Atmospheric effects on the Earth gravity field featured by TU Vienna





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Abstract

Satellite missions like GRACE (Gravity Recovery and Climate Experiment) and GOCE (Gravity field and steady-state Ocean Circulation Explorer) which explore the Earth gravity field observe the instantaneous distribution of mass in the Earth, including all solid, liquid and gaseous components. Due to the fluctuation of those masses at various temporal and spatial scales, a long observation period does not guarantee that the introduced variations in the gravity field are cancelled out. Therefore, to avoid aliasing effects, the mass variations have to be modeled and corrected with respect to the mean state.Within project GGOS Atmosphere, funded by the Austrian Science Fund (FWF) at the Institute of Geodesy and Geophysics (IGG) of the Vienna University of Technology, different methods for the determination of Atmospheric Gravity field Coefficients (AGC) are evaluated. Results indicate that for a proper modelling the vertical structure of the atmosphere has to be taken into account, as already applied for GRACE data processing. Further, atmosphere loading adds a significant signal to the gravity change which has to be considered, in particular at longer wavelengths. The choice of different data structures of the ECMWF (European Centre for Medium-range Weather Forecasts), i.e. model or pressure level data, does not have a significant impact on the final AGC. All findings confirm the data processing strategy of the GRACE Science Data System([4] Flechtner, 2007), providing the operational GRACE AOD1B (level 1B atmosphere and ocean de-aliasing) product.

Keywords: Gravity field, atmosphere, GRACE, de-aliasing

Kurzfassung

Satelliten-Missionen wie GRACE (Gravity Recovery and Climate Experiment) und GOCE (Gravity Field and steady-state Ocean Circulation Explorer), die das Erdschwerefeld erkunden, beobachten die momentane Verteilung der Massen im System Erde, einschließlich aller festen, flüssigen und gasförmigen Bestandteile. Aufgrund der Fluktuation dieser Massen auf verschiedenen räumlichen und zeitlichen Skalen garantiert eine lange Beobachtungszeit nicht, dass die durch sie verursachten Variationendes Schwerefeldeseliminiert werden. Um so genannte Aliasing-Effekte zu vermeiden, muss deshalb der bekannte Teil der Massenvariationen modelliert und bezüglich eines mittleren Zustandes korrigiert werden. Innerhalb des Projekts„GGOS Atmosphäre“, finanziert vom Österreichischen Wissenschaftsfonds (FWF) am Institut für Geodäsie und Geophysik (IGG) der TU Wien, werden verschiedene Methoden zur Bestimmung der atmosphärischen Schwerefeldfeldkoeffizienten (AGC) ausgewertet. Die Ergebnisse zeigen, dass für eine adäquate Modellierung die vertikale Struktur der Atmosphäre zu berücksichtigen ist. Außerdem hat die Auflast der Atmosphäre einen signifikanten Einfluss auf die Schwerkraftvariation und ist somit ebenfalls zu berücksichtigen. Die Wahl unterschiedlicher Datenstrukturen des ECMWF (European Centre for Medium-range Weather Forecasts), nämlich„model“ oder „pressure level“ Daten, hat keinen entscheidenden Einfluss auf die AGC. Alle Ergebnisse bestätigen die Strategie zur Datenverarbeitung des GRACE Science Data Systems ([4] Flechtner, 2007), welches das GRACE AOD1B (Stufe 1B Atmosphäre und Ozean de-Aliasing) Produkt bereitstellt.

Schlüsselwörter: Schwerefeld, Atmosphäre, GRACE, de-aliasing

1. Introduction

Exploring the Earth gravity field requires the removal of short term (sub-daily) mass variations in the system Earth, including all solid, liquid and atmospheric particles. Due to the fluctuation of those masses at various temporal and spatial scales(like high and low atmospheric pressure systems) as well as due to a strong dependency on the sampling rate of the ground track of the satellite, a long observation time does not guarantee that the introduced variations in the gravity field are cancelled out by the mean operator. De-aliasing then denotes incorporating such instantaneous variations in the atmospheric masses with respect to a static mean state of the atmosphere, either during the preprocessing of observations or during the estimation procedure of the gravity field solution. The same holds for all other mass variation effects inside the system Earth; only that within the atmosphere also the centre of mass of the atmospheric column is varying, which interferes again on the satellite observations ([5] Gruber et.al,2009).

To eliminate the aliasing signals the determination of accurate Atmospheric Gravity field Coefficients (AGC) is indispensable. For the determination of AGC it has become state of the art to use high resolution Numerical Weather Models (NWM),which take into account the three-dimensional distribution of the atmospheric mass. By subtracting the gravity spherical harmonics of the instantaneous atmosphere from the ones of the mean atmospheric field, the residual gravity spherical harmonic series are obtained. These describe the deviation of the actual gravity field from the mean gravity field due to atmospheric mass variations.

In Section2 we contrast the formulation of the AGC under different hypotheses, i.e. the thin layer assumption and the 3D approach.Section3 is devoted to the different data structures, the preprocessing of the NWM data, and the strategy used for the computation of the AGC. The computational results are given in Section4.

2. From mass to gravity

The atmosphere is nearly in a hydrostatic equilibrium, which means that the change in atmospheric pressure on the surface is proportional to the change of mass in the corresponding atmospheric column, including variations in water vapour mass as well as in the dry air mass. ρ describes the density along the column which can be expressed in terms of surface load σ ([2] Boy et al. 2002, [4] Flechtner, 2007) and which is linked directly to the surface pressure variation Δp.

$$\Delta p = g_0 \int_{r_s}^{\infty} \Delta\rho dr = g_0 \Delta\sigma \,, \tag{1}$$

$$\Delta\sigma = \frac{\Delta p}{g_0} \,, \tag{2}$$

where g_0 is the mean gravity acceleration at the Earth surface, Δp the pressure variation and r_s denotes the Earth surface.

The atmosphere affects the Earth gravity field in two different ways: a direct attraction of the atmospheric masses acting on the orbiting satellite and a much smaller indirect effect introduced by the deformation of the Earth's surface due to elastic loading. Both effects are always evaluated with respect to a mean atmosphere model. This approach is described in detail by [8] Torge (1989).

This section is exclusively devoted to the direct effect, whereas Section 4.2will deal with the indirect effect. A mathematical description of the gravitational potential can be given in terms of a spherical harmonic expansion (see [8] Torge, 1989):

$$V = \frac{GM}{r}\sum_{n=0}^{\infty}\sum_{m=0}^{n}\left(\frac{a}{r}\right)^{n} P_{nm}(\cos\theta)\left(C_{nm}\cos m\lambda + S_{nm}\sin m\lambda\right), \tag{3}$$

$$\begin{Bmatrix} C_{nm} \\ S_{nm} \end{Bmatrix} = \frac{1}{(2n+1)Ma^n} \cdot \iiint_{Earth} r^n P_{nm}(\cos\theta) \begin{Bmatrix} \cos m\lambda \\ \sin m\lambda \end{Bmatrix} dM \,, \tag{4}$$

where $dM = \rho r^2 dr \sin\theta d\theta d\lambda$. (5)

GM is the geocentric gravitational constant multiplied with the Earth's mass (solid Earth + oceans + atmosphere), a denotes the radius of a spherical Earth, r is the distance to the centre of mass of the Earth, θ and λ are co-latitude and longitude, C_{nm} and S_{nm} are dimensionless coefficients and P_{mn} are the fully normalized associated Legendre functions, both depending on degree n and order m.

Due to mass redistribution in the atmosphere the potential V changes with time. This time-dependency of atmospheric density $\Delta\rho$ can be represented in terms of time-dependent ΔC_{nm} and ΔS_{nm} coefficients, taking into account Equations (4) and (5), as follows:

$$\begin{Bmatrix} \Delta C_{nm} \\ \Delta S_{nm} \end{Bmatrix} = \frac{1}{(2n+1)Ma^n} \cdot \iint_{Earth} \left[\int_{r_s}^{\infty} \Delta\rho r^{n+2} dr\right] P_{nm}(\cos\theta) \begin{Bmatrix} \cos m\lambda \\ \sin m\lambda \end{Bmatrix} \sin\theta d\theta d\lambda \tag{6}$$

2.1 Thin layer approximation

In the simplest approach the vertical extent of the atmosphere is neglected and all the atmospheric masses are concentrated in a thin layer at the Earth surface. This can be done under the assumption that most of the mass changes occur in the lower 10km of the atmosphere and act as variable loading effects on the solid Earth's surface ([1] Boy et al.,2005).

Surface loads are defined as mass per surface element; therefore the density change in the atmosphere can be expressed in terms of surface load as follows:

$$\begin{Bmatrix} \Delta C_{nm} \\ \Delta S_{nm} \end{Bmatrix} = \frac{a^2}{(2n+1)M} \cdot \iint_{Earth} \Delta\sigma P_{nm}(\cos\theta) \begin{Bmatrix} \cos m\lambda \\ \sin m\lambda \end{Bmatrix} dS , \tag{7}$$

considering that the mass element.

$$dM = \rho r^2 dr \sin\theta d\theta d\lambda = \\ = r^2 \sigma \sin\theta d\theta d\lambda = r^2 \sigma dS .$$

Following the definition of the surface load σ in Equation (2), the surface pressure p_s can be introduced, whereas a mean pressure field $\underline{p_s}$, representing a static mean state of the atmosphere, has to be subtracted to obtain the mass variation:

$$\begin{Bmatrix} \Delta C_{nm} \\ \Delta S_{nm} \end{Bmatrix} = \frac{a^2}{(2n+1)Mg_0} \cdot \iint_{Earth} \left(p_s - \underline{p_s}\right) P_{nm}(\cos\theta) \begin{Bmatrix} \cos m\lambda \\ \sin m\lambda \end{Bmatrix} dS \tag{8}$$

2.2 Vertical integration of the atmospheric column

As mentioned in the introduction, also the change of the centre of mass of the atmospheric column has an impact on the orbiting satellite, not only the mass change itself. This variation of the centre of mass is not addressed in the thin layer approximation but has to be taken into account for satellite gravity missions such as GRACE (Gravity Recovery and Climate Experiment) ([4] Flechtner, 2007; [7] Swenson and Wahr, 2002; [11] Velicogna et al., 2001).

This deficiency can be overcome by considering the whole vertical structure of the atmosphere by performing a vertical integration of the atmospheric masses. To do so, Numerical Weather Models(NWM) which describethe vertical structure by introducing various numbers of pressure or model levels are needed. The structure and the processing of these data will be explained in Section 3.

To formulate the vertical integration (VI) we start from the basic Equations(3) and (4), introducing the volume element from Equation(5) (for details see [4] Flechtner, 2007; [9] Zenner et al., 2010; [10] Zenner et al., 2011).

$$\begin{Bmatrix} C_{nm} \\ S_{nm} \end{Bmatrix} = -\frac{1}{(2n+1)Ma^n} \cdot \iint_{Earth} \left[\int_{r_s}^{\infty} r^{n+2} \rho dr \right] P_{nm}(\cos\theta) \begin{Bmatrix} \cos m\lambda \\ \sin m\lambda \end{Bmatrix} \sin\theta d\theta d\lambda \tag{9}$$

Adopting the hydrostatic equation $\rho dr = -\frac{dp}{g_r}$, where gr is the gravity acceleration at each level, we get:

$$\begin{Bmatrix} C_{nm} \\ S_{nm} \end{Bmatrix} = -\frac{1}{(2n+1)Ma^n} \cdot \iint_{Earth} \left[\int_{Ps}^{0} \frac{r^{n+2}}{g_r} dp \right] P_{nm}(\cos\theta) \begin{Bmatrix} \cos m\lambda \\ \sin m\lambda \end{Bmatrix} \sin\theta d\theta d\lambda \tag{10}$$

Again, to analyze gravity field variations caused by atmospheric effects, a quantity $\underline{p_{VI}}$ representing the mean state of the atmosphere, has to be subtracted from the inner integral, leading to:

$$\begin{Bmatrix} \Delta C_{nm} \\ \Delta S_{nm} \end{Bmatrix} = -\frac{1}{(2n+1)Ma^{n+2}g_0} \cdot \iint_{Earth} \left(\left[\int_{Ps}^{0} r^{n+4} dp \right] - \underline{p_{VI}} \right) P_{nm}(\cos\theta) \begin{Bmatrix} \cos m\lambda \\ \sin m\lambda \end{Bmatrix} \sin\theta d\theta d\lambda \tag{11}$$

3 Data and processing

3.1 Numerical Weather Models

For this work NWM data from the European Centre for Medium-range Weather Forecasts (ECMWF) are used. Generally, the results of the ECMWF analysis are provided on individual layers, realized as model or pressure level data. The model level data presently consist of 91 model levels. The concept of model levelsaddresses the problem of discontinuities in the atmosphere, for example mountains, by creating atmospheric levels that follow the contours of the Earth's surface in the lower and mid-troposphere, the so-called orography. In high altitude the effect of the orography diminishes until the layers in the upper atmosphere becomeparallel to layers of constant pressure.

From the model level data the so-called pressure levels are retrieved, where the vertical discretization is implemented through 25 levels instead of 91, following continuous surfaces of equal pressure from 1000hPa to 1hPa, which can also lie underneath the topography. At each level, among other parameters, the temperature, the specific humidity, and the geopotential height are available. For this paper, pressure level data on global equidistant grids with a horizontal resolution of 1°×1° and a temporal resolution of 6 hours (00, 06, 12, 18UTC)were used.

3.2 Pre-processing: from geopotential height to the topography

As can be seen in Equation (10), not the geopotential height of each level is needed but the geocentric radius, which is not delivered by ECMWF. Equations and approximations for the usage of the geopotential height can be found in [4] Flechtner (2007). Otherwise, the radii of the individual levels as well as the gravitational acceleration at each level have to be calculated in the pre-processing.

At TU Vienna, the data from the ECMWF are downloaded daily as rectangular, three-dimensional grids in the grib-format, containing the geopotential Z, the specific humidity Q, and the temperature T at discrete points on each pressure level and at each epoch (00, 06, 12, 18 UTC). Further meta-data like time and date, spatial resolution and number of nodes are included.

In the pre–processing the following steps are performed:

1. The geographical co-latitude θ given by ECMWF is transformed to the WGS84 ellipsoid by setting it equal to the geocentric latitude ψ.
2. In order to get the longitude and latitude dependent gravity acceleration at each level, it is necessary to introduce a gravity model. We used the fully normalized degree 2 coefficients and the corresponding gravity acceleration of the tide-free EGM96 model. Further, the geoid undulation is needed to retrieve exact ellipsoidal heights. At this point the EGM96 geoid as given by the IGFS (International Gravity Field Service) on a 1°x1° ellipsoidal grid is used. The differences to geocentric latitudes are again neglected. Finally, the geocentric radii, the corresponding gravity acceleration and the ellipsoidal height of all layer grid points are computed. Additionally, the density and the virtual temperature T_v ([4] Flechtner, 2007) are calculated and stored.
3. The ECMWFmodel level data are not based on topography but on orography, i.e., an envelope of the actual topography, with the consequence that smaller details or rapid height changes are not represented. To overcome this deficiency we reduce all the parameters retrieved during step 2 to the topography of the ETOPO5 model (http://www.ngdc.noaa.gov/mgg/global/etopo5.HTML).
4. For all layers and all nodes block-mean values are calculated to be consistent with the theory of spherical harmonic expansions.

Although we introduce longitude and latitude dependent radii and gravity acceleration for the Earth surface instead of the constant a and g_0 in Equations (8) and (10), investigations have shown, that at the current accuracy levels of the GRACE processing, this alteration has no significant influence.

3.3 Calculation strategy for the Atmospheric Gravity Coefficients (AGC)

For both approaches, the thin layer approach as well as the vertical integration, a reference (pressure) field is needed. In the first case as a 2D field at the surface, and for VI approach it has to represent the three-dimensional structure of the atmosphere. For the thin layer approach we use the Global Reference Pressure model GRP developed at our institute (Schuh et al.,2010). It is a 2D surface pressure field computed from the atmospheric data of ECMWF ERA-40 and referenced to the ETOPO5 topography. Given its nature, GRP cannot be used for VI, where a 3D model corresponding to the calculation model has to be used. For this purpose, Equation(10) was evaluated for the years 2008 and 2009 and a mean was formed. Consequently, this mean field is not a surface pressure field, but consists of mean Atmospheric Gravity Coefficients (AGC).

3.3.1 Thin layer approach

Starting from the block-mean value obtained in the pre-processing of the ECMWF data, the mean pressure field GRP is subtracted from the actual surface pressure to get the pressure variation. Those differential values are then entered in Equation (8) and integrated numerically over the entire Earth's surface. The obtained integral value is then transformed into the actual potential by multiplication with the expression in front of the surface integral. This procedure is repeated for each degree and order.

3.3.2 Vertical Integration (VI)

For the VI approach we evaluate Equation (10) for the actual epoch. The inner integral is computed first, starting from the highest level, down to the topography. This value is then entered in the same procedure as used for the thin layer approach. Unlike the thin layer approach, we do not calculate the difference of the 3D-pressure beforehand but afterwards by subtracting the coefficients of the mean 3D field from the ones just calculated for the actual epoch.

All the coefficients are derived up to degree and order 100 and stored as text file in (n, m, C_{nm}, S_{nm}) format on our central server (http://ggosatm.hg.tuwien.ac.at/GRAVITY/). The GRP model can be downloaded from there as well.

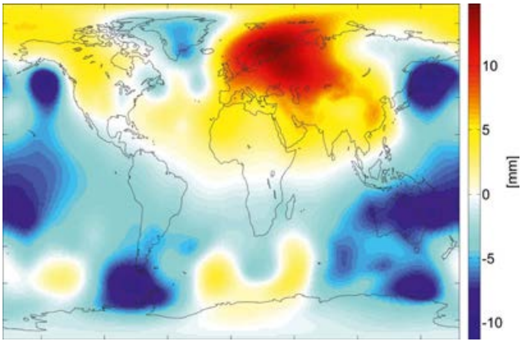

Fig. 1a: *Geoid height variation (VI approach) with respect to the mean field (over 2008 and 2009) in mm on January 1st, 2008, 00 UTC (min: -11.35 mm, max: 14.81 mm, rms: 2.9 mm).*

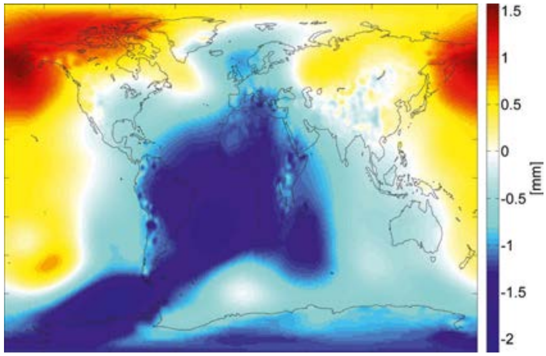

Fig. 1b: *Difference between the AOD1B "atm" product and our (TU Vienna) VI approach in geoid height (min: –2.25 mm, max: 1.77 mm, rms: 0.7 mm).*

4. Results

The real impact of aliasing effects and other miss-modelling of the atmosphere cannot be estimated straight forwardly. Therefore, we rely on comparisons of degree standard deviations in geoid height and global plots of the geoid heights.

All the results presented here base on the 6-hourly pressure information of the year 2008. (Mind that the mean field for the VI approach was determined for 2008 and 2009.) As an example the first epoch (00 UTC) of January 1st 2008 is selected. Figure 1a (left plot) depicts the geoid height variation following the VI approach and Fig. 1.b (right plot)the difference between the official AOD1B "atm" product and our (TU Vienna) VI approach is shown, also expressed in geoid height.

Both solutions, AOD1B and TU Vienna, show a good agreement, also in terms of degree standard deviation (Figure 2) or distinct coefficients (Figure 3). The differences are most prominent at long wavelengths and can be attributed to the different definition of the static mean field of the atmosphere (AOD1B: mean over 2001+2002, VI approach by TU Vienna: mean over 2008+2009) and to the fact that in the VI approach by TU Vienna the S1 tide is still included.

To evaluate the significance of the vertical structure of the atmospheric column, the spherical harmonic series resulting from the thin layer approach and the ones of the VI approach are compared. In Figure 2 the degree standard deviations of the coefficients for the year 2008 up to degree 100 are compared to the AOD1B co-

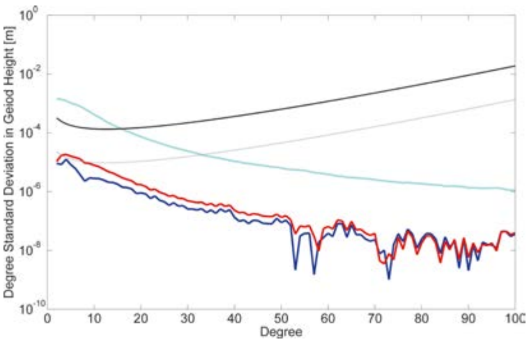

Fig. 2: *Degree standard deviation in terms of geoid height for the year 2008, in cyan for the AOD1B product for the atmosphere, in blue the difference of the VI approach by TU Vienna with respect to the AOD1B product, in red the corresponding difference of our thin layer approach w.r.t. AOD1B. The black line marks the actual error level of GRACE, the grey one the theoretical error as obtained by pre-launch simulations.*

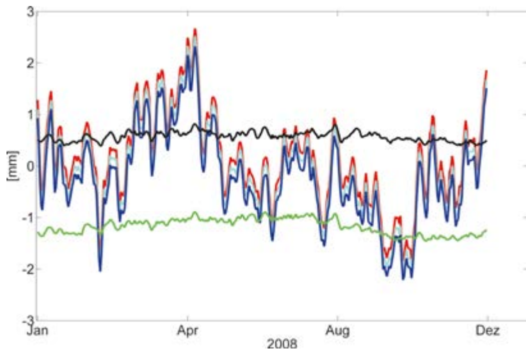

Fig. 3: *Time variation of the C_{20} coefficient in geoid height for the year 2008, in blue for the VI approach by TU Vienna, in red the thin layer approach, in cyan for the AOD1B product. The difference between the VI approach by TU Vienna and AOD1B is shown in black, the difference between the thin layer approach and AOD1B in green, both differences multiplied by a factor of 10.*

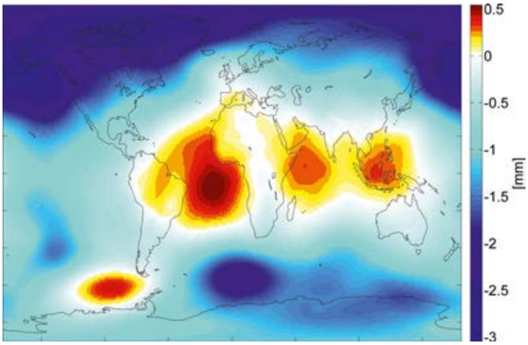

Fig. 4a: Difference of the geoid height variation between the VI approach and the thin layer approach for January 1st 2008, 00 UTC (min: -3.05 mm, max: 0.54 mm, rms: 1.2mm).

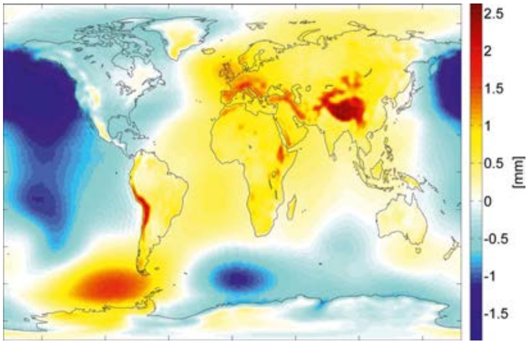

Fig. 4b: Difference of the reference fields for the VI and the thin layer approach, expressed in geoid height (min: -2.62 mm, max: 1.86 mm, rms: 0.52 mm).

efficients. Figure 3 exemplarily shows the geoid height variability for the C_{20} coefficients. The results indicate that at the current error level the differences between the two approaches by TU Vienna and the official product are negligible, thus also confirming the approach by the GRACE science team.

In a second step, the resulting potential fields obtained from the two different approaches (thin layer and VI, both from TU Vienna) for January 1st, 2008, 0 UTC and the two corresponding mean fields are compared, always in terms of geoid height. Figure 4a on the left shows the difference between the thin layer approach and the vertical integration approach, and for both approaches the respective mean fields are subtracted. Figure 4b shows the discrepancy between the two mean fields(average per latitude band was removed).In Figure 5 the absolute values (no mean field subtracted) for the two methods at the actual epoch are plotted. In order not to have a dominating effect of the topography, a land-sea mask was applied. If the different approaches (thin layer vs. VI) would be the cause of the differences in Figure 4a, similar structures should appear also in the discrepancy of the total atmosphere in Figure 5;however this is not the case. Therefore those signals are introduced somewhere else, probably due to the different definition of the mean fields mentioned in Section3.3.Obviously,besides topographical signals due to the different reference height, i.e. surface and centre of mass, also some signals coming from the atmosphere are still present in Figure 4b, showing some correlation with the artefacts in Figure 4a. This leads to the conclusion that those signals are introduced and then propagated to the final AGC.

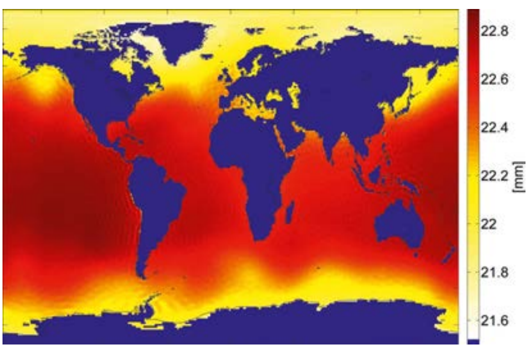

Fig. 5: Difference of the total atmosphere between VI and the thin layer approach, expressed in geoid height for January 1st 2008, 0 UTC (min: 21.34 mm, max22.89 mm, rms: 22.44 mm).

This discrepancy can be overcome, if a consistent mean pressure field would be calculated (from Equation(10)). However, due to the enormous computational expense to process the full ERA-40 dataset in 3D, this task was abandoned for now. Although the effect is too small to have a significant influence on the resulting ACG for the actual GRACE mission, improved versions of reprocessed gravity solutions might demand to take this factor into account.

4.2 Loading

In all the calculations up to now the indirect effect, i.e. the elastic deformation of the solid Earth due to atmospheric loading was not considered. This effect is counteracting the direct effect due to the deformation towards the geocentre. In general, for small deformations the additional change in the potential ΔV depends linearly on the potential (Equation(3)), following [3] Farrell (1972):

$$\Delta V_n^{ind} = k_n \Delta V \,, \tag{12}$$

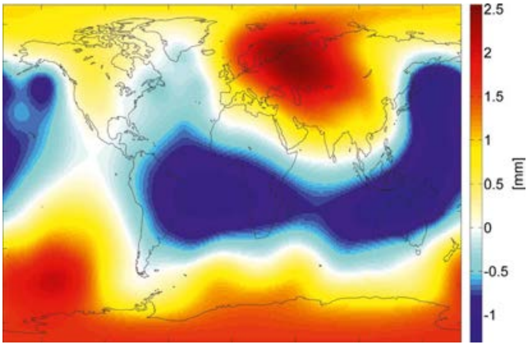

Fig. 6: Difference of the geoid height variation for the thin layer approach between the variants with and without loading for January 1st, 2008, 0 UTC;(min:-1.32 mm, max: 2:45 mm, rms: 0.78 mm).

$$\Delta V_n^{tot} = \Delta V + k_n \Delta V = (1 + k_n)\Delta V \ . \qquad (13)$$

kn denote the degree dependent Load Love numbers and represent the deformational behaviour based on the rheology of the Earth. For our processing, we use Load Love numbers defined in the centre of mass framecalculated by Pascal Gegout,provided by Jean-Paul Boy, and downloaded from http://astrogeo.org/agra/Load_Love2_CM.dat.

Figure 6 shows the difference between a solution without considering loading and one which includes loading, both for the thin layer approximation. As expected only differences at a big spatial scale appear since Earth's elastic surface deformation due to mass redistribution is sensitive to large scale pressure variations with wavelengths greater than 2000 km, corresponding to n<10 (Boy et al. 2002).This result is confirmed by the degree standard deviation expressed in geoid height calculated for the year 2008 (Figure 7).

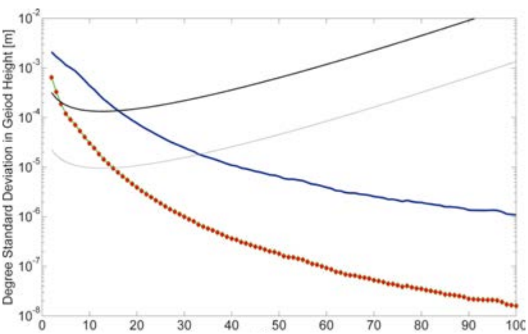

Fig. 7: Degree standard deviation in terms of geoid height for the year 2008, in blue for the VI approach with loading, in green the corresponding difference of the VI approach without loading, in red the corresponding difference of the thin layer approach without loading. The black line marks the actual error level of GRACE, the grey one the theoretical error as obtained by pre-launch simulations.

Given the fact that the differences up to degree 4 lie above the actual error level and up to degree 15 above the predicted error level, the indirect effect has to be accounted for, as it is of course done for the AOD1B product. The same conclusion is drawn looking at the difference between with and without loading in terms of geoid height variability for low degrees (Figure 8), considering the aimed precision of GRACE to be a few micrometers for degrees 3 to 5.

4.3 Pressure and modellevel data

As mentioned before, the ECMWF data can be downloaded as pressure or model level data. The biggest difference between those two representations is the method of discretisation of the vertical structure of the atmosphere. Whereas the model level data reach up to approximately 80 km, the pressure level search up to a height of about 46 km. The lowest model level, i.e. the one nearest to the surface, follows the orography used by the ECMWF; the lowest pressure level is at 1000 hPa. In Figure 9a (left plot) the difference between topography and orography is shown; the majority of the differences appear in mountainous regions like the Himalaya or the Andes, but the most prominent anomalies (more than 1 km) can be found in the Antarctica.

To determine the influence of the data structure on the AGC results, the difference between the VI solutions computed with pressure level data and model level data was calculated and plotted in Figure 9b in terms of geoid height. Small non-zero features over the continents appear, most prominent in the Himalaya region.

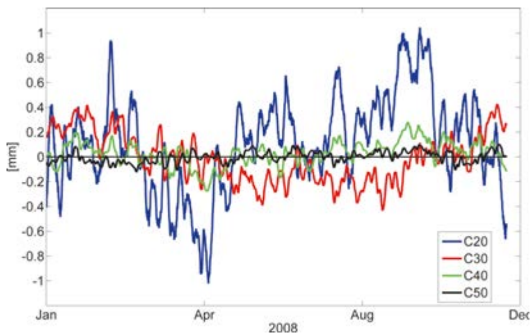

Fig. 8: Time variations for low degree coefficients, calculated with and without loading, expressed in geoid height.

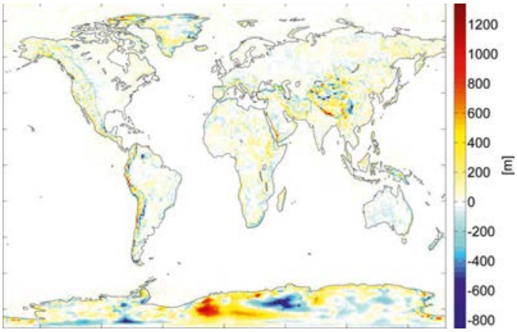

Fig. 9a: Difference of the geocentric surface radii for the orography used by the ECMWF and ETOPO5 (min: -855.76 m, max: 1350.18 m, rms: 51m).

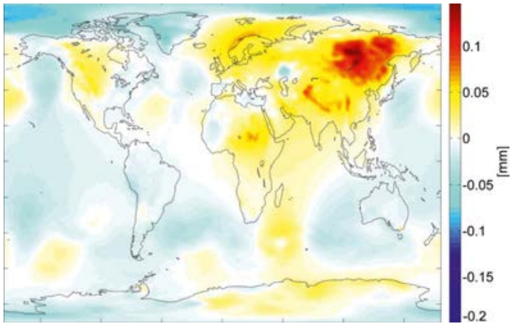

Fig. 9b: Difference of the geoid height variation between the VI approaches with pressure level and model level data for January 1st 2008, 00 UTC (min: -0.09 mm, max: 0.15 mm, rms: 0.018 mm).

Although some features propagate into the final results, their impact is small. This leads to the conclusion that the definition of the Earth surface and the method of vertical discretisation of the atmosphere do not have a significant impact on the actual GRACE processing. Although the differences in height, especially in the Antarctica are huge, those features do not show up in the AGC.

5. Conclusion and outlook

Our de-aliasing product shows good agreement with the official AOD1B product provided by GFZ ([4] Flechtner, 2007), the source for the discrepancies seems to be the different definition of the static mean field of the atmosphere. The current and future space gravity missions demand a very high accuracy in modelling atmospheric effects, both the direct and the indirect effects. We have confirmed that for the actual GRACE mission, in order to reach the predicted error level, the 3D structure of the atmosphere must not be neglected. Also the indirect effect, i.e. loading, has to be modelled, at least for wavelengths longer than 2000 km. Therefore both are applied for the operational GRACE short-term atmosphere and ocean de-aliasing product. Concerning the data sets provided by the ECMWF, the differences between model and pressure level data can be neglected.

Considering the massive computational effort to calculate the VI approach, we developed a new processing strategy, where only a 2D pressure field like for the thin layer approach and the height of the centre of mass of the atmospheric column is needed. First results look promising, especially for the low degrees, but further investigations need to be carried out.

In the results presented here the atmospheric tides (S1 and S2) were not modelled, although they have an impact on the orbiting satellite, as many other forces, too. They will be included in the processing of AGC in the next version to be available at http://ggosatm.hg.tuwien.ac.at/.

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