The potential of ground gravity measurements to validate GRACE data

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Abstract. New satellite missions are returning high precision, time-varying, satellite measurements of the Earth's gravity field. The GRACE mission is now in its calibration/-validation phase and first results of the gravity field solutions are imminent. We consider here the possibility of external validation using data from the superconducting gravimeters in the European sub-array of the Global Geodynamics Project (GGP) as 'ground truth' for comparison with GRACE. This is a pilot study in which we use 14 months of 1-hour data from the beginning of GGP (1 July 1997) to 30 August 1998, when the Potsdam instrument was relocated to South Africa. There are 7 stations clustered in west central Europe, and one station, Metsahovi in Finland. We remove local tides, polar motion, local and global air pressure, and instrument drift and then decimate to 6-hour samples. We see large variations in the time series of 5–10 μgal between even some neighboring stations, but there are also common features that correlate well over the 427-day period. The 8 stations are used to interpolate a minimum curvature (gridded) surface that extends over the geographical region. This surface shows time and spatial coherency at the level of 2–4 μgal over the first half of the data and 1–2 μgal over the latter half. The mean value of the surface clearly shows a rise in European gravity of about 3 μgal over the first 150 days and a fairly constant value for the rest of the data. The accuracy of this mean is estimated at 1 μgal, which compares favorably with GRACE predictions for wavelengths of 500 km or less. Preliminary studies of hydrology loading over Western Europe shows the difficulty of correlating the local hydrology, which can be highly variable, with large-scale gravity variations.

Key words. GRACE, satellite gravity, superconducting gravimeter, GGP, ground truth

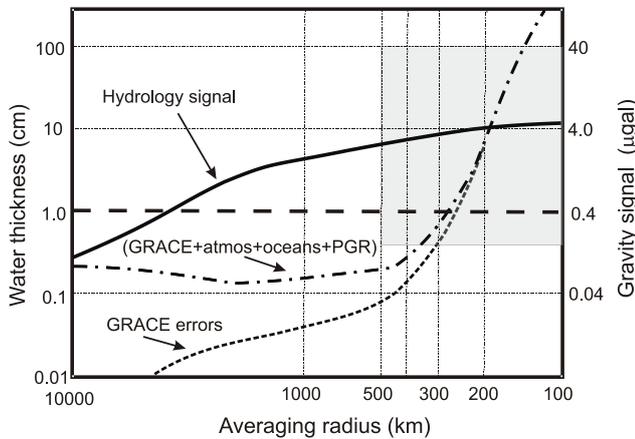*Correspondence to:* D. Crossley (crossley@eas.slu.edu)

1 Introduction

The stimulus for this study originated with Wahr et al. (1998) who discussed the expected accuracy of surface gravity fluctuations for the proposed GRACE mission. They demonstrated that a careful accounting for the various contributions to time varying gravity would permit the determination of small time-varying signals such as variations in continental water storage. It was immediately clear that we should consider the possibility of combining satellite data and ground-based data from the Global Geodynamics Project (GGP; Crossley et al., 1999). The GGP superconducting gravimeter (SG) network is far too sparse geographically to be suitable as a global gravity field, but there are sub-arrays of instruments, particularly in Asia and Europe, that warrant closer consideration. Preliminary attempts at producing a ground-based map of gravity variations were reported by Crossley and Hinderer (1999) and later by Crossley and Hinderer (2002).

Recently, Velicogna and Wahr (2001) suggested that ground based gravity measurements cannot usefully contribute to the validation or analysis of GRACE data. They argue that the radius over which a single ground-based measurement extends (several 10's km) is incompatible with the wavelengths of satellite-derived fields (> 200 km). Second, using GGP data from the International Centre for Earth Tides, they used statistical arguments to argue against any correlation of the signals over long time spans. The wavelength argument is true for a single station but the limitation can be overcome, to some extent, by the use of a gravity array. The question of the treatment of GGP data can only be answered by taking care in the analysis to preserve long-term integrity of each data set. Here we address both issues by processing the GGP data for a specific epoch, finding that the correlation between gravity variations over distances of several hundred km and time spans of several months is quite convincing.

Table 1. European GGP stations used in the analysis

Station	Code	Country	Instrument	Latitude	Longitude
Brussels	BE	Belgium	T003	50.7986	4.3581
Membach	MB	Belgium	C021	50.6093	6.0066
Medicina	MC	Italy	C023	44.5219	24.3958
Metsahovi	ME	Finland	T020	60.2172	24.3958
Potsdam	PO	Germany	T018	52.3806	13.0682
Strasbourg	ST	France	C026	48.6217	7.6838
Vienna	VI	Austria	C025	48.2493	16.3579
Wettzell	WE	Germany	SG103	49.1440	12.8780

Fig. 1. Hydrology recovery from GRACE.

Fig. 2. GGP stations July 1997 – August 1998.

2 GRACE goals

GRACE (Gravity Recovery and Climate Experiment) is a joint venture of NASA (USA), DLR (Germany), UTCSR (Texas), and GFZ (Potsdam). The mission has now been actively collecting data for about 9 months and the first results are to be reported soon (AGU Abstracts, Fall Meeting 2002). The high accuracy anticipated of GRACE data should enable subtle time variations in the gravity field to be found, i.e. changes in continental water storage, the variability of ocean bottom pressure, and the redistribution of snow and ice. These changes will be determined by successive spherical harmonic solutions of the data with a limiting ground resolution of 100–200 km and intervals of 2–4 weeks.

The methodology follows the sequence (Wahr et al., 1998):

– assume a density change $\Delta\rho$ in a layer of thickness H (10–15 km) surrounding the Earth's surface (i.e. the lower atmosphere and upper hydrosphere).

– convert $\Delta\rho$ to a surface density distribution $\Delta\sigma$ by integrating over H.

– expand $\Delta\sigma$ in spherical harmonics, with coefficients $(\hat{C}_l^m, \hat{S}_l^m)$

– relate these harmonics to the harmonics $(\hat{C}_l^m, \hat{S}_l^m)$ of the gravity field, determined from the GRACE satellite orbit data, approximately every 14 days.

– deduce $\Delta\sigma$ from $(\hat{C}_l^m, \hat{S}_l^m)$, and thus infer $\Delta\rho$ by assuming H.

Note that $\Delta\sigma$ does not distinguish between water, ice, or snow. It is also evident that the GRACE data will be time-aliased if there is any unmodeled variation of gravity on time scales less than 2 weeks, as seems probable for the atmosphere and oceans (e.g. Flechtner et al., 2002).

One of the examples considered by Wahr et al. (1998) is for Manaus, Brazil, in the Amazon River Basin (Fig. 1). The upper curve shows the predicted hydrology signal, the middle curve is the expected errors in GRACE data with all sources of modeling (PGR is post glacial rebound), and the lower trace is for GRACE errors alone. The accuracy of the recovery using the full 5 years of data is 2 mm of water at length scales longer than 400 km; a more recent estimate indicates better than 1 cm at 200 km or longer (Swenson et al., 2002). The errors at shorter wavelengths rise rapidly, becoming excessive at wavelengths less than 200 km. The shaded box is the region where GRACE errors and GGP network errors are expected to overlap. To be competitive, ground-based gravity measurements need (a) to cover wavelengths between 100 and 1000 km and (b) to reach accuracies of less than 0.4 μgal at wavelengths between 200 and 300 km. If both conditions are satisfied, we may claim that ground-based (in
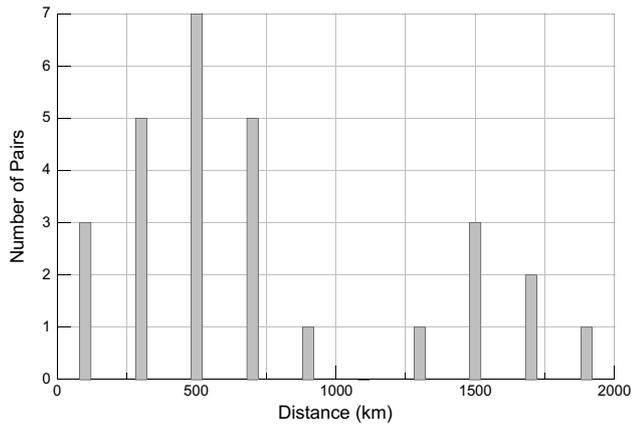
Fig. 3. Station distribution by pairs.

this case GGP) gravity can be used to 'validate' satellite measurements.

3 GGP data

The stations used in this study are shown in Table 1 and Fig. 2. All stations except ME are in the middle of the Western European landmass. The time period of this analysis was chosen to begin at the start of GGP (97/7/1) and continue to the end of the recording at PO (98/8/31). The PO SG was then reconstructed as a dual sphere instrument and moved to South Africa (Neumeyer et al., 2001). Station MC is not officially a GGP station, but data are available for this study through the work of Zerbini et al. (2001). Station BE stopped recording in 2000 and the instrument at Wettzell, which was a prototype compact dewar model (designation SG103) with unusually large drift (Harnisch et al., 2000), has been replaced with a new dual sphere model. Also a new station, Moxa, was started in 2000 (Kroner et al., 2001). The distribution, or spacing, of the 8 stations taken in pairs, is plotted as a histogram in Fig. 3. The distance range of 200–1000 km is well covered, but the inclusion of a single distant station (ME) extends the coverage up to 2000 km.

4 Processing

The first step is to remove a modeled tide from each station using local tidal gravimetric factors (δ, κ) obtained from independent analyses of data from each station. We include all waves with periods up to a month. For semi-annual and longer periods we use nominal elastic gravimetric values of (1.16, 0) to avoid fitting artificially the residual annual signals. We also remove the effect of local atmospheric pressure using a nominal admittance of $-0.3\,\mu$gal mbar^{-1}; using slightly different values will not be a major source of error in the final result. The residual series are displayed in Fig. 4. It is clear station WE has a large negative drift that appears
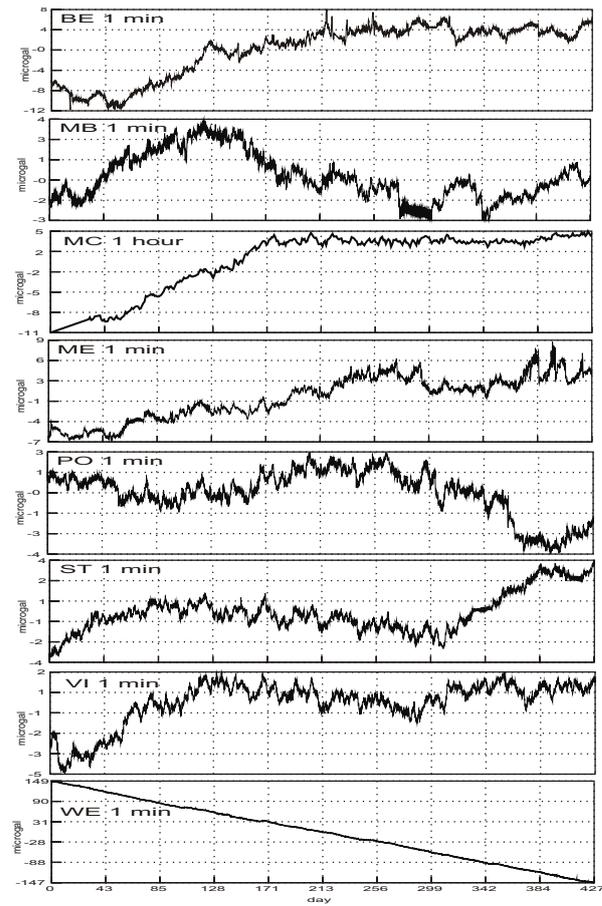

Fig. 4. Gravity residuals after removal of tides, local pressure and polar motion. Note the different scales of each data set.
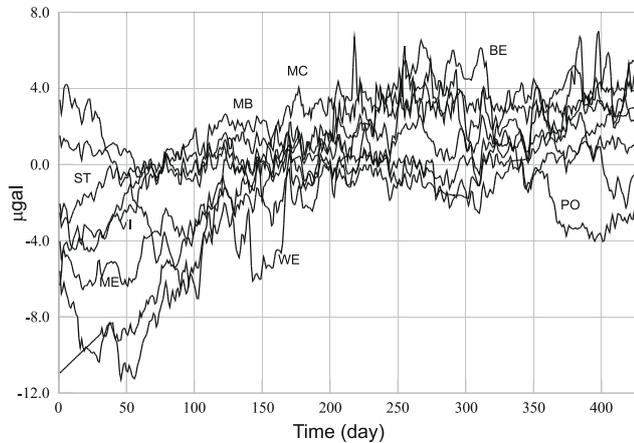
linear. IERS-derived polar motion was also subtracted from each data set.

5 Instrument drift

For WE, we fit simultaneously a linear drift function and a series of offsets (already corrected in Fig. 4) at fixed time locations; this is done iteratively to arrive at an appropriate correcting function. In Fig. 5 the residuals are now plotted on a common axis, with the drift of station WE removed. We now remove the instrument drift at the other stations. Rather than do the analysis solely on the basis of the 14 months of data at our disposal, we requested assistance from the station operators who have analyzed their data over much longer time periods. The results are shown in Table 2. It is seen that apart from WE (discussed above), most of the stations have drift rates in the 1–4 μgal yr^{-1} range. The most reliable estimates come from comparisons with Absolute Gravimeter data, but at some stations this was not possible. At one station (MC), the data were further checked using a series of GPS measurements (Zerbini et al., 2001) to establish vertical motion. The drift of station MB is a high compared to the other stations,

Table 2. Drift functions removed for each station

Station	Author	Drift (μgal/yr)	Function	Method
BE	B. Ducarme	0.0	exponential	Estimated from last disturbance
MB	M. van Camp et al.	4.90	linear	Comparison with AG
MC	B. Richter	2.50	exponential	Comparison with AG and GPS
ME	H. Virtanen	3.75	linear	Comparison with AG
PO	J. Neumeyer	1.64	linear	Fit to long series
ST	J. Hinderer et al.	3.65	linear	Comparison with AG
VI	B. Meurers	2.25	linear	Fit to long series
WE	G. and M. Harnisch	-253.95	linear	Fit to series

Fig. 5. Gravity residuals, 1 hour, mean values removed.

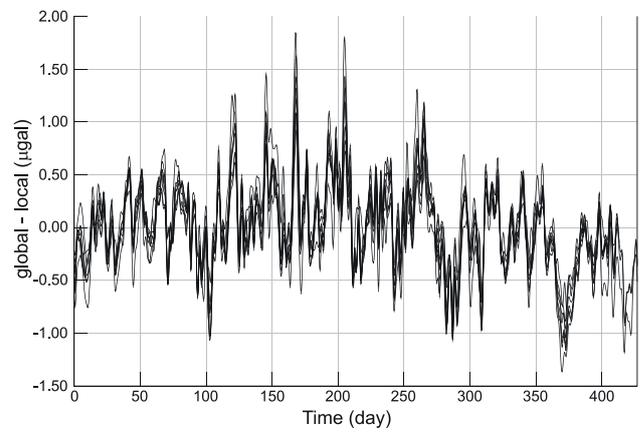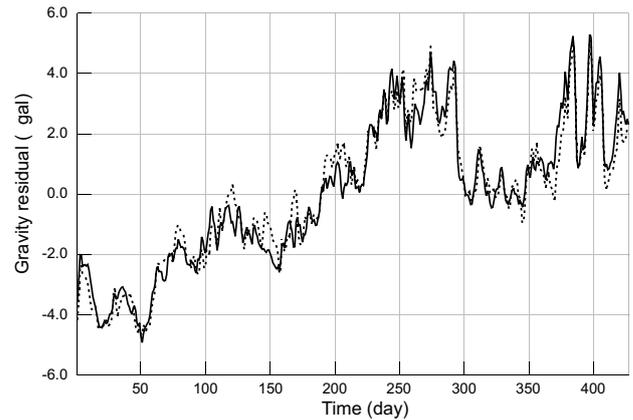**Fig. 6.** Global vs local atmospheric pressure loading.

but it has been carefully checked and has found to be reliable (van Camp et al., 2002).

6   Global pressure loading

We now correct for the non-local atmospheric pressure effects, first decimating the data further to 6-hour samples. The global atmospheric loading has been calculated by Jean Paul Boy (personal communication) using the method described in Boy et al. (2001). The assumption is that the vertical column is hydrostatic and so the mass attraction and loading are dependent only on the surface pressure, here obtained from the ECMWF.

The results for the 8 stations are shown in Fig. 6 as the difference between local and global loading. The differences between global and local corrections are significant (-1.5 to $+2\,\mu$gal) over short periods, but there is little or no long term trend. More importantly, all stations respond in a similar fashion, indicating that the global loading is intergrating over atmospheric masses of the same size or larger than this station distribution.

To illustrate the effect of the global loading on the gravity residuals, we show the results for ME, where the differences are the largest (Fig. 7, in which the dashed line is the

Fig. 7. Effect of global pressure loading at station ME.

local correction). Clearly the inclusion of global pressure does not significantly affect the trend of the gravity residuals. This type of computation has since been updated (Boy and Chao, 2002) by using a three-dimensional atmospheric model and they find that seasonal changes due to global loading are more significant than those shown here. Their method will be incorporated in future work on this project.

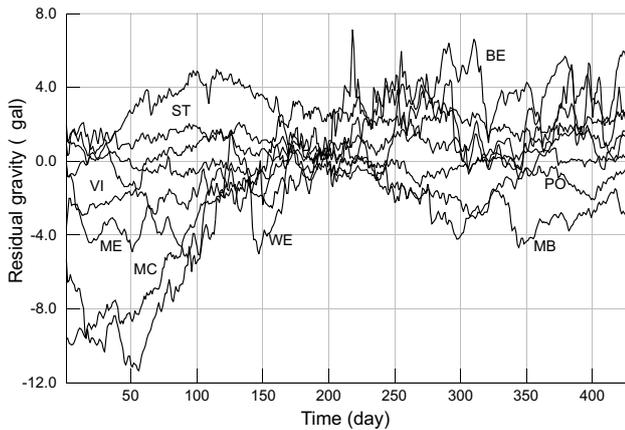Figure 8 shows the final residual gravity, corrected for the

Fig. 8. Final residual gravity after all corrections.

Fig. 9. Spatially smoothed residuals, 6-hour sampling.

Fig. 10. Spatially smoothed residuals, 14-day sampling.

drift in Table 2 and the atmospheric loading in Fig. 6. It can be seen the the series are somewhat flatter than Fig. 5, but still with a significant spread of values, especially during the first 150 days.

7   Spatial averaging

We need to consider how to spatially average the individual station residuals to simulate the integrating effect of the satellite measurements. To estimate the spherical harmonic coefficients of a global model from such a limited amount of ground data would definitely yield poor coefficients, so we proceed differently.

We first of all fit a minimum curvature surface to the data points on which Fig. 8 is derived. This fit is performed for each 6-hour sample of the field, and one of the properties of the surface is that it goes through each of the original points. This is therefore a good interpolation procedure and we can produce contour maps of the surface as a function of time. These maps do not do any spatial averaging of the field and neighboring stations with conflicting series (e.g. BE and MB) still show up as inconsistencies (we cannot shows these maps here due to lack of space). As a second step we therefore fit a polynomial surface to the data using the whole of this interpolated surface (not just the original data points), in order to get a robust least squares solution. We choose a third degree polynomial because this gives a reasonable smoothing about a wavelength of 500 km. Higher order polynomials may also be justified, but we have not investigated all possibilities. The resulting surface is then re-sampled at each of the original station locations and a set of smoothed time series is produced (Fig. 9). The series now show much less deviation and the data that stands out from the rest is ME, due partly to its distance from the other stations and the poor control due to the lack of neighboring stations.

We now re-sample the residuals to 14 days to represent satellite repeat determinations of the field, to produce the series in Fig. 10. We claim that this figure represents our inter-
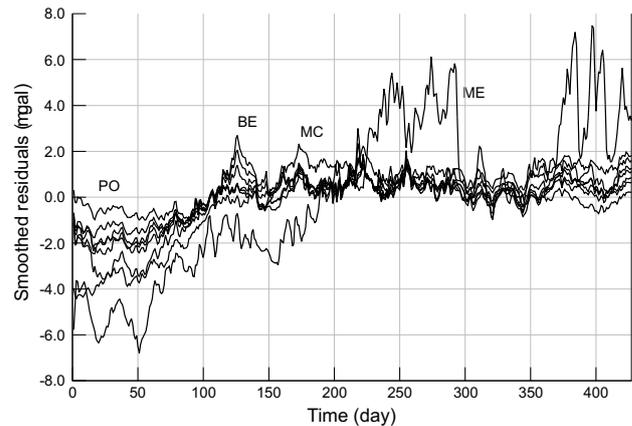
pretation of a time and space averaged picture of the gravity field. It is difficult to place error estimates on these series due to the various processing steps involved, but one further step might be to average all the series together (except ME since this is an outlying station) and compute the errors in this final average. This is done in Fig. 11, which is our final result. It shows that the 1σ errors are of the order of 1–2 μgal for the first 150 days, and less than 1 μgal for the reset of the time period.

8   Hydrology

So far we have made no allowance for hydrology, either large scale or local, because in fact this signal will also be seen by the satellite. The variability of continental water storage is a prime target for GRACE and clearly one that affects surface gravity measurements. Van Dam et al. (2001) showed that at the GGP stations their models predict large variations of water-induced gravity changes that are frequently dominated by annual variations of the order of 10 μgal. Such effects are
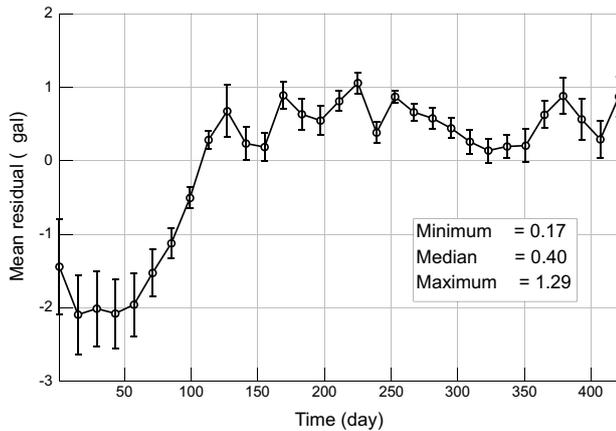
Fig. 11. Evolution of the mean field, ME omitted.

difficult to separate from other possible annual signals in the residuals, e.g. global pressure effects.

At the same time the analysis of local hydrology is often usefully done to remove a water signal in gravity recordings, e.g. Crossley et al. (1998), Kroner (2001), and Virtanen (2001). Frequently there is a high correlation between local measures, such as rainfall and water table depth, and residual gravity and this correlation is used to justify the modeling. One source of uncertainty is the interpretation of the resulting admittance, as this depends on the local porosity. Frequently the empirical estimate of porosity obtained from the admittance is difficult to verify on physical grounds.

A long-term project has been initiated to estimate continental water storage in central Europe. We (Florsch and Llubes, 2002) have performed loading estimates of large-scale (2000 × 3000 km) hydrology over Europe by placing a 1 m water load on the continental crust. The vertical crustal displacement reaches a maximum of 4.6 cm at the center of the load and less than 1.0 cm outside the area. This gives a central effect of 14μgal in gravity to which must be added about 42μgal in direct Newtonian attraction; the total effect is therefore 0.56μgal per cm water (at 100% porosity). At a smaller scale we have simulated a loading of 1 m water over the Alsace region of the lower Rhine Graben (about 20 × 300 km) and find a very localized loading effect of 1.6μgal, with again the dominating direct effect of 42μgal, thus a total of 0.44μgal per cm water.

As is widely recognized, hydrology can be extremely complex at regional scales. A good example can be found at the website (http://aesn.brgm.fr/bulletin/nappes. html) run by l'Agence de L'Eau Seine-Normandie, in which the water table has been monitored at over 60 sites over the Seine Basin since 1975. The correlation between neighboring sites is often poor because of the geological variability, (http://aesn.brgm.fr/bulletin13/images/situation_ nappe.gif), whereas some sites are well correlated even at relatively large distances. Determining the true hydrological signal will inevitably require a combination of direct water table measurements and gravity observations, the latter including both ground-based and satellite data.

9 Discussion and conclusions

One may, with some justification, question the somewhat ad-hoc procedure used to get from the gravity residuals (Fig. 8) to the smoothed integrated curve of the gravity field evolution (Figs. 9 to 11). At the present time we are considering alternative ways of doing this. Nevertheless, it is evident that even in Fig. 8. there is spatial and temporal coherency of the field over the 427 days, and this becomes more obvious in the smoothed product, Figs. 9 and 10. Station ME is unusual in more than its geographic isolation from the other stations. As Virtanen et al. (2002) have shown, the loading effects of the Baltic Sea are quite strong and account for much of the variability seen in Fig. 10. So far this loading has not been corrected in the current study, but it will be seen by a satellite, so one has to be careful to compare fields that have been consistently processed.

Further work is being done to extend these series to more recent years, in particular into 2000 when the CHAMP satellite started to produce results. It is difficult to produce a gravity surface in real time from GGP data due to the care needed in processing and the need to make systematic absolute gravity measurements to check the drift. Nevertheless the GGP data certainly enables such maps of the evolution of the European gravity field to be made. A longer-term goal might be to establish further SGs in the missing regions (e.g. Spain, Poland, Northern Germany) that would undoubtedly significantly improve the quality of the gravity field estimation. In future we believe this work will provide a useful source of data with which GRACE and other satellite missions may be compared.

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