (or more) points, thus largely eliminating correlated errors. This can take two approaches (e.g., Rummel, 1979, 1986; Rummel et al., 1999; Jekeli, 1999; Rummel et al., 2002): satellite-to-satellite tracking (SST) or a dedicated gravity gradiometer instrument onboard a satellite.

The SST methods can use either low-low inter-satellite tracking (ll-SST), where two low-Earth orbiting satellites track one another (Wolff, 1969; Kaula, 1983; Wagner, 1987; Cui and Lelgemann, 2000; Cheng, 2002), or high-low inter-satellite tracking (hl-SST), where high-Earth orbiting satellites (notably GPS) track the low-Earth orbiting satellite(s) (Schrama, 1991; Visser and van den IJssel, 2000). The satellite(s) being tracked should be in low orbits, with the proof masses isolated, as-best-as-possible, from the perturbing effects of atmospheric drag. Both SST methods can be applied to satellite gravity gradiometry (cf. Ditmar et al., 2003).

Various such missions have been proposed for over two decades, such as GRAVSAT (Piscane, 1982; Wagner, 1983), STAGE (Jekeli and Upadhyay, 1990), Aristotles (Visser et al., 1994), STEP (Albertella et al., 1995; Petrovskaya, 1997) and SAGE (Sansò et al., 2000). However, only now have dedicated satellite gravity field missions been launched, most notably GRACE and GOCE.

GRACE

GRACE (the Gravity Recovery and Climate Experiment) is a joint US-German mission to map both the static and a time-variable parts of the Earth’s external gravity field (Tapley et al., 2004; http://www.csr.utexas.edu/grace/; http://grace.jpl.nasa.gov/). Temporal gravity variations have been monitored every month (30 days) or less (1-10 days) by different groups. GRACE is used for a variety of scientific applications, including oceanography (surface and deep-ocean currents, mass and heat content change, sea-level change); hydrology (seasonal storage of surface and subsurface water, evapotranspiration); glaciology (ice-sheet mass change, sea-level change); solid Earth geophysics (glacial isostatic adjustment, mantle viscosity, lithosphere density) and geodesy (global and regional geoid modelling, precise satellite orbit determination).

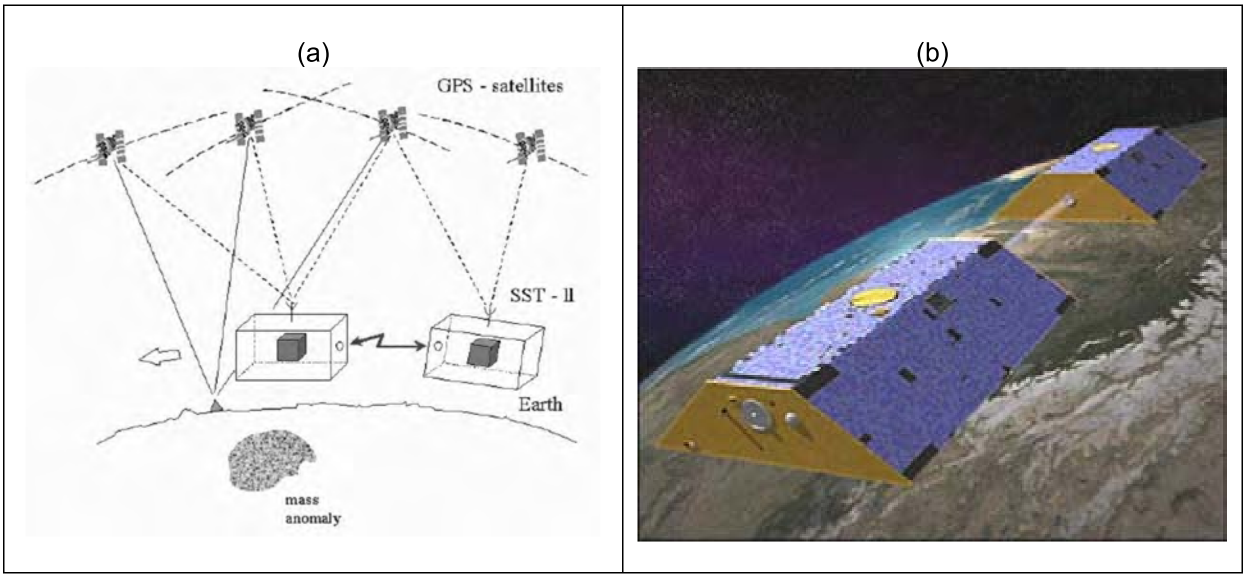

Figure 2. (a) The GRACE concept of satellite-to-satellite tracking in the low-low mode combined with satellite-to-satellite tracking in the high-low mode (from Rummel et al., 2002). (b) Artist’s impression of the GRACE satellites in orbit (from http://www.csr.utexas.edu/grace/).

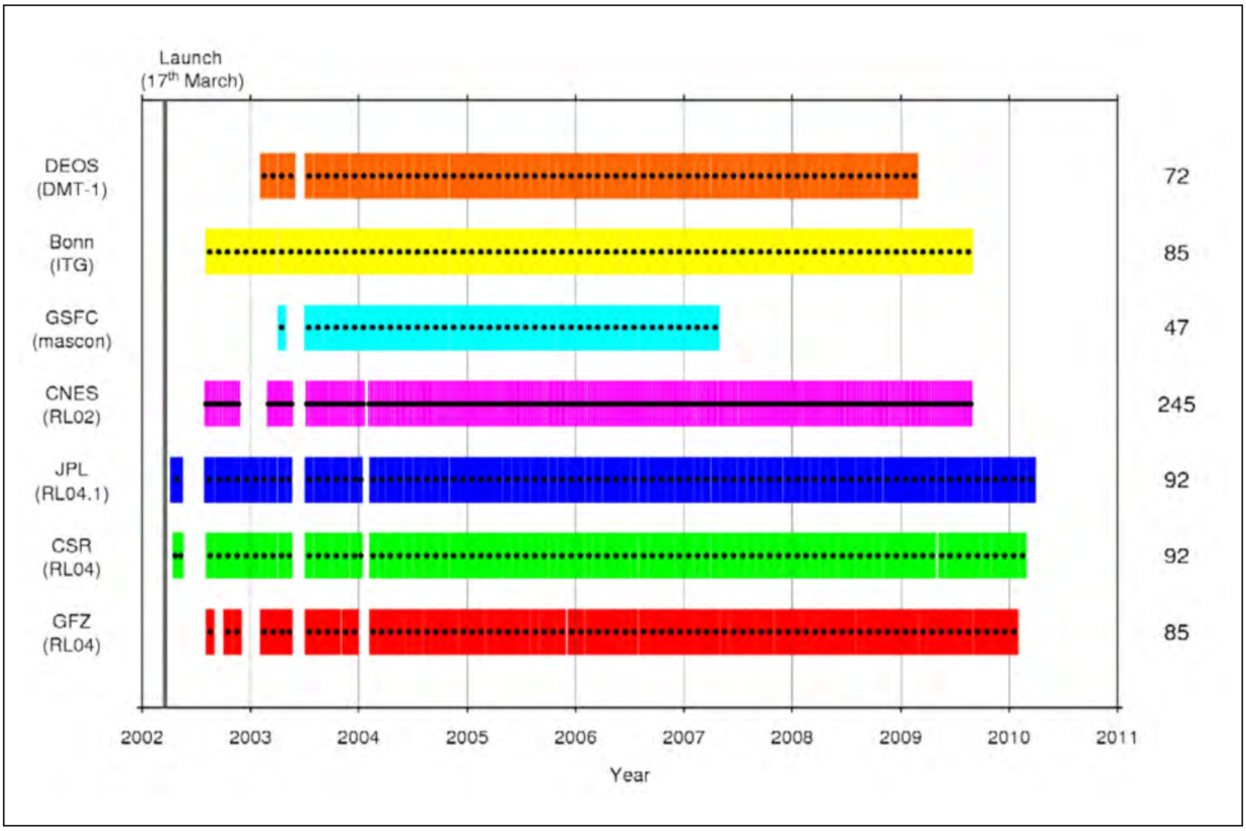

Figure 3. Time-variable GRACE gravity field solutions from various groups. The total number of solutions for each group are shown on the right hand side. Delft University of Technology (DEOS), University of Bonn (ITG), NASA's Goddard Space Flight Centre (GSFC), The French Centre National d'Études Spatiales (CNES), NASA's Jet Propulsion Laboratory (JPL), University of Texas (CSR) and the German GeoForschungs Zentrum (GFZ). The gaps in some of the time-series are when the GRACE satellites were going through periods of orbital resonance. (Image courtesy of Kevin Fleming, Curtin University)

The GRACE mission consists of two near-identical satellites following one other in the same near-circular polar orbit, at a ~498-km launch altitude that has now decayed to ~460 km, and separated by a distance of ~220 km; a so-called tandem formation. The ll-SST is measured using K-band ranging (KBR), coupled with GPS-based hl-SST tracking of both satellites (Figure 2). These data are processed in different ways by several different groups worldwide to yield models of the Earth's external gravity field (Figure 3). GRACE was launched on 17 March, 2002 with an anticipated five-year lifespan, but the mission is still operating and may extend to, 2013 or, 2015 (12-14 years) because of the design life of the satellite components (batteries, altitude, propellant, thrusters, solar panels).

GRACE-based time-variable and static gravity models suffer from some deficiencies: the time-variable solutions are only really reliable to spherical harmonic degree ~60 and have to be 'destriped' to remove correlated errors (Swenson and Wahr, 2006) and filtered, usually using a Gaussian filter of width 300-600 km (Wahr et al., 1988), though many other filters are available. Spectral and spatial leakages also have to be handled before interpretation (e.g., Baur et al., 2009). The exact methods vary among 'users' and there is not yet a general consensus on the 'best' approach to these problems. The stripes also contaminate the static GRACE gravity field solutions, so they either have to be truncated to a lower degree or destriped. These problems can also be removed by combining the GRACE static solution with terrestrial gravimetry (see section on EGM2008).

GOCE

GOCE (the Gravity field and steady-state Ocean Circulation Explorer) is a European Space Agency (ESA) satellite mission to map the global static gravity field using gravity gradiometry (http://www.esa.int/export/esaLP/goce.html; Drinkwater et al., 2003; Johannessen et al., 2003). Due to the use of a low-Earth orbiting (~250 km) gradiometer, and based on numerous simulations, it should be able to determine the static gravity field to an accuracy of ~1 mGal in terms of gravity anomalies and ~1-2 cm in terms of geoid undulations (cf. Tscherning et al., 2002) down to spatial scales of ~100 km. A geoid model of this accuracy is important as a global reference surface for geodesy (e.g., unification of height datums), and studies of Earth-interior processes, ocean current circulation, ice motion and sea-level change, among others.

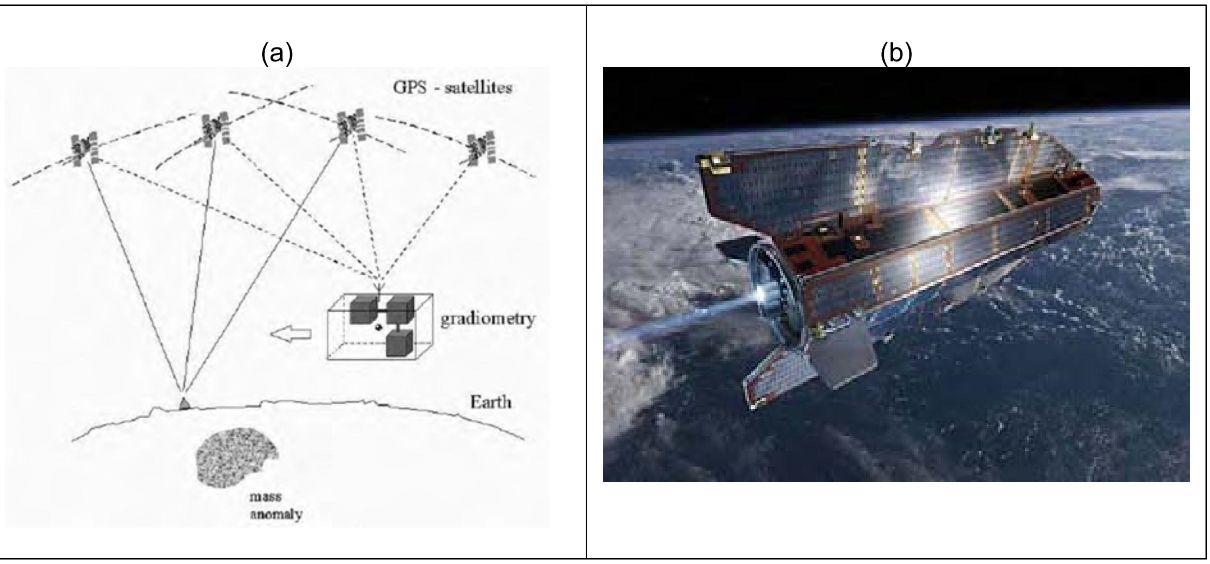

Figure 4. (a) The GOCE concept of satellite gravity gradiometry combined with high-low satellite tracking (from Rummel et al., 2002). (b) Artist's impression of the GOCE satellite in orbit (from http://www.esa.int/export/esaLP/goce.html).

The GOCE satellite was launched on 17 March, 2009, with an expected mission duration of ~20 months, but given the experience with GRACE it is possible that this could be extended, but atmospheric drag is more of a restriction for the lower orbiting GOCE spacecraft (~250 km vs ~460 km). GOCE orbits in a dawn-dusk Sun-synchronous orbit at 96.7 degrees inclination. Importantly, the GOCE satellite will house a dedicated three-axis electrostatic gravity gradiometer, which is used to determine the static gravity field in combination with hl-SST using GPS (Figure 4). This will allow determination of the stationary global gravity field at a spatial resolution of ~100-km, though there will be circular data gaps centred at the poles (e.g., Sneeuw and van Gelderen, 1997; Albertella et al., 2001; Rudolph et al., 2002). This is why airborne gravimetry should be collected in Antarctica (see later), which can be integrated to refine GOCE data (cf. Bouman and Koop, 2001).

EGM2008

EGM2008 (Pavlis et al., 2008) is a recent combined global gravity field model. It is provided in terms of fully normalised spherical harmonic coefficients to degree 2190 and order 2160. This corresponds to a spatial resolution (half-wavelength) of ~10 km at the equator. These coefficients can be used to synthesise any gravity field functional, including the full gravity gradient tensor, using public-domain software provided at the NGA website http://earth-info.nga.mil/GandG/wgs84/gravitymod/egm2008/index.html.

EGM2008 uses a GRACE-derived satellite-only model from the Institute for Theoretical Geodesy (ITG) at the University of Bonn, Germany. While this GRACE-only model resolves the gravity field to degree and order 180, the higher degree coefficients are far less reliable (see earlier). This results in large errors in areas with no terrestrial gravity data, notably in Antarctica where errors in the geoid height can be larger than one metre (Morgan and Featherstone, 2009). There is also the problem of striping in the static GRACE solution that is not removed in areas devoid of terrestrial gravity data. These issues provide justification for airborne gravimetry over Antarctica.

In other areas where gravity data are protected by commercial or military confidentiality clauses, the EGM Development Team had to reconstruct free-air gravity anomalies from Bouguer anomaly maps and topographic elevations from the Shuttle Radar Topographic Mission (SRTM). Sri Lanka is one example of where this occurred, but where we were able to access observational data to test EGM20008 (Abeyratne et al., 2009). The standard deviation of fit of EGM2008 to, 207 GPS-levelling points is ±0.184 m and to, 20 gravity points on benchmarks is ±6.743 mGal, showing the reconstruction to be effective.

In Australia, Claessens et al. (2009) have shown that EGM2008 reveals some known and unknown problems in the AUSGeoid98 national geoid model (Featherstone et al., 2001). Figure 5 shows long-wavelength differences related to the use of GRACE data in EGM2008 (cf. Featherstone, 2007), linear differences offshore due to unadjusted ship-track gravity data (cf. Featherstone, 2009), and the large difference in the Gulf of Carpentaria is probably due to mis-modelling because of a 1m non-barotropic tide (Tregoning et al., 2008).

EGM2008 performs very well in Australia because a substantial amount of gravity observations have been in the public domain for many years. We helped ensure that the EGM Development Team was provided access to GADDS (http://www.geoscience.gov.au/gadds), but they also had access to additional gravity data not stored in GADDS (Claessens et al., 2009). There are several datasets in the Australian region with which to evaluate EGM2008, the results of which are summarised in Table 1. Of interest is the rather good fit to the BRAGS airborne gravity survey (Sproule et al., 2001) over the northern Barrier Reef. For a more complete analysis, which includes various subsets and localised study areas, see Claessens et al. (2009).

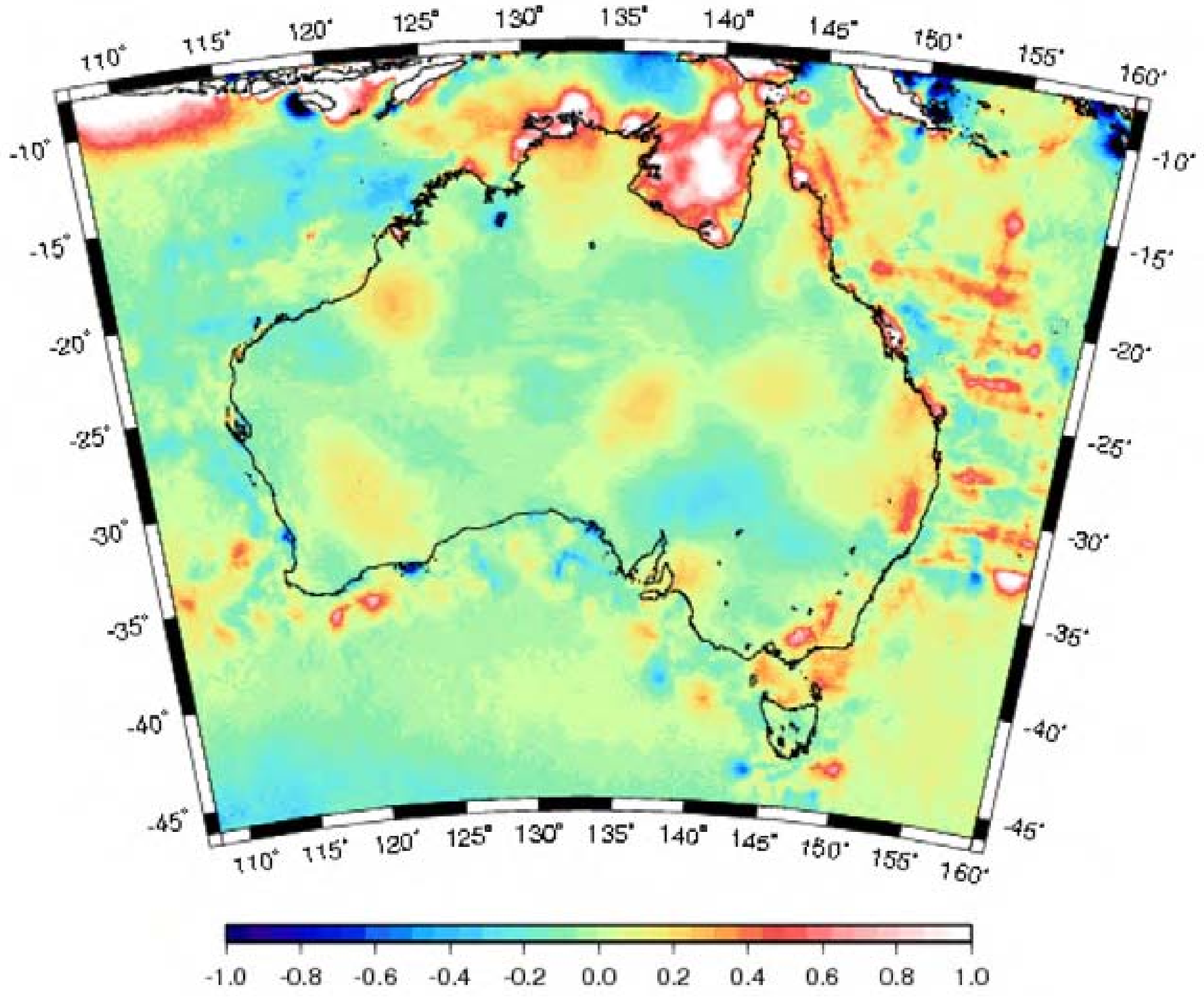

Figure 5. Differences (metres) between EGM2008 and AUSGeoid98 [Lambert projection].

Table 1. Fit of EGM2008 to Australian gravity-related data (from Claessens et al., 2009). Units of measurement for the maximum, minimum, mean and standard deviations statistics are metres.

data	*# points*	*max*	*min*	*mean*	*std*
All GA GADDS land free-air gravity anomalies (mGal)	1,304,904	+192.294	−88.756	+0.498	±5.541
GA GADDS free-air anomalies not used in EGM2008 (mGal)	548,787	+191.677	−67.641	+0.566	±6.373
BRAGS airborne free-air anomalies (mGal)	6,725	+13.239	−22.434	−2.495	±3.954
GPS-levelling data (m), no bias and tilt applied	254	+0.648	−0.535	+0.063	±0.242
GPS-levelling data (m), with bias and tilt applied	254	+0.571	−0.701	+0.000	±0.173
north-south vertical deflections (arc-sec)	1,080	+17.69	−6.99	−0.62	±1.17
east-west vertical deflections (arc-sec)	1,080	+8.70	−11.34	+0.10	±1.28

AUSGeoid09

AUSGeoid09 will soon replace AUSGeoid98 as the national standard for the transformation of GPS heights to the Australian Height Datum (AHD), and should be released by Geoscience Australia as soon as final tests are completed. Its gravimetric component was computed by the author and colleagues at Curtin University of Technology. This was based on EGM2008 to spherical harmonic degree 2190 and adds regional gravity data via numerical integration of a deterministically modified Stokes integral using efficient FFT techniques. The full details will be in a forthcoming paper in the Journal of Geodesy (Featherstone et al., 2010).

The input data to AUSGeoid09 comprise EGM2008, land gravity anomalies recomputed from raw data in the GADDS database (http://www.geoscience.gov.au/gadds), DNSC2009GRA marine gravity anomalies derived from re-tracked satellite radar altimetry (Andersen et al., 2010), gravimetric terrain corrections computed from the GEODATA DEM of Australia (cf. Kirby and Featherstone, 2002), and around 1,000 GPS-levelling points first used to test the computations.

Because EGM2008 is already such a good fit to the Australian gravity field (refer to the section on EGM2008), only a small residual quasigeoid signal of up to ~20-30 cm is needed (Figure 6). The larger differences in mountainous regions are most probably due to the use of different digital elevation models: EGM2008 uses a DEM based on SRTM data, whereas AUSGeoid09 uses the GEODATA v3 Australian DEM. Other differences are omission errors because EGM2008 has a spatial resolution of 5 arc-mins, whereas AUSGeoid09 has a spatial resolution of 1-arc-min.

The improvements in the fit to GPS-levelling data were only very marginal. The standard deviation of fit to a newly reprocessed set of 911 GPS heights at readjusted AHD benchmarks only improved from ±0.138 m for EGM2008 alone to ±0.129 m for the gravimetric-only component of AUSGeoid09. However, the quality of the AHD heights is the major limiting factor in this evaluation (cf. Featherstone, 2004, 2006; Featherstone and Filmer, 2008; Filmer and Featherstone, 2009).

Because of the deficiencies in the AHD, the released version of AUSGeoid09 has been fitted to the AHD using least-squares collocation (Featherstone and Sproule, 2006) to give a model of the base of AHD rather than the classical quasigeoid (cf. Featherstone, 1998). This approach is pragmatic because the majority of the users of AUSGeoid09 will want to determine AHD heights from Global Navigation Satellite Systems (GNSS, notably GPS). The gravimetric version of AUSGeoid09, called AGQG2009, is available from the author for scientific purposes only, but must not be distributed in case it causes confusion with the official AUSGeoid09 product, especially as AUSGeoid09 will be a national standard.

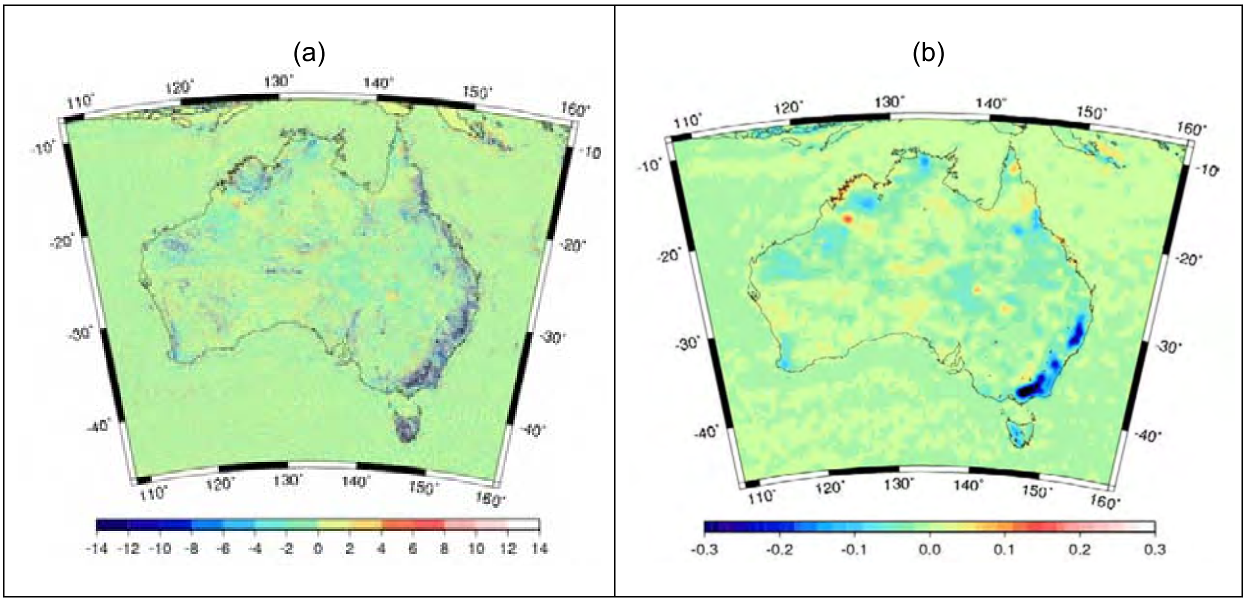

Figure 6. (a) Residual terrestrial gravity anomalies (in mGal), and (b) residual quasigeoid undulations (in m) used to compute AUSGeoid09 as a regional refinement to EGM2008.

Suggestion 1: Airborne gravity over the Australian coasts

Although AUSGeoid09 has dispensed with erroneous ship-track data (cf. Featherstone, 2009) and used marine gravity anomalies from re-tracked satellite radar altimetry (Andersen et al., 2010), it is still likely to be less accurate in coastal and estuarine zones (cf. Andersen and Knudsen, 2000). This is because ship-track gravimetry is limited by near-shore navigation restrictions (and poorly modelled tides in the coastal zone) and radar re-tracking is imperfect because of backscatter from the land and inland water bodies (cf. Deng et al., 2002). Also see the discussion by Hipkin (2000) on the problems of modelling the geoid in the coastal zone. A poor geoid model in these zones is of particular concern in Australia because the majority of the population and thus users of AUSGeoid09 reside near the coast. At present, there are no obvious means with which to solve these problems with the existing data restrictions.

As such, it is suggested that Australia collects airborne gravimetry over coastal and estuarine zones. Initially, and based on cost considerations, acquisition should be carried out near the most populated areas or those that are low-lying and thus vulnerable to marine inundation (sea level change, storm surges, tsunamis). Ideally, however, the whole coastline should be airborne gravity surveyed. In this regard, the US National Geodetic Survey has already embarked on its GRAV-D project to collect airborne gravity in coastal zones, with a view to better mapping of the geoid in the coastal zone (http://www.ngs.noaa.gov/grav-d/). There is no reason why Australia should not replicate this initiative. It will also allow for the better integration of topographic and bathymetric data that have been collected by GPS-controlled air- and ship-borne surveys (e.g., LIDAR and SONAR).

Suggestion 2: Airborne gravity over Antarctica

As shown in Morgan and Featherstone (2009) and summarised earlier, EGM2008 does not perform well in Antarctica. This is simply because the EGM Development Team did not have access to many terrestrial gravity observations over this region. Instead, they had to rely on the degree-180 GRACE-only ITG model from the University of Bonn. GRACE-derived gravity field models are very unreliable at these high degrees, where correlated noise and aliasing cause striped patterns. Because satellites cannot sense the high-frequency gravity field, the only solution in Antarctica is to collect terrestrial gravity (i.e., ground, vehicle, or airborne observations). These data will also help address the polar gap problem in GRACE- and GOCE-only models (cf. Sneeuw and van Gelderen, 1997; Albertella et al., 2001; Rudolph et al., 2002).

Because of the harsh and restricted access in Antarctica, airborne gravity is the most logical choice. Moreover, the experience from the Arctic Gravity Project (Kenyon et al., 2008; http://earth-info.nga.mil/GandG/wgs84/agp/) and some area-limited surveys in Antarctica (Diehl et al., 2008;

Scheinert et al., 2008; Jordan et al., 2009; McLean et al., 2009) shows that it is feasible. There is no reason why Australia should not replicate the Arctic Gravity Project.

Concluding comment

This paper has overviewed the GRACE and GOCE satellite mission concepts, presented results of some assessments of EGM2008, and described the computation and evaluation of AUSGeoid09. All were set in the framework of the restrictions of current data and how airborne gravimetry can help to provide solutions. One proposal is to collect airborne gravimetry in key Australian coastal zones, but preferably the entire coastline! Another proposal is to collect airborne gravimetry over Antarctica to close the polar-gap problem. While both are ambitious projects, it is important to consider them as valuable applications of airborne gravimetry.

Acknowledgments

Will Featherstone is the recipient of an Australian Research Council Professorial Fellowship (project number DP0663020). This work was also supported under the Australian Research Council's Discovery Projects funding scheme (project number DP0877381). The views expressed herein are those of the author and are not necessarily those of the Australian Research Council. Thanks go to J. Hicks for proofreading and resulting productive discussions and to the reviewers of this paper.

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