Creutzfeldt, B., Güntner, A., Klügel, T., Wziontek, H. (2008): Simulating the influence of water storage changes on the superconducting gravimeter of the Geodetic Observatorc Wettzell, Germany. - Geophysics, 73, 6, WA95-WA104

DOI: 10.1190/1.2992508.

GRAVITY MODELING OF HYDROLOGICAL VARIATIONS

Simulating the influence of water storage changes on the superconducting gravimeter of the Geodetic Observatory Wettzell, Germany.

Benjamin Creutzfeldt[1]

Andreas Güntner[1]

Thomas Klügel[2]

Hartmut Wziontek[3]

[1] Helmholtz-Centre Potsdam – German Research Centre for Geosciences (GFZ), Section Engineering Hydrology, Telegrafenberg, D-14473 Potsdam, Germany. Email: benjamin.creutzfeldt@gfz-potsdam.de; guentner@gfz-potsdam.de

[2] Federal Agency for Cartography and Geodesy (BKG), Section Geodetic Observatory Wettzell, Sackenrieder Straße 25, D-93444 Bad Kötzting, Germany. Email: kluegel@fs.wettzell.de

[3] Federal Agency for Cartography and Geodesy (BKG), Section National Reference Systems for Gravity, Richard-Strauss-Allee 11, D-60598 Frankfurt am Main, Germany. Email: hartmut.wziontek@bkg.bund.de

Manuscript submitted: July 21, 2008

Abstract

Superconducting gravimeters (SG) measure temporal changes of the Earth's gravity field with high accuracy and long term stability. Variations in local water storage components (snow, soil moisture, groundwater, surface water and water stored by vegetation) can have a significant influence on SG measurements and – from a geodetic perspective – add noise to the SG records. At the same time, this hydrological gravity signal can provide substantial information about the quantification of water balances.

A 4D forward model with a spatially nested discretization domain was developed to investigate the local hydrological gravity effect on the SG records of the Geodetic Observatory Wettzell, Germany. The possible maximum gravity effect was investigated using hypothetical water storage changes based on physical boundary conditions. Generally, on flat terrain, a water mass change of one meter in the model domain causes a gravity change of 42 μGal. Simulation results show that topography increases this value to 52 μGal. Errors in the Digital Elevation Model can influence the results significantly. The radius of influence of local water storage variations is limited to 1000 m. Detailed hydrological measurements should be carried out in a radius of 50 to 100 m around the SG station. Groundwater, soil moisture and snow storage changes dominate the hydrological gravity effect at the SG Wettzell. Using observed time series for these variables in the 4D model and comparing the results to the measured gravity residuals show similarities in both seasonal and shorter-term dynamics. However, differences exist, e.g. the range comparison of the mean modeled (10 μGal) gravity signal and the measured (19 μGal) gravity signal, making additional hydrological measurements necessary in order to describe the full spatio-temporal variability of local water masses.

Introduction

The interrelation of hydrology and gravity is attracting increasing attention in hydrological and geodetic sciences. From a hydrological perspective, the estimation of water storage and its spatio-temporal variation is important in order to quantify water balances and for effective water use and management. Direct measurements of water storage changes (WSC) are, however, still a challenging task. Gravity observations provide a promising tool. From a geodetic perspective, the local hydrological gravity effect is an interfering signal which adds noise to gravimetric measurements and must therefore be eliminated from the gravity records.

Generally, the hydrological gravity effect is caused by the gravitational attraction of water masses and their variation. The interrelation of WSC and gravity variation is expressed by Newton's law of gravitation and is the basis for ground-based observations with gravimeters as well as for remote sensing carried out by the GRACE satellite mission. In addition to Newton's attraction term, variations of the Earth's gravity field due to the elastic deformation of the Earth's crust caused by the water load have to be taken into account at the global and regional scale (Farrel, 1972).

Bonatz (1967) was the first to study the relationship between hydrology and gravity by simulating the effect of soil moisture on gravity measurements. He concluded that it is not useful, for geodetic applications, to develop gravimeters with higher accuracy because soil moisture can contribute to the gravimetric signal up to 10 μGal and more. Nonetheless, more accurate gravimeters were developed – superconducting gravimeters (SG), for example. These are relative gravimeters with a measurement resolution of 0.01 μGal but, due to noise of atmospheric or seismic origin, their measurement accuracy decreases and ranges between 0.1 to 1 μGal (Goodkind, 1999; Hinderer et al., 2006). Damiata and Lee (2006) investigated the relationship between groundwater properties and the gravitational response. They concluded that it is necessary, from the hydrological perspective, to develop inexpensive gravimeters with sub-μGal accuracy. SGs are cost intensive in acquisition and operation, but are state of the art in terms of temporal resolution, stability and accuracy. Hence, they are suitable for studying the interrelationship of hydrology and gravity. For geodetic and hydrological applications, the problem remains that gravimetric records are an integral signal of accelerations of different origins (Earth and ocean tides, mass redistribution in the atmosphere, oceans, polar motion, continental hydrology, etc.). It is therefore still a challenging task to separate the influence of WSC from the rest of the signal.

Generally, gravity variations caused by water storage changes can be represented either by empirical relationships or by a physically-based approach. The empirical relationship is either established using a simple regression between hydrological and gravity data (Imanishi et al., 2006; Harnisch et al., 2006a), or by translating the WSC to gravity changes on the basis of the Bouguer approximation. Within the Bouguer approximation, the water mass change is treated as the change in thickness of an infinitely extended plate (e.g. one meter change in water height causes a change in gravity of 42 μGal). Here, the information on WSC is often derived from precipitation and/or groundwater level measurements (Peter et al., 1995; Bower and Courtier, 1998; Crossley et al., 1998). The advantage here is that detailed hydrological properties and processes do not have to be considered and the gravity signal can be corrected for the hydrological influence without knowledge of the complex hydrological system.

A more physically based approach is the analytical solution of a circular disc to calculate gravity change from WSC in the geometric body (Bonatz, 1967). Although the spatial distribution of masses can be considered to a certain extent by dividing the cylinder into many disks and nesting different cylinders (Abe et al., 2006), complex situations cannot be modeled with that method.

Because water masses are highly variable in space and time, emphasis was placed on the development of physically-based 4D models to investigate the hydrological gravity effect (Virtanen, 2001; de Jong and Ros, 2004; Hasan et al., 2006; Hokkanen et al., 2006; Meurers et al., 2006; Van Camp et al., 2006; Naujoks et al., 2007a; Prutkin and Klees, 2007). The actual model implementation varies depending on the subject of the study. Often, rectangular prisms are used for the abstraction of the topography and the subsurface (Hasan et al., 2006; Kroner et al., 2007), but spherical segments (Neumeyer et al., 2004) or 3D polygons (TINs) (Hokkanen et al., 2006) can also be used. These models allow for the exact calculation of the gravity signal caused by WSC and for the consideration of the topography and spatial distribution of masses.

The beginnings of these different methods, initially developed in geodesy and applied geophysics, date back to the 1960s. They are mainly used to eliminate the effect of topographic masses on gravity anomalies (terrain correction), to fulfill the boundary condition in determining the Earth's gravitational potential, or for exploring the composition of the upper crust. Nonetheless, computation of hydrologically-induced gravitational effects differs from standard application in terrain correction in terms of variability of density in space and time, in the magnitude of the effect and intended accuracy in the sub μGal range.

Decomposition of masses into rectangular homogeneous prisms is common both in geophysics and in geodesy. This is reflected by the extensive literature on the subject, including Mader (1951), Nagy (1966), Kolbenheyer (1967), Ehrismann (1973) and Forsberg (1984). An alternative and more general approach is the use of homogeneous polyhedra (see e.g. Rausenberger (1888), Götze (1976), Pohanka (1988) and Petrović (1996)), which tends to be a more efficient method in terms of the relationship between processing time and accuracy (Petrović and Skiba, 2001). Neglecting the dimensions of the volume elements and assuming the concentration of the mass at a single point is a simple approach, allowing direct application of Newton's law. An intermediate solution is given by MacMillan (1958), where a series expansion for rectangular homogeneous prisms leads to a modified point mass representation. Especially in the near field, accuracy depends on the distance between the element center and the point of observation and on the element size. Mufti (1973) shows that the MacMillan approach is applicable for terrain corrections at all points with a distance not less than three times the side-length of the corresponding element.

To overcome accuracy limitations in the near field, Klügel and Wziontek (2007) combined an analytical solution with the point-mass equation in a correction model for atmospheric effects on gravimetric time series. In their approach, the mass attraction in the immediate area around the gravimeter is calculated using an analytical expression for cylindrical disks, while the point-mass model is used for more distant areas. The approach of using different models for different areas of influence around a gravimeter was also pursued by Leirião (2007) and He (2007). For the closest area around the gravimeter, they use a prism equation which integrates fully over each volume element. To reduce computing time for large distances, the point-mass equation is used, while for intermediate distances the MacMillan equation is used. Finally, this model switches between these equations depending on a criterion that relates distance to cell size.

Some open questions remain regarding 4D modeling of WSC. Firstly, topographic information needed for 4D modeling is generally provided by a Digital Elevation Model (DEM). Its accuracy depends on the data source (topographic map, laser scan, survey) and the resolution. Consequently, the question is: How do topography and the DEM accuracy influence the gravity calculation? The second question is related to the scale. Llubes et al. (2004) concluded that for hydrological effects on gravimeters, it is only necessary to distinguish between the local and global scale. In their study, the influence of the local scale was set roughly to a few kilometers. For local 4D modeling, however, appropriate hydrological data on WSC are vital and therefore more exact knowledge of the radius of influence is needed for proper instrumentation with hydrological measurement devices. As total WSC is usually composed of several water storage components (snow, soil moisture, groundwater, surface water, water stored by vegetation), the question is how the gravitational signal is influenced by each of these storage components and how the radius of influence varies for each component. A final question is whether a planar approximation of the Earth's gravity field is sufficient in the model domain if the local scale is set to a few kilometers.

In this study, an appropriate 4D model is set up to investigate these open questions using the SG at the Geodetic Observatory Wettzell in Germany as an example. The possible maximum influence of all hydrological WSC components is assessed and the influence of real hydrological data on SG measurements is analyzed.

Study area

The study area is the area around the SG, which is located in a small building at the Geodetic Observatory Wettzell, operated by the Federal Agency for Cartography and Geodesy (BKG). The dual sphere SG CD029 is positioned near ground level and is based on a concrete foundation (Figure 1). Accuracy estimates for the SG gravity change recordings are better than 1 μGal and may even reach 0.1 μGal because the station is located in an area with little noise interference. Earlier studies showed that the SG records are influenced by WSC. Based on a simple regression model, Harnisch and Harnisch (2002, 2006b) estimated the effect of precipitation and groundwater level on the SG Wettzell to be up to 14 μGal and 20 μGal, respectively.

The station is located on a mountain ridge of the Upper Palatinate Forest. The mean annual precipitation amounts to 863 mm, the potential evapotranspiration according to the Haude equation is 403 mm, and the mean annual temperature is 7 °C. Land use is characterized mainly by grassland and forestry.

The geology of the station can be classified into 4 different zones: (1) the soil zone with mainly loamy-sandy brown soils (Cambisols) and with an underlying solifluction layer, (2) the weathered zone mainly out of grus (physically-weathered gneiss), (3) the fractured zone and (4) the basement zone. This classification is based on data from 11 boreholes with a mean depth of 22 m (see Table 1 for a detailed description of the different zones).

The hydrogeological situation is characterized by a highly-variable, complex and unconfined groundwater table with a mean groundwater level of 8 m and a seasonal fluctuation of 3-4 m. The two surface water bodies, the Höllenstein (storage volume 1.4 km³) and the Blaibacher (storage volume 1.5 km³) reservoir, are located at a distance of 1500 m to 3000 m from the station (Figure 2).

Method

The model for simulating the homogeneous elementary body attraction of spatially-distributed WSC is implemented in MATLAB and is based on the MacMillan equation (MacMillan, 1958) presented by Leirião (2007):

$$\Delta g = G\Delta\rho\Delta x\Delta y\Delta z\left[-\frac{z}{d^3}-\frac{5}{24}\frac{\left(\alpha x^2+\beta y^2+\omega z^2\right)z}{d^7}+\frac{1}{12}\frac{\omega z}{d^5}\right] \tag{1}$$

with $\alpha = 2\Delta x^2 - \Delta y^2 - \Delta z^2$, $\beta = -\Delta x^2 + 2\Delta y^2 - \Delta z^2$, $\omega = -\Delta x^2 - \Delta y^2 + 2\Delta z^2$ and $d = \sqrt{x^2 + y^2 + z^2}$. X, y and z are the centre coordinates of an elementary body relative to the sensor (m). Δx, Δy and Δz are the side lengths of a rectangular body (Δxy = Δx = Δy) (m), G is the universal gravitational constant (Nm²/kg²) and Δρ is the density change in an elementary body (kg/m³). The z component of gravity variation is calculated for each body and the total gravity effect is derived by summation of all gravity changes in each elementary body. The MacMillan equation has the advantage over the point-mass approximation, in that the shape of cuboids is considered (Leirião, 2007). In contrast to the prism approach (Heck and Seitz, 2007), which solves the volume integral of each body, it is possible to calculate WSC using the MacMillan approach on the basis of matrices, which significantly reduces the calculation time. This is vital for 4D gravity modeling. A nested discretization of the model domain was used to reduce the approximation error of the MacMillan equation in the near field (Figure 2). This error was assessed by comparing the results of the MacMillan approach to the gravity change, which was calculated using an analytical solution of a cylinder with the same mass variation, distance to the sensor, height and volume as in the MacMillan approach. Nested discretization also makes it possible to use high resolution DEMs in the near field and therefore allows for the consideration of detailed topographic and subsurface structures. Using a higher resolution reduces the step effect resulting from the abstraction of a continuous landscape by raster DEMs. Finally, the nested approach takes into account the fact that data availability for SG studies concerning hydrology is generally better in the immediate vicinity of the SG than at larger distances.

Two DEMs were available for the study area: A DEM 25 (cell size 25 m, extent 20 km and mean height accuracy 1-3 m) (BKG, 2004), and a DEM 10 (cell size 10 m, extent 5 km and mean height accuracy 1 m) (LVG, 2007). Data on river lines, lakes (BKG, 2005) and the base plate of the houses of the Geodetic Observatory Wettzell were also available. During a differential GPS (DGPS) survey, around 14,000 height points were collected in a region spanning 300 m around the SG (height error < 0.1 m). From these different data sources a final DEM with varying resolution depending on the distance to the SG was created by excluding the most inaccurate data (DEM DGPS) (Figure 2). Two additional DEMs were processed to assess the influence of DEM accuracy on gravity modeling: The DEM 10, where only the DGPS points had been excluded and the DEM 25 with the exclusion of the DGPS points and the DEM 10. Finally, the quality of each DEM was assessed by an independent validation set derived during the DGPS survey (Table 2). The positional reference for all data used is the projected coordinate system – Gauss Krüger, Zone 4, Bessel Ellipsoid. The normal heights are given using the DHHN 92 height system. Because all subsequent computations refer to a planar approximation of the Earth's gravity field, the influence of the curvature of the equipotential surfaces was estimated by comparing the heights with a spherical approximation using the equation of Petrahn (2000):

$$h_{corr} = h_{DEM} - \frac{1}{2R} d^2 \qquad (2)$$

where h_{DEM} is the height of the DEM (m) referring to the plane, R the Earth's radius of around 6370 km and d is the distance of a raster cell to the reference point (m). For example, at a distance of 10 km, the height decreases by about 7.8 m.

In the simulation model, the effect of WSC is calculated by assuming a homogeneous density change in an elementary body. However, WSC in ground, surface and snow water is equivalent to water level changes, meaning that the density change is heterogeneously distributed within each body. This simplification error can be minimized by reducing the layer thickness of the model.

Possible maximum influence of hydrological masses on the SG sensor was estimated assuming hypothetical maximum changes of different water storages based on real boundary conditions. Therefore, the effect of the SG house was considered in this step, assuming no WSC in the soil and weathered zone and only snow storage change on the roof of the SG house. Mass changes in the subsurface were estimated on the basis of water level change, soil porosity and specific yield (see Table 1). The mean depth of the groundwater table was calculated to be 8 ± 3 m based on three groundwater wells 200 m to 300 m away from the SG. The soil porosity was calculated from the measured bulk density and an assumed particle density of 2650 kg/m³. In the soil zone, porosity was used to estimate the maximum water mass change as soil moisture can range between saturation and a water content close to 0 Vol%. In deeper zones, it is unlikely that the moisture content would decrease to values lower than field capacity. For deeper zones, therefore, the specific yield – the drainable porosity – was used to derive the possible change of water masses. For the weathered zone, the specific yield was determined from undisturbed soil probes which were taken at depths between 1.5 m and 6.3 m, and was deduced on the basis of water retention curves (pF curves). For the fractured zone, the specific yield was roughly estimated to be around 3 %. This value was derived from a pump test using the Cooper-Jacob straight line method. According to Rubbert (2008), the specific yield varies from less than 0.1 % to up to 5 % for the fractured zone (see

Table 1). The maximum measured snow height in the study area is 0.9 m, and snow density can vary between 50 kg/m³ (freshly fallen snow) and 500 kg/m³ (old snow), so the mean snow density was set at 275 kg/m³. The effect of surface water change in both reservoirs on gravity was assessed by draining them completely, using a stage-storage relationship and starting at their maximum possible water level. Finally, the effect of water stored in the vegetation and intercepted on the vegetation surface was investigated. The interception storage capacity depends on the climate conditions and on the morphology of the vegetation cover and can be about 5 mm (1 mm water corresponds to 1 kg/m²) (Baumgartner and Liebscher, 1996). During the vegetation period, the water variations in the vegetation were set to 5 mm (Schulz, 2002). Maximum water storage change due to vegetation then amounts to up to 10 mm.

Finally, real data (daily values taken from 29 Nov 2000 to 31 Dec 2006) for groundwater, soil moisture and snow were used to model the effect of WSC on the gravitational signal. Groundwater level data were available from a monitoring well located at a distance of 200 m from the SG, where the groundwater data were adjusted to match the mean groundwater depth of 8 m. The specific yield of 5 % for the groundwater was derived by using the average specific yield of the weathered and fractured rock zone (see Table 1). Soil moisture was measured close to the station at a depth of 0.5 m with a TRIME sensor (measurement accuracy of ± 4 %). These data were applied to the topsoil layer (0-1 m depth) only. Variations in snow water storage were derived from precipitation, temperature and snow height data. Gaps in the climate time series have been filled using data from the nearby Höllenstein reservoir climate station. Precipitation and temperature are measured at the Wettzell Observatory climate station. Snow height was derived by interpolating snow heights measured at two different stations around the observatory. The snow water equivalent was determined by considering as snow all precipitation that fell when the temperature was below 0.7 °C, where snowfall takes place only when the snow height measured is greater than zero (input into snow storage). The melted snow water equivalent was calculated using the degree-day method, which is based on the positive daily temperatures (snow storage reduction) (Dyck and Peschke, 1983). For each day, the total snow storage was derived from the snow storage input and reduction. The snow density was calculated using the storage and height data. The temperature threshold of 0.7 °C and the degree-day factor of 1.1 were calibrated by detecting snow fall and thawing on webcam pictures from 2005. Visual validation using pictures from 2006 proved their suitability. The hydrological gravity effect of these real data was calculated for a quadratic area with a side length of 100 m x 100 m and the SG in the centre, assuming that water variations measured by the different sensors are constant for the whole area. 1 January 2000 was used as the reference date for gravity variations induced by WSC.

Results and Discussion

A comparison of the nested simulation model and the analytical cylinder solution shows that the difference in the results of both models is below the 0.01 µGal range for the whole area. This confirms the suitability of the nested simulation model for this study. Using the height correction according to Equation 2 results in a difference of 0.03 µGal in the simulated gravity response for the SG. This implies that the curvature of equipotential surfaces of the gravity field can be neglected in local studies.

Taking the topography into account, a water mass change of one meter (density 1000 kg/m^3) causes a gravity change of 52.49 µGal in a quadratic layer with a side length of 20 km, with a vertical distance of the WSC layer of 1 m below the gravimeter (i.e. in the top meter of the soil) and the gravimeter located in the centre of this area. The corresponding spatial distribution of the gravity response is shown in Figure 3. Gravity response refers to the total gravity signal of the SG that is caused by the assumed water mass change in the model domain. In this context, the spatial distribution of the gravity response refers to the contribution of each elementary body in the model domain to the total SG signal. Figure 3 illustrates that some areas have a negative contribution because they are located at higher elevations than the SG. Note that the spatial distribution in Figure 3 shows some discontinuities because the size of the elementary bodies changes depending on the distance to the SG in the nested modeling approach. When increasing the vertical distance of the layer where WSC occurs to 10 m and 20 m below the SG, the signal decreases to 51.33 μGal and 50.60 μGal, respectively (Figure 4). While some areas with a higher topographical position to the west of the SG show a negative gravity effect for a vertical distance of 1 m (Figure 4a), the spatial distribution becomes more symmetrical and includes only positive values for deeper WSC (Figures 4b, 4c). Comparing these results of the model for real topography with a flat terrain model, the same water mass changes in three different vertical distances amount to 41.90 μGal, 41.71 μGal and 41.53 μGal, respectively. These values are almost identical to the values derived by the Bouguer approximation, which also indicates that for surficial mass changes the main part of the signal is generated in the model domain. The effect of water mass changes on the SG sensor in Wettzell is thus intensified by 20 %, 19 % and 18 % as a consequence of the topography. For other SG stations too, Meurers et al. (2006) demonstrate that the topographical setting around the SG may have very different but significant effects on the SG signal.

As topography has a significant effect on the gravitational signal, the influence of DEM errors must be assessed. The results obtained with the different DEMs are displayed in Table 3. Assuming the same WSC as above, the difference between the results for DEM DGPS and DEM 10 falls within the range of a few μGal, whereas the difference between the results for DEM DGPS and DEM 25 is one magnitude higher and may amount to as much as 20 %. In the near field of the SG, in particular, DEM accuracy has a significant effect. At larger distances, the difference is smaller because the terrain effect weakens (Table 3) and also the topography data basis converges. DEMs with an RMSE of 1 m may still be suitable for gravity modeling when the acceptable error margin is set to 1 μGal and additional local topographic information is available. Similarly, Virtanen (1999) stresses that elevation should be mapped in detail up to a distance of about 200 m from the SG. However, DEMs with higher RMSE such as the DEM 25 are unsuitable for gravity modeling when the focus is on hydrological processes.

In the same context, information on the subsurface structures in the immediate vicinity of the SG has to be considered in the model. As presented in Figure 1, the SG is placed on a concrete foundation. In the above example, however, we assumed a water mass change of one meter in the topsoil for the entire model domain. When we take the concrete foundation of the SG into account and assume no water mass change in this particular area, the total SG signal decreases markedly by 8.47 μGal to 44.02 μGal. Similarly, the effect of the base plate of the SG building and its umbrella effect – which inhibits infiltration of rainfall into the soil – should be taken into account in the model.

Figure 5 summarizes the gravity response of the model – for real topography and taking into account the foundation – as a function of the radius of influence and a function of the depth of the mass changes. Here, radius of influence (R) refers to a square with the SG located in its centre and the square side length being twice the radius R. Both parameters – radius and depth – describe the area which gives rise to the SG gravity response. For the flat terrain model with a vertical layer distance of 1 m (and 20 m), 98 % (66 %) of the signal is generated in an area of R=50 m, 100 % (95 %) in an area of R=500 m and 100 % (98 %) in an area of R=2000 m. Signal generation for the model with real topography is as follows: 80 % (52 %) with R=50 m, 91 % (84 %) with R=500 m and 97 % (93 %) with R=2000 m. This shows that the gravity response does not simply depend on the radius of influence, but is also a function of the mass change distribution over depth and a function of topography. It follows from this that it is important to distinguish between the different water storage components because near-surface mass variations like soil moisture have a smaller sphere of influence than deeper mass changes like groundwater (see Figure 4).

In the following, we analyze separately the possible maximum effect of WSC on gravimetric observations for the different water storage components. Figure 6 shows the relationship between the gravity response and the WSC in the snow cover, soil moisture and groundwater. A groundwater table rise of 4 m from 10 to 6 m below the terrain surface over the entire model domain in an aquifer with an average specific yield of 5 %, results in a gravity change of around 10 μGal (lower section of Figure 6), whereas this value amounts to 22 μGal for a maximum specific yield of 11 %. The distinct gravity response in this theoretical experiment indicates that in addition to detailed groundwater level monitoring, precise estimation of the specific yield and of the interface between weathered and fractured zone are also vital for accurate modeling of the effect of groundwater variations. Also, van Camp (2006) pointed out the need for a detailed mapping of subsurface characteristics in combination with the recording of time series.

The unsaturated zone was divided into two layers for this modeling experiment, i.e., into soil (0-2 m) and weathered zone (2-6 m) (Figure 6). This was done not only because of the geologic situation, but also to consider the dominant processes of precipitation, evapotranspiration and drainage in the soil and of drainage in the weathered zone. In the weathered zone, a maximum soil moisture change of 7 ± 4 Vol% causes a gravity response of 17 ± 10 μGal. Water saturation in the topsoil causes a maximum gravity change of 21 ± 2 μGal. Measurements of the topsoil water content, however, show that soil moisture varies between 15 Vol% and 40 Vol%. When translated into gravity change, this results in a value of 12 μGal. The infiltration process can have a significant effect on the gravity signal because saturation and drainage account for mass redistribution in the soil column.

While WSC below the terrain surface generally leads to a gravity increase, snow reduces the gravity because it accumulates mainly above the SG sensor and on the roof of the SG building (top of Figure 6). At greater distances, the effect of snow masses is reversed because the SG is located on an elevated mountain ridge and, consequently, snow mass accumulation in the valleys adds to the gravimetric signal. Nonetheless, the highest signal change can be expected during the period of snow melt because the mass redistributes from above to below the sensor due to infiltration of snow melt water into the soil.

The gravitational effects of surface and vegetation water change are below the significance level of the SG recordings. In the unlikely event that both reservoirs are completely drained from maximum storage capacity, a gravity change of 0.19 μGal is calculated. As there are no other major water surface bodies in the study area, the effect of surface water on the SG records can be neglected in future gravity studies at the Wettzell site. However, as shown by Bonatz and Sperling (1995), water redistribution in surface water bodies may cause a significant gravity response for particular locations. The water storage capacity of vegetation is too small to cause marked gravity changes, which results in a maximum vegetation effect of 0.07 μGal.

When summarizing the above results for the individual water storage components, it follows that local hydrological mass variations can contribute up to 49 μGal to the SG records of the Geodetic Observatory Wettzell, where as much as 64 % of this signal is generated within a radius of 50 m, 90 % within a radius of 500 m, 97 % within a radius of 2000 m and 99 % within a radius of 5000 m. Note that these simulation results were obtained on the basis of parameters that represent the maximum plausible water mass variation in each storage component from a hydrological perspective. In real-world settings, however, it is very unlikely that all the changes will occur at the same time. For example, the annual phases of snow storage or groundwater level variations are frequently shifted in time compared to soil moisture variations (Güntner et al., 2007). Individual effects therefore compensate for each other in part. Also, maximum water storage change is not expected to occur homogeneously over the entire study area, as assumed here. In addition, other hydrological processes, such as evapotranspiration or groundwater discharge, which may reduce the variation in hydrological storages, are disregarded in the theoretical modeling experiments. As a consequence, the hydrologically-induced gravity response shown here can be considered as an extreme. It might only be observable in long-time records of the SG because of the negative correlation between frequency and intensity of extreme events in hydrology.

Figure 7 shows the modeled gravity response derived from the measured data for the period of Nov 2000 to Dec 2006, with the corresponding minimum and maximum possible response. The mean, maximum and minimum range of the modeled WSC effect amounts to 10.24 μGal, 19.49 μGal and 6.06 μGal. Generally, groundwater has the biggest share in the hydrological gravity signal and contributes up to 8.28 μGal by a specific yield of 5 %, 18.21 μGal by a specific yield of 11 % and 1.66 μGal by a specific yield of 1 %, followed by soil moisture with 5.46 μGal and snow with -1.25 μGal. Groundwater and soil moisture-induced gravity variations were estimated with a LaCoste & Romberg gravimeter by Naujoks (2007b) at the Geodynamic Observatory Moxa, Germany and by Mäkinen and Tattari (1988) at the Hyrylä Station, Finland to be up to 17 μGal and 13 μGal, respectively. The order of magnitude of snow-induced gravity variations is also similar to those reported by Virtanen (1999) for the Methsähovi Station in Finland. The time series of the groundwater effect is dominated by a seasonal pattern. On top of this, significant short-term variations during wet periods can also be observed. These events are also partly represented with smaller magnitudes in the soil moisture time series. During 2003, in which Central Europe experienced an exceptionally dry summer period, groundwater and soil moisture storage were depleted causing a minimum in the gravity signal. The modeled gravity signal shows a reasonable temporal correlation with the gravity residuals of the SG (corrected by tides, earth oscillation, instrument drift and atmospheric pressure). Especially during wet periods, the modeled and measured variations fit well, whereas for dry periods the time series diverge. Apart from the uncertainties due to the processing of the gravity residuals, these differences may have several reasons which are discussed in the following part.

The model based on the observation data is a simplification of the real conditions because WSC are highly variable in space and the sensors installed cannot capture this variability. With regard to groundwater, the area surrounding the observatory is highly complex due to the undulating topography combined with the geological conditions. Groundwater level measurements in nearby monitoring wells are only weakly correlated and so the assumption that the aquifer is a continuous layer that follows the topography is not valid. The aforementioned result that the modeled and measured series deviate especially during dry periods may be due to the fact that the groundwater well used here is located in a depression while the SG building is built on a ridge. This implies that groundwater below the gravimeter depletes faster while the decrease of the groundwater level at the gauge is delayed due to inflowing water from the surrounding area.

For the calculated soil moisture gravity effect it should be considered that, firstly, in hydrology, soil moisture is regarded as a highly variable parameter in space and time. Therefore, only one sensor cannot be considered as a representative sample to estimate the soil moisture dynamics and patterns around the SG. Secondly, no information was available on the soil moisture distribution over depth. Finally, apart from soil moisture measurements being influenced by different soil parameters, requiring site-specific calibration for accurate measurements, the results depend on the soil moisture measurement technique itself.

The accuracy of the snow model was not assessed statistically, but it was found to match closely with webcam photos when a visual comparison was made. Snow density, however, could not be validated by this method.

Conclusions

The results indicate that the hydrological effect of local water storage variations has to be considered in high-accuracy studies of gravity monitoring. A nested model discretization is especially useful for investigating the gravity response of SGs caused by WSC. Topography plays an important role in gravity signal modeling. Therefore, not only must topography be included in the model domain, but also DEM accuracy has to be taken into account and, if necessary, DEM uncertainties have to be reduced with supplementary measurements. The exact location of the SG, the surrounding and subsurface structures in the immediate vicinity are vital for calculation of the WSC effect on gravity. With regards to the radius of influence of mass changes on the SG recordings, gravity is not only a function of the distance of mass changes but also depends on the topography and on the distribution of mass changes over depth. For local studies at sites of superconducting gravimeters with current accuracies of about 0.1 to 1 μGal, a limitation of the radius of influence to 1000 m seems to be justified for hydrological mass variations because around 90 % of the signal is generated in this area. The greatest benefits from detailed studies of WSC can be expected within a radius of 50 to 100 m around the SG station. In local studies, the curvature of equipotential surfaces can be neglected.

The results of the modeled gravity effect for theoretical WSC, as well as for real data, lead to the conclusion that groundwater has the largest contribution to the gravity signal, followed by soil moisture and snow; but they may also indicate that until now the influence of WSC in the vadose zone might be underestimated. For the Wettzell Observatory, surface water and WSC in the vegetation cover can be neglected. The two time series of the total modeled gravity signal and the measured gravity residuals show similar seasonal and shorter-term dynamics. At the same time, several differences exist due to the limited coverage and representativeness of hydrological data, as they do not fully describe the spatio-temporal variations of water storage for the individual components. Apart from temporal changes to the groundwater level, the spatial characterization of the aquifer in terms of porosity and the transition between the weathered and fractured zone are fundamental. Therefore, complementary measurements of groundwater level variations in additional boreholes and the estimation of the physical properties of the borehole cores in combination with geophysical methods are the subjects of ongoing work in the study area.

Soil moisture contributes significantly to the gravitational signal, but is highly variable in space and time. To estimate 4D soil moisture patterns for future analyses, a detailed multi-sensor soil moisture monitoring system has recently been installed around the SG Wettzell. The snow effect can account for several μGal, but generally has the reverse effect compared to groundwater or soil moisture. Up to now, the snow storage change is estimated using a simple model, but in future it will be estimated based on direct measurements of the snow water equivalent.

Tables

Table 1. Characterization of the different underground zones by thickness (m), depth (m), porosity (%) and specific yield (%). Count refers to the number of probes. The values are mean values with the corresponding standard deviation ([1]Specific yield for the fractured zone was derived from one pump test. The value of the standard deviation was estimated from literature (see text)).

		Soil zone	Weathered zone	Fractured zone	Basement zone
Count		10	10	9	9
Mean thickness		2.0 ± 0.8	5.5 ± 3.9	8.8 ± 5.0	-
Mean depth	From	0 ± 0	2.0 ± 0.8	7.3 ± 4.3	16.1 ± 4.9
	To	2.0 ± 0.8	7.5 ± 4.1	16.1 ± 4.9	-
Porosity	Count	60	16	-	-
	Mean	43 ± 5	37 ± 4	-	-
Specific yield	Count		4	1	-
	Mean		7 ± 4	3 ± 2[1]	-

Table 2. Validation results for the different DEMs. *Count* refers to the number of validation points. *Minimum*, *Maximum*, and *Mean* are the minimum, maximum, and mean of the residuals in meters, *RMSE* is the root mean square error (m), and *ME* is the mean error (m).

	Count	Minimum	Maximum	Mean	RMSE	ME
DEM DGPS	147	−0.60	0.48	−0.01	0.13	0.08
DEM 10	147	−2.27	6.88	0.37	1.08	0.72
DEM 25	147	−5.24	5.02	−2.01	2.46	2.12

Table 3. Gravity response caused by a water mass change of 1 m (density 1000 kg/m³) calculated for the different DEMs as a function of vertical layer distance and radius of influence R.

DEM	Vertical layer distance (m)	Response (μGal) R: 50 m	Response (μGal) R: 500 m	Response (μGal) R: 2000 m	Response (μGal) R:10000 m
DEM DGPS	1	42.04	47.79	51.08	52.46
	20	27.91	44.59	48.91	50.60
DEM 10	1	42.72	48.65	51.94	53.35
	20	27.87	44.68	49.00	50.69
DEM 25	1	33.96	38.26	41.45	42.86
	20	28.46	44.04	48.27	49.96

References

Abe, M., S. Takemoto, Y. Fukuda, T. Higashi, Y. Imanishi, S. Iwano, S. Ogasawara, Y. Kobayashi, S. Dwipa, and D. S. Kusuma, 2006, Hydrological effects on the superconducting gravimeter observation in Bandung: Journal of Geodynamics, **41**, 288-295.

Baumgartner, A., and H.-J. Liebscher, 1996, Lehrbuch der Hydrologie: Allgemeine Hydrologie — Quantitative Hydrologie: Borntraeger.

BKG, 2004, ATKIS® DGM-D: Vermessungsverwaltungen der Länder und BKG (Bundesamt für Kartographie und Geodäsie).

BKG, 2005, ATKIS® Basis-DLM: Vermessungsverwaltungen der Länder und BKG (Bundesamt für Kartographie und Geodäsie).

Bonatz, M., 1967, Der Gravitationseinfluss der Bodenfeuchte: Zeitschrift für Vermessungswesen, **92**, 135-139.

Bonatz, M., and D. Sperling, 1995, Gravitation effects at the Vianden storage power station: Presented at the Cahiers du Centre Européen de Géodynamique et de Séismologie.

Bower, D. R., and N. Courtier, 1998, Precipitation effects on gravity measurements at the Canadian Absolute Gravity Site: Physics of the Earth and Planetary Interiors, **106**, 353-369.

Crossley, D., S. Xu, and T. v. Dam, 1998, Comprehensive Analysis of 2 years of SG Data from Table Mountain, Colorado: Presented at the 13th International Symposium on Earth Tides, Observatoire Royal de Belgique.

Damiata, B. N., and T.-C. Lee, 2006, Simulated gravitational response to hydraulic testing of unconfined aquifers: Journal of Hydrology, **318**, 348-359.

de Jong, B., and G. Ros, 2004, The effect of water storage changes on gravity near Westerbork: Thesis, University of Wageningen.

Dyck, S., and G. Peschke, 1983, Grundlagen der Hydrologie: Ernst.

Ehrismann, W., 1973, Ein allgemeines Verfahren zur digitalen Berechnung der Schwerewirkung von Modellkörpern: Zeitschrift für Geophysik, **39**, 131-166.

Farrel, W. E., 1972, Deformation of the earth by surface loads: Reviews of Geophysics, **10**, 761-797.

Forsberg, R., 1984, A study of terrain reductions, density anomalies and geophysical inversion methods in gravity field modelling: Department of Geodetic Science and Surveying, Ohio State University.

Goodkind, J. M., 1999, The superconducting gravimeter: Review of Scientific Instruments, **70**, 4131-4152.

Götze, H.-J., 1976, Ein numerisches Verfahren zur Berechnung der gravimetrischen und magnetischen Feldgrößen für dreidimensionale Modellkörper: Doctoral dissertation, TU Clausthal.

Güntner, A., J. Stuck, S. Werth, P. Döll, K. Verzano, and B. Merz, 2007, A global analysis of temporal and spatial variations in continental water storage: Water Resources Research, **43**, W05416 doi:10.1029/2006WR005247, accessed March 10, 2008.

Harnisch, G., and M. Harnisch, 2006b, Hydrological influences in long gravimetric data series: Journal of Geodynamics, **41**, 276-287.

Harnisch, G., M. Harnisch, and R. Falk, 2006a, Hydrological influences on the gravity variations recorded at Bad Homburg: Bulletin d'Information des Marées Terrestres, **142**, 11331-11342.

Harnisch, M., and G. Harnisch, 2002, Seasonal variations of hydrological influences on gravity measurements at Wettzell: Bulletin d'Information des Marées Terrestres, **137**, 10849-10861.

Hasan, S., P. A. Troch, J. Boll, and C. Kroner, 2006, Modeling the hydrological effect in local gravity at Moxa, Germany: Journal of Hydrometeorology, **7**, 346-354.

He, X., 2007, Using hydrological models to simulate time-lapse gravity changes at Danish permanent gravity stations: Master's thesis, Technical University of Denmark.

Heck, B., and K. Seitz, 2007, A comparison of the tesseroid, prism and point-mass approaches for mass reductions in gravity field modelling: Journal of Geodesy, **81**, 121-136.

Hinderer, J., C. de Linage, and J. P. Boy, 2006, How to validate satellite-derived gravity observations with gravimeters at the ground?: Bulletin d'Information des Marées Terrestres, **142**, 11433-11442.

Hokkanen, T., K. Korhonen, and H. Virtanen, 2006, Hydrogeological effects on superconducting gravimeter measurements at Metsähovi in Finland: Journal of Environmental & Engineering Geophysics, **11**, 261-267.

Imanishi, Y., K. Kokubo, and H. Tatehata, 2006, Effect of underground water on gravity observation at Matsushiro, Japan: Journal of Geodynamics, **41**, 221-226.

Klügel, T., and H. Wziontek, 2007, Atmosphärische Schwereeffekte: Modell-Zeitreihen aus 3-dimensionalen Wettermodellen: Presented at the Jahrestagung der DGG.

Kolbenheyer, T., 1967, Die Schwerewirkungen eines geraden Prismas mit rechtwinkligem Querschnitt: Studia Geophysica & Geodætica, **11**, 262-270.

Kroner, C., T. Jahr, M. Naujoks, and A. Weise, 2007, Hydrological signals in gravity — Foe or friend?, *in* C. Rizos, ed., Dynamic planet — Monitoring and understanding a dynamic planet with geodetic and oceanographic tools: Springer.

Leirião, S., 2007, Hydrological model calibration using ground based and spaceborne time-lapse gravity surveys: Master's thesis, Technical University of Denmark.

Llubes, M., N. Florsch, J. Hinderer, L. Longuevergne, and M. Amalvict, 2004, Local hydrology, the Global Geodynamics Project and CHAMP/GRACE perspective: Some case studies: Journal of Geodynamics, **38**, 355-374.

LVG, 2007, DGM 5 aus Höhenlinien: Landesamt für Vermessung und Geoinformation Bayern.

MacMillan, W. D., 1958, Theoretical mechanics: The theory of the potential: Dover Publications, Inc.

Mader, K., 1951, Das Newtonsche Raumpotential prismatischer Körper und seine Ableitungen bis zur dritten Ordnung: Österreichische Zeitschrift für Vermessungswesen, Sonderheft **11**.

Mäkinen, J., and S. Tattari, 1988, Soil moisture and groundwater: Two sources of gravity variations: Bulletin d'Information des Marées Terrestres, **64**, 103-110.

Meurers, B., M. V. Camp, and T. Petermans, 2006, Correcting superconducting gravity time-series using rainfall modelling at the Vienna and Membach stations and application to earth tide analysis: Journal of Geodesy, **81**, 703-712.

Mufti, I. R., 1973, Rapid determination of cube's gravity field: Geophysical Prospecting, **21**, 724-735.

Nagy, D., 1966, The gravitational attraction of a right rectangular prism: Geophysics, **31**, 362-371.

Naujoks, M., C. Kroner, T. Jahr, P. Krause, and A. Weise, 2007a, Gravimetric 3D modelling and observation of time-dependent gravity variations to improve small-scale hydrological modelling: Presented at the IUGG XXIV General Assembly.

Naujoks, M., A. Weise, C. Kroner, and T. Jahr, 2007b, Detection of small hydrological variations in gravity by repeated observations with relative gravimeters: Journal of Geodesy, doi:10.1007/s00190-007-0202-9, accessed March 6, 2008.

Neumeyer, J., J. Hagedoorn, J. Leitloff, and T. Schmidt, 2004, Gravity reduction with three-dimensional atmospheric pressure data for precise ground gravity measurements: Journal of Geodynamics, **38**, 437-450.

Peter, G., F. Klopping, and K. A. Berstis, 1995, Observing and modeling gravity changes caused by soil moisture and groundwater table variations with superconducting gravimeters in Richmond, Florida, USA: Presented at the Cahiers du Centre Européen de Géodynamique et de Séismologie.

Petrahn, G., 2000, Grundlagen der Vermessungstechnik: Cornelsen.

Petrović, S., 1996, Determination of the potential of homogeneous polyhedral bodies using line integrals: Journal of Geodesy, **71**, 44-52.

Petrović, S., and P. Skiba, 2001, Polyhedra in the gravity field modeling: Proceedings of the International Workshop on Perspectives of Geodesy in South-East Europe, Mitteilungen der geodätischen Institute der Technischen Univerasität Graz, **89**, 115-128.

Pohanka, V., 1988, Optimum expression for computation of the gravity field of a homogeneous polyhedral body: Geophysical Prospecting, **36**, 733-751.

Prutkin, I., and R. Klees, 2007, Environmental effects in time-series of gravity measurements at the Astrometric-Geodetic Observatorium Westerbork (Netherlands), *in* C. Rizos, ed., Dynamic planet — Monitoring and understanding a dynamic planet with geodetic and oceanographic tools: Springer.

Rausenberger, O., 1888, Lehrbuch der analytischen Mechanik I.: B.G. Teubner.

Rubbert, T. K., 2008, Hydrogeologische Modellbildung eines kombinierten porös-geklüfteten Grundwasserleitersystems des Bayerischen Waldes: Doctoral dissertation, Ruhr-Universität.

Schulz, J., 2002, Die Ökozonen der Erde: Eugen Ulmer.

Van Camp, M., M. Vanclooster, O. Crommen, T. Petermans, K. Verbeeck, B. Meurers, T. v. Dam, and A. Dassargues, 2006, Hydrogeological investigations at the Membach station, Belgium, and application to correct long periodic gravity variations: Journal of Geophysical Research, **111,** B10403 doi:10.1029/2006JB004405, accessed November 26, 2007.

Virtanen, H., 1999, On the Observed Hydrological Environmental Effects on Gravity at the Metsähovi Station, Finland: Presented at the Cahiers du Centre Européen de Géodynamique et de Séismologie.

Virtanen, H., 2001, Hydrological studies at the gravity station Metsähovi in Finland: Journal of the Geodetic Society of Japan, **47**, 328-333.

List of Figures

Figure 1. The superconducting gravimeter of the Geodetic Observatory Wettzell (coordinates in Gauss Krüger, zone 4, Bessel Ellipsoid: x = 4564212 m, y = 5445520 m). (1) Concrete foundation with the size of 1.4 × 1.4 × 1.2 m (width × depth × height), (2) Soil, (3) Baseplate (height = 613.80 m), (4) Sensor 1 (height = 614.2 m), (5) Sensor 2 (height = 614.4 m) and (6) SG building.

Figure 2. Spatial extent of the nested model domains (black squares) and the topography (DEM DGPS) for (a) the total model area and for (b) the vicinity of the SG. The model elementary cell size (Δxy) varies with the domain radius R (half the side length of the domain square). Dark gray dots are the measured DGPS points, and light gray lines are height contours – (a) 100-m interval; (b) 1-m interval.

Figure 3. Spatial distribution of the gravity response resulting from a water mass change of 1 m (density change 1000 kg/m³) at the terrain surface (vertical layer distance to the sensor = 1 m), calculated for each element (with varying element size Δxy) for the whole nested model.

Figure 4. Spatial distribution of the gravity response resulting from a water mass change of 1 m (density change 1000 kg/m³) in a layer with a vertical distance of (a) 1 m, (b) 10 m, and (c) 20 m to the sensor, for an area of 100 × 100 m around the SG.

Figure 5. Gravity response resulting from a water mass change of 1 m (density change 1000 kg/m³) as a function of the radius of influence and the vertical layer distance of the mass change.

Figure 6. Gravity response for water storage change (WSC) in each storage compartment (snow, soil moisture, vadose zone soil moisture and groundwater). The storage compartments are progressively saturated with water from bottom-up. Dashed lines represent the maximum range of the gravity response for R=50 m taking the uncertainty of relevant parameters into account.

Figure 7. (a) The modeled gravity effect of the different water storage components from November 29, 2000 to December 31, 2006 and (b) the gravity residuals measured by the SG and the total sum of the modeled gravity signal with a mean specific storage of 5 % for groundwater. Dotted blue lines are the total sum of the modeled gravity for maximum (11 %) and minimum (1 %) specific storage.