Time-variable aliasing effects of ocean tides, atmosphere, and continental water mass on monthly mean GRACE gravity field

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[1] The Gravity Recovery and Climate Experiment (GRACE) satellite mission will provide new measurements of Earth's static and time-variable gravity fields with monthly resolution. The temporal effects due to ocean tides and atmospheric mass redistribution are assumed known and could be removed using current models. In this study we quantify the aliasing effects on monthly mean GRACE gravity estimates due to errors in models for ocean tides and atmosphere and due to ground surface water mass variation. Our results are based on simulations of GRACE recovery of monthly gravity solution complete to degree and order 120 in the presence of the respective model errors and temporal aliasing effects. For ocean tides we find that a model error in S_2 causes errors 3 times larger than the measurement noise at $n < 15$ in the monthly gravity solution. Errors in K_1, O_1, and M_2 can be reduced to below the measurement noise level by monthly averaging. For the atmosphere, model errors alias the solution at the measurement noise level. The errors corrupt recovered coefficients and introduce 30% more error in the global monthly geoid estimates up to maximum degree 120. Assuming daily CDAS-1 data for continental surface water mass redistribution, the analysis indicates that the daily soil moisture and snow depth variations with respect to their monthly mean produce a systematic error as large as the measurement noise over the continental regions. *INDEX TERMS:* 1214 Geodesy and Gravity: Geopotential theory and determination; 1223 Geodesy and Gravity: Ocean/Earth/atmosphere interactions (3339); 1227 Geodesy and Gravity: Planetary geodesy and gravity (5420, 5714, 6019); 1241 Geodesy and Gravity: Satellite orbits; 1255 Geodesy and Gravity: Tides—ocean (4560); *KEYWORDS:* temporal aliasing, GRACE mission, time-variable gravity

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1. Introduction

[2] The Gravity Recovery and Climate Experiment (GRACE) launched on March 17, 2002 for a mission life time of 5 years or longer. The mission consists of two identical co-orbiting spacecrafts with a separation of 220 ± 50 km at a mean initial orbital altitude of 500 km with a circular orbit and an inclination of 89° for near-global coverage [*Thomas*, 1999; *Bettadpur and Watkins*, 2000]. The scientific objectives of GRACE include the mapping and understanding of climate-change signals associated with time-varying mass distribution within the solid Earth-atmosphere-ocean-cryosphere-hydrosphere system with unprecedented accuracy and resolution [e.g., *Wahr et al.*, 1998]. New models of Earth's static and time-variable gravity fields will be available every month as one of the science products from the GRACE mission. It has been expected that monthly averaged time-varying mass redistribution of the Earth in the form of climate-sensitive signals can be measured with subcentimeter accuracy in units of column of water movement near Earth surface with a spatial resolution of 250 km or longer, and a temporal resolution of a month for a time-span of 5 years [*Wahr et al.*, 1998; *Han et al.*, 2003; *Nerem et al.*, 2003].

[3] For the recovery of temporal gravity fields, the two largest systematic and high (temporal) frequency signals, ocean tides and atmosphere, are assumed known and removed from GRACE observables. This is accomplished by using currently available models, e.g., CSR4.0 [*Eanes and Bettadpur*, 1995] or NAO99 [*Matsumoto et al.*, 2000] for ocean tides, and ECMWF (European Centre for Medium-range Weather Forecast) or NCEP [*Kalnay et al.*, 1996] for the atmosphere. The errors in these models produce mismodeling of tidal and atmospheric perturbations on the satellite orbit, thus corrupting the recovered gravity fields. In addition, the required time resolution of the gravity field products, i.e., every month, and the distinct characteristics of orbital sampling of GRACE satellites induce high frequency temporal aliasing effects on the monthly mean gravity field estimates. That is, the short period temporal mass variations not present in ocean tide and atmosphere models, as well as continental water mass signals alias into the longer period components and systematically contaminate the monthly mean gravity field estimates. These variations are significant as shown, for example, by *Cheng* [2002] and *Ray et al.*

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[2003] who demonstrate the existence of range-rate signal due to unmodeled ocean tide.

[4] None of the previous studies rigorously deal with the temporal aliasing problem in the recovered GRACE monthly gravity solutions. For example, simulations by *Wahr et al.* [1998] and *Nerem et al.* [2003] started with spherical harmonic coefficients for ocean and continental surface water hydrology based on the reference models. Random noise was added to these coefficients according to the GRACE error degree variance model assuming independent errors among different order coefficients. Finally, spatial averaging was applied in a spectral domain to reduce errors coming from ill-determined higher degree and order coefficients, thus yielding long wavelength estimates of temporal gravity. No rigorous inversion of GRACE data nor a time-wise approach was used in this approach [e.g., *Wahr and Velicogna*, 2003]. These studies are limited to an analysis in the spectral domain and do not completely assess the temporal aliasing error in GRACE monthly mean gravity solutions. As *Velicogna et al.* [2001] mentioned, the aliasing effects cannot be predicted without a detailed orbital simulation, because it is a complex output resulting from GRACE's local sampling (in the space domain as a function of time) averaged over a month. Temporal mass distributions of signals over some regions may not necessarily be "observed" by GRACE because of its distinct orbital sampling characteristics.

[5] In order to identify the realistic aliasing effects on the recovered gravity field, we simulated the perturbations due to the time-variable ocean tidal and atmospheric model errors along the GRACE orbit for a month. Then we fully inverted one month of data to solve the geopotential coefficients up to the maximum degree and order 120 in the presence of the measurement noise, and errors in the ocean tides and atmospheric models. By investigating the recovered monthly gravity coefficient estimates and comparing them with estimates in the presence of measurement error only, we are able to quantify the effect of time-variable tidal and atmospheric effects on the monthly mean GRACE gravity solutions. In the following sections, we describe the models and data used in the GRACE simulation for ocean tides, atmosphere, and continental hydrology. The methodology to generate the perturbations due to these temporal gravity signals will be presented and an efficient method to invert monthly GRACE data for a gravity solution will be concisely described. We will present and analyze in detail the aliasing in the recovered gravity solution due to errors in ocean tide and atmosphere models in both the spatial and spectral domains. Also, the effect of the daily variability (with respect to the monthly mean) of continental surface water hydrology on the monthly mean GRACE gravity estimates will be discussed. Finally, a possible way to reduce the aliasing effects on the monthly mean field is suggested.

2. Data and Method

2.1. GRACE Observable and Inversion

[6] The GRACE mission provides precise range measurements between two satellites tracking each other along approximately the same orbit. The post-processed range-rate data are sampled at the rate of 0.2 Hz and have an accuracy of 0.1 μm/sec in a total RMS [*Kim et al.*, 1999]. This type of along-track measurement is very sensitive to the geophysical fluid mass redistribution under the orbits and it is a fundamental quantity used to solve for the Earth's global (static and temporal) gravity field. The measurements can be transformed into gravitational potential differences using alternative approaches, either on the basis of an energy condition [*Jekeli*, 1999; *Han*, 2004] or on the basis of traditional orbit perturbation analysis [*Tapley*, 1973]. In either case, modeling error affects the geopotential observable directly or indirectly, and, consequently, it affects the derived geopotential model. We conduct an analysis by considering direct aliasing effects on the monthly mean gravity field due to along-orbit sampling of the modeling error. This is most easily visualized as a direct effect in the in situ (energy equation) approach, being a systematic error in the geopotential observable. However, the results hold equally well for the orbital perturbation approach, where model errors indirectly propagate to the geopotential field solution.

[7] The gravitational potential difference between two satellites, V_{12}, can be expressed using the geopotential coefficients as follows:

$$
\begin{aligned}
V_{12}(r_1,\theta_1,\lambda_1,r_2,\theta_2,\lambda_2) &= V(r_1,\theta_1,\lambda_1) - V(r_2,\theta_2,\lambda_2) \\
&= U_0 \sum_{m=0}^{N\max} \sum_{n=m}^{N\max} \left\{ \left\{ \left(\frac{R}{r_1}\right)^{n+1} Y^c_{nm}(r_1,\theta_1,\lambda_1) \right.\right. \\
&\quad \left. - \left(\frac{R}{r_2}\right)^{n+1} Y^c_{nm}(r_2,\theta_2,\lambda_2) \right\} \overline{C}_{nm} \\
&\quad + \left\{ \left(\frac{R}{r_1}\right)^{n+1} Y^s_{nm}(r_1,\theta_1,\lambda_1) \right. \\
&\quad \left.\left. - \left(\frac{R}{r_2}\right)^{n+1} Y^s_{nm}(r_2,\theta_2,\lambda_2) \right\} \overline{S}_{nm} \right\}, \quad (1)
\end{aligned}
$$

where $U_0 = \frac{GM}{R}$, $Y^c_{nm}(r_i, \theta_i, \lambda_i) = \overline{P}_{nm}(\cos\theta_i)\cos m\lambda_i$, $Y^s_{nm}(r_i, \theta_i, \lambda_i) = \overline{P}_{nm}(\cos\theta_i)\sin m\lambda_i$, i = 1 or 2. GM is the gravitational constant times the mass of the Earth, R is the Earth's mean radius, $(r_1, \theta_1, \lambda_1)$ and $(r_2, \theta_2, \lambda_2)$ are the coordinates (radius, colatitude, and longitude) of the first and second satellites, respectively, $\overline{P}_{nm}$ is the fully normalized, associated Legendre function of degree n and order m, and $\overline{C}_{nm}$ and $\overline{S}_{nm}$ are the gravitational spherical harmonic coefficients of degree n and order m. This is the fundamental model used to analyze the potential difference generated by the time-variable gravitational fields, i.e., due to ocean tides, atmosphere, and hydrology. If the spherical harmonic coefficients for the time-variable sources are available from some model or data, we can compute their gravitational effects in terms of the potential difference along the orbit through (1). Subsequently, having the total gravitational potential difference observables along the orbit, we can estimate the spherical harmonic coefficients in a least squares sense using (1). This process is an inversion process, which is usually computational intensive. We have developed and used an efficient inversion method [*Han*, 2004] for assessing and quantifying the aliasing effects in GRACE monthly gravity solutions.

2.2. Model for Ocean Tide Error

[8] As a possible indicator for the modeling error of ocean tides, the difference between two distinct tide

models can be used. Here, we used the difference between a hydrodynamic model, NAO99 [*Matsumoto et al.*, 2000] and the altimetry-based model, CSR4.0 [*Eanes and Bettadpur*, 1995]. The effect of tide model errors on GRACE monthly solution have been studied by *Ray et al.* [2001, 2003], *Knudsen and Andersen* [2002], and *Knudsen* [2003]. *Ray et al.* [2001] first studied the effect of M_2 error on GRACE by averaging 12.4 hours of M_2 error for a 3-month averaged GRACE sensitivity. *Knudsen and Andersen* [2002] followed with a comparable study and computed the tidal aliasing frequencies [*Parke et al.*, 1987] for the four most energetic constituents. They computed a monthly mean tidal error by applying convolution in a time domain using a block averaging function with corresponding aliasing periods. In the more realistic investigations such as *Knudsen* [2003] and *Ray et al.* [2003], they considered orbital sampling and found sectorial anomalies in the recovered gravity field. *Cheng* [2002] used a semianalytic formulation and modeled the GRACE range-rate observations to study the detailed perturbations due to ocean tides. In this study, the effects of the tidal model error, defined as the difference between CSR4.0 and NAO99, were transformed into the GRACE potential difference observables. By inverting these observables, the effects of their temporal variation on the monthly mean gravity field estimates were determined. The analysis considers orbital sampling and the periods of each tide constituent. This approach provides a more realistic assessment on the effect of time-variable tidal error on monthly GRACE gravity solutions.

[9] The time-variable ocean tide was decomposed into temporal sine and cosine components and each component was expanded into spherical harmonic coefficients. Therefore each tidal constituent consists of 4 sets of coefficients, $\overline{C}^C_{nm}$, $\overline{S}^C_{nm}$, $\overline{C}^S_{nm}$, and $\overline{S}^S_{nm}$. Coefficients with the superscript, C, pertain to the cosine component, and those with the superscript, S, pertain to the sine component. The corresponding gravitational potentials at satellite altitude generates by a particular constituent with frequency, ω, and initial phase, ϕ^0, given by

$$\Delta V_{12}(r_1,\theta_1,\lambda_1,r_2,\theta_2,\lambda_2;t) = U_0 \sum_{m=0}^{Nmax}\sum_{n=m}^{Nmax}\left\{A_{nm}\Delta\overline{C}^t_{nm}(t) + B_{nm}\Delta\overline{S}^t_{nm}(t)\right\}, \quad (2)$$

where $\Delta\overline{C}^t_{nm}(t) = \Delta\overline{C}^C_{nm}\cos(\omega t + \varphi^0) + \Delta\overline{C}^S_{nm}\sin(\omega t + \varphi^0)$, $\Delta\overline{S}^t_{nm}(t) = \Delta\overline{S}^C_{nm}\cos(\omega t + \varphi^0) + \Delta\overline{S}^S_{nm}\sin(\omega t + \varphi^0)$. The symbol Δ stands for the difference between two tidal models and t is time. A_{nm} and B_{nm} are the same factors appearing in front of the coefficients in (1), which depend on the orbital coordinates. The in situ potential difference error, ΔV_{12}, induced by the ocean tidal model error, is expressed by a function of time and positions of the two satellites, and it can be computed along the simulated GRACE orbit.

2.3. Model for Atmosphere Error

[10] The surface pressure fields can be used to compute global atmospheric mass redistribution [*Chao and Au*, 1991]. Even though the nominal thickness of the atmosphere is around $10 \sim 15$ km, the vertical variation of the atmospheric water mass is disregarded in this study. The entire atmosphere is assumed to be condensed onto a very thin layer on the Earth's surface. Global surface pressure data are available through the European Center for Medium-range Weather Forecast (ECMWF) and National Centers for Environmental Prediction (NCEP). Both global circulation models assimilate common data including the barometric surface pressure data; thus the two models are not completely independent and they lack information over Antarctica [*Ge et al.*, 2002; *Nerem et al.*, 2000]. Here, we use the difference between the two models as the atmospheric modeling error (residual atmosphere). A commonly used scaling factor, $1/\sqrt{2}$ [*Velicogna et al.*, 2001; *Nerem et al.*, 2003], was not applied because the two models are obviously not independent and the error level might, otherwise, be underestimated.

[11] The surface pressure data, p_s, is converted to the equivalent water thickness, h, using the following relationship:

$$h(\theta,\lambda,t) = \frac{p_s(\theta,\lambda,t)}{g\sigma_w}, \quad (3)$$

where h(θ, λ, t) also represents the anomalous surface mass in terms of water height. g is the nominal gravity value, and σ_w is the density of water (1000 kg/m^3). We use the quadrature equation (as used by *Wahr et al.* [1998] and *Hwang* [2001]) to compute the spherical harmonic coefficients of atmospheric mass change based on a regular grid of equivalent water thickness data, considering also the elastic Earth's direct response to loading:

$$\left\{\begin{array}{c}\overline{C}^a_{nm}(t)\\ \overline{S}^a_{nm}(t)\end{array}\right\} = \frac{3(1+k_n)\sigma_w}{4\pi R\sigma_E(2n+1)}\iint h(\theta,\lambda,t)\overline{P}_{nm}(\cos\theta) \cdot \left\{\begin{array}{c}\cos m\lambda\\ \sin m\lambda\end{array}\right\}\sin\theta d\theta d\lambda, \quad (4)$$

where $\bar{C}^a_{nm}(t)$ and $\bar{S}^a_{nm}(t)$ are the spherical harmonic coefficients of the time-variable surface mass change. k_n is the load Love number of degree n that describes the Earth's elasticity, and σ_E is the average density of the Earth.

[12] On the basis of NCEP and ECMWF gridded surface pressure data, the corresponding spherical harmonic coefficients were computed up to maximum degree 60 at 6 hour intervals for a month (that is, 120 sets of coefficients with maximum degree of 60). Assuming one of them is the "truth" atmosphere and the other is its estimate, the coefficients for the residual atmosphere were computed every 6 hours by differencing the ECMWF and NCEP models. From the computed coefficients, the perturbations in the GRACE potential difference measurements due to the residual atmosphere were computed along the GRACE orbits for one month using (2). We will quantify and analyze how much these short period (6 hour) perturbations affect the monthly mean gravity estimates.

2.4. Models for Continental Surface Water

[13] Although previous temporal gravity studies [*Wahr et al.*, 1998; *Rodell and Famiglietti*, 1999; *Velicogna et al.*, 2001] predicted the successful recovery of continental surface water mass from the GRACE mission, it is still uncertain how much the variability of the surface water mass affects the monthly mean GRACE gravity field. Using a similar

approach as for the atmosphere model error, the error due to the variability of the groundwater mass can be quantified in the measurement domain as well as the spectral domain based on a certain model which can reasonably describe the short period variations of continental surface water.

[14] The daily continental water storage can be computed by using NCEP/NCAR (National Centers for Environmental Prediction/National Center for Atmospheric Research) reanalysis products (http://www.cdc.noaa.gov) including the daily mean soil moisture within 2 layers (0–10 cm, 10–200 cm) and the snow accumulation data (referred to as CDAS-1). Data in terms of a volumetric fraction are converted into equivalent water heights. They represent the daily mean water contents in the upper 2 meters of the soil layer (IERS Global Geophysical Fluids Center, Special Bureau for Hydrology, http://www.csr.utexas.edu/research/ggfc). The surface water mass redistribution is caused by the hydrological cycle, i.e., precipitation, evaporation, transpiration, and runoff. The GRACE satellites are supposed to be influenced by this mass redistribution phenomena, and this redistribution is recoverable from the GRACE data on a monthly basis [*Rodell and Famiglietti*, 1999]. The regularly gridded data cover the entire continents with a spatial and temporal resolution of about 2 degrees and a day, respectively, from 1979 to 2002. Data over Greenland and Antarctica are excluded in this study, because the NCEP reanalysis defines constants over those areas; however, unrealistic variability can also be found along the coastlines (J. Chen, personal communication, 2002).

[15] The water storage anomaly (WSA) at a certain time is defined as the difference between water content at that time and water content averaged over some years (say, 5 years). The one-month average of daily WSA's is called the monthly mean WSA (MWSA), which is what the GRACE mission would improve every month. MWSA indicates the monthly mean continental surface water mass redistribution with respect to the 5-year mean field. The 5-year mean is highly correlated with the static GRACE gravity field, thus hardly separable from the static gravity estimates. Therefore what GRACE can identify is not a total continental surface water mass, but an anomaly field like MWSA. Daily mean WSA (DWSA) can be computed in the same way as MWSA, on the basis of the daily water storage data. In this investigation, we quantify the effect of daily variability on the monthly mean GRACE gravity field. The WSA is converted to equivalent water thickness, which then expresses the anomalous surface mass in terms of water height. It can be computed by dividing the surface density (mass per area) of anomalous mass by the volume density (mass per volume) of the water. The corresponding geopotential coefficients are computed using (4) and the perturbations in the GRACE potential difference measurements due to the daily hydrology are computed along the GRACE orbits for one month using (2). Near the boundary between the ocean and continent, there will be effects caused by Gibb's phenomenon. However, the investigation of this effect is outside our scope and neglected in this study.

3. Ocean Tides

[16] To obtain the magnitude of the tide model error, the square root of degree variances (the degree RMS) of four mean ocean tidal constituent errors in terms of the geoid height effect were calculated and compared with the monthly GRACE sensitivity (Figure 1). The GRACE sensitivity curve is based on our own noise-only simulation (assuming 0.1 μm/s (RMS) error in range-rate or the corresponding potential difference error 0.8×10^{-3} m^2/s^2 [see *Wolff*, 1969; *Rummel*, 1980; *Jekeli and Rapp*, 1980]. It follows approximately the one used in other studies [e.g., *Wahr et al.*, 1998]. The mean tide error was computed by averaging the power of time-variable residual tidal coefficients over a complete tidal cycle of each constituent as done by *Ray et al.* [2001]. That is, the mean M_2, S_2, K_1, and O_1 tidal errors (differences between the two distinct models; see section 2.2) were calculated by averaging the sinusoidal variation over their periods of 12.42, 12.00, 23.93, and 25.82 solar hours, respectively [see also *Knudsen and Andersen*, 2002, Figure 1; *Ray et al.*, 2001, Figure 2]. Two time-variable gravity signals, monthly mean ocean and groundwater storage anomaly, were plotted in the same figure. Figure 1 indicates that the lack of knowledge about the ocean tide would affect coefficients less than degree 20 significantly compared to the monthly GRACE sensitivity.

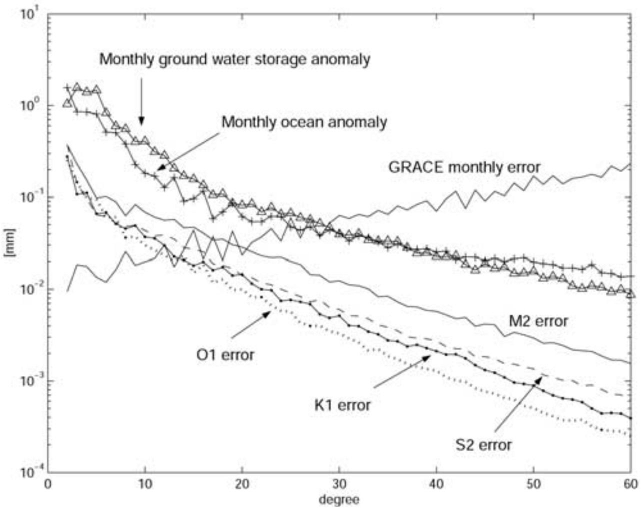

Figure 1. Degree RMS of four mean ocean tidal constituent errors, monthly mean temporal gravity signals, and monthly GRACE sensitivity in terms of the geoid.

[17] However, the foregoing mean tidal error analysis disregards the characteristics of the time-varying tidal model error and the orbital sampling from GRACE. The ocean tide model as well as its error varies periodically over time at every location, which implies that the ocean tidal model errors can be mitigated by some suitable sampling and averaging. The monthly mean gravity field solution might be less corrupted by the tidal model errors because they will be averaged over one month. The realistic time-varying tidal model errors were computed along the GRACE orbit for 30 days in terms of the potential difference and are presented in Figure 2. The true period of each constituent was used, but the initial phase was disregarded in this simulation study.

[18] Figures 2a to 2d show four global maps of four (month-long) time series of each tidal constituent error. The error tends to be small over the middle ocean areas, which

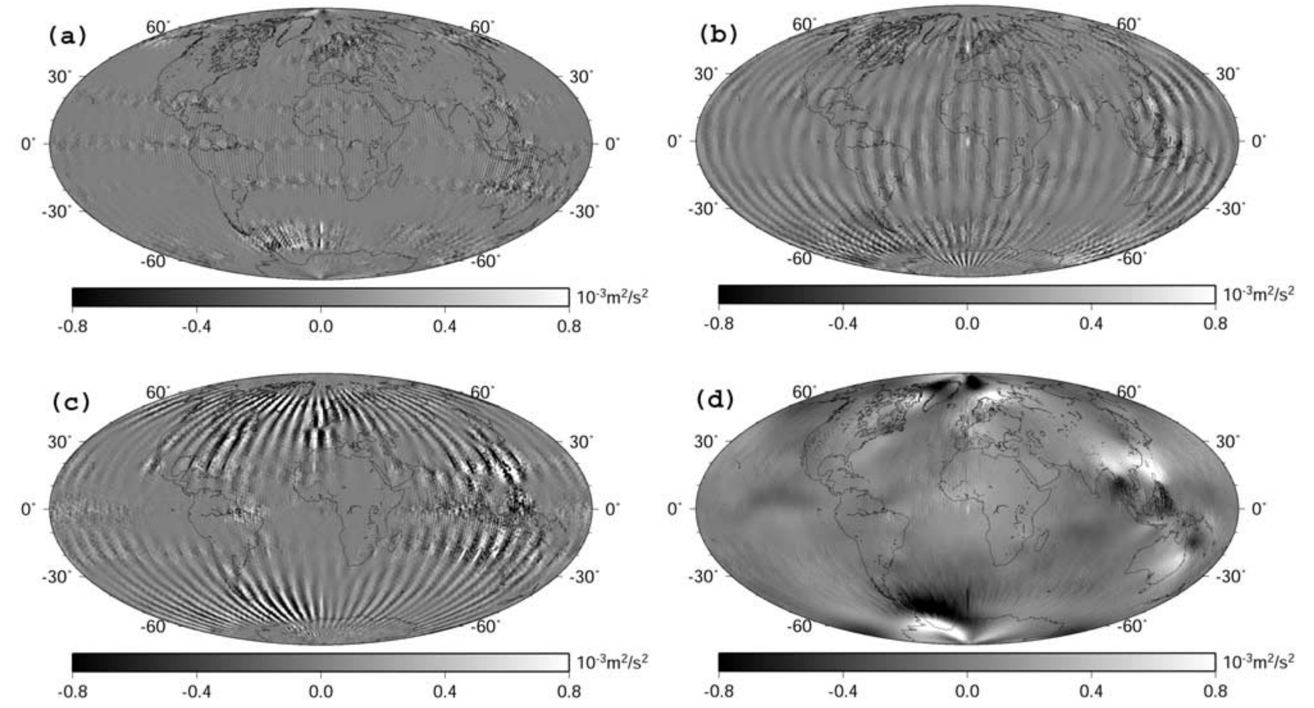

Figure 2. The time-varying tidal model errors computed along GRACE orbit for 30 days in terms of the potential difference (mapped with respect to the leading satellite): (a) K_1, (b) O_1, (c) M_2, and (d) S_2.

may not represent a realistic error level in the model because the two models, NAO99 and CSR4.0, used the same TOPEX/POSEIDON (T/P) data. For constituents such as K_1, O_1, and M_2, the short-wavelength sectorial variation of the error is dominant, while the long wavelength features are dominant in the S_2 error map. The errors for K_1, O_1, and M_2 show consecutive positive and negative values along the longitude with a resolution of about 5 $\sim$ 6 degrees, corresponding to spherical harmonic orders 30 $\sim$ 36. We can expect that these sectorial variations affect the harmonic coefficients of all degrees and corresponding specific orders in the monthly mean gravity field estimates. In order to reduce the contribution of the short wavelength (high degree and order) components of the tidal model errors and check their effects on low degree and order harmonic coefficients of the monthly mean field, the Gaussian averaging function was applied in the spatial domain. A horizontal radius of 800 km was used to filter out the anomalies having wavelengths shorter than 1600 km. Therefore components beyond degree 25 will be reduced using the radius of 800 km. Except for the S_2 tidal error, the filter reduces the errors significantly. It indicates that tidal errors from the K_1, O_1, and M_2 constituents affect less the low degree and order coefficients (n, m $\leq$ 25) of the monthly mean geopotential estimates, while the S_2 tidal error corrupts them significantly. The global RMS values were reduced from $0.4 \sim 0.5 \times 10^{-3}$ m^2/s^2 to 0.08×10^{-3} m^2/s^2 for the cases of the K_1, O_1, and M_2 error after low-pass filtering, while the RMS of S_2 remains the same at 0.3×10^{-3} m^2/s^2.

[19] The characteristic that the sun-synchronous constituent, S_2, is not canceled at all, can be explained by analyzing the aliasing period of each tidal constituent. An under-sampling of a certain signal makes a sampled signal appear as if it has a longer period (called the aliasing period) than what the original signal has. The aliasing period, T_a, depends on the period of the signal and the Nyquist period (inverse of Nyquist frequency), and can be computed as follows [*Parke et al.*, 1987]:

$$\frac{1}{T_a} = \mathrm{abs}\left[\mathrm{mod}\left(\frac{1}{T_k} + \frac{1}{T_N}, \frac{2}{T_N}\right) - \frac{1}{T_N}\right], \tag{5}$$

where T_k is a period of a signal, and T_N is the Nyquist period. *Knudsen and Andersen* [2002] used a half sidereal day (0.4986 solar day) as a sampling interval, $T_{N/2}$, for GRACE, because each tidal constituent will be sampled approximately twice per sidereal day along the ascending and descending tracks. In this approximation, the aliasing period of K_1 is 23.94 solar hours, which is slightly longer than the original period, 23.93 solar hours. The aliasing period of O_1 is same as the original one, 25.82 solar hours, because there is no aliasing for O_1. However, M_2 and S_2 have aliasing periods of 13.7 and 182.5 solar days, respectively, the same as given by *Knudsen and Andersen* [2002]. Errors of constituents such as K_1, O_1, and M_2 are expected to cancel in the monthly mean gravity estimates, because their aliasing period is still less than a month. However, the S_2 error does not cancel in monthly mean field, because of its much longer aliasing period. It generates a kind of systematic long wavelength anomaly over the globe as shown in Figure 2d with semiannual characteristics in the temporal gravity field estimates.

[20] In the presence of these four systematic tidal model errors (but no measurement noise), four monthly mean gravity fields were recovered up to degree and order 120. For the purpose of comparison, the simulation was also

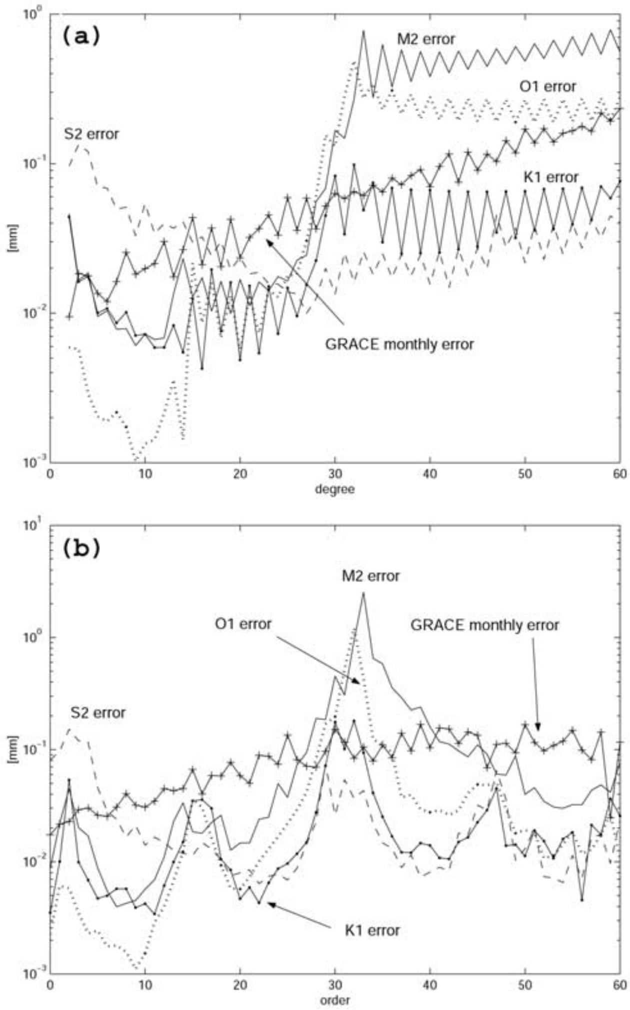

Figure 3. (a) Degree RMS and (b) order RMS of recovered coefficients in the case of noise only and noise combined with residual ocean tide perturbation.

performed in the presence of measurement noise only (but no tidal errors). The true coefficients were subtracted from each of the five sets of estimated geopotential coefficients to determine the distinct effects due to the tidal constituent error and the measurement noise. Figure 3a shows the degree RMS of the errors (in terms of geoid) existing in the recovered spherical harmonic coefficients. Except for the S_2 error, the tidal errors affect the low degree coefficients (below degree 30) less than the measurement noise does. The degree RMS jumps at degree 31 and remain constant at higher degrees. However, it should be mentioned that not all coefficients beyond degree 30 were corrupted by tidal error, because its effect is limited to a certain order and all degrees (see the following paragraph and Figure 3b). The curve for the S_2 error indicates that it remains and significantly corrupts the low degree harmonic coefficients in the monthly averaged field. It does not cancel out, having nearly the same power as its mean error shown in Figure 1, because its aliasing period is much longer than one month. In order to identify the reason for the jumps at degree 31 in Figure 3a, the order RMS (square root of order variances) were computed and are depicted in Figure 3b. Some errors turned out to be significant relative to the effect of the measurement noise; at orders less than 10 for S_2 and at $30 \sim 36$ for the other constituents. While the error in the S_2 model tends to corrupt all low degrees and orders, errors in the other constituents significantly corrupt all degrees and but only certain orders $30 \sim 36$.

4. Atmosphere

[21] The perturbations due to the residual atmosphere (6 hour ECMWF-NCEP) were computed along one month of GRACE orbits at 30-second intervals in terms of the potential difference, depicted in Figure 4. Its global RMS is 0.7×10^{-3} m^2/s^2, which corresponds almost to the same level as the expected total RMS of post-processed GRACE instrument precision (10 μm in range and 0.1 μm/sec in range-rate (J. R. Kim, personal communication, 2002)). In this map, notice that the global low degree and order features remain as in the case of perturbations due to the S_2 tidal error for a month. In addition, high degree and order features are found because of GRACE's orbital sampling of the time-variable residual atmosphere.

[22] In the presence of noise and residual atmosphere, monthly geopotential difference data were inverted and the spherical harmonic coefficients were computed up to degree and order 120. For the comparison, the same simulation was done in the presence of noise only. Figures 5a and 5b show degree RMS and order RMS of coefficients errors (in terms of geoid error) for both cases of noise only and noise with residual atmospheric perturbation. The monthly global mean of ECMWF and NCEP surface pressure data were computed and the corresponding spherical harmonic coefficients were calculated. The degree RMS of coefficient differences between monthly mean ECMWF and NCEP were computed and presented in Figure 5a. From Figure 5a, we see two distinct characteristics near the boundary of degree $20 \sim 30$. The coefficient errors less than degree 20 (in the presence of noise with the residual atmosphere) show a trend and magnitude similar to the monthly global mean ECMWF-NCEP. If the residual atmosphere does not vary in time and is fixed as its global mean for a month, then the error curve of the recovered coefficients from GRACE monthly data would be similar to the sum of the two curves of monthly global mean ECMWF-NCEP and monthly

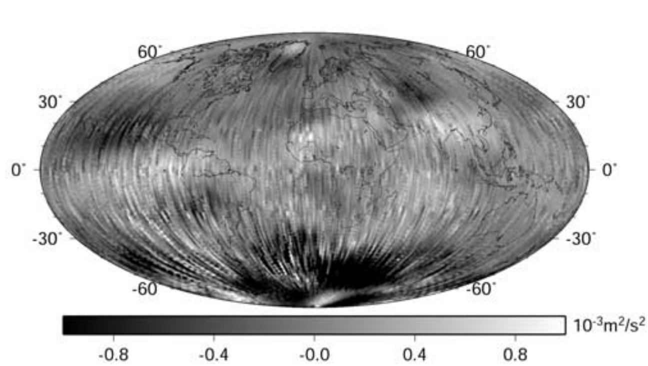

Figure 4. Time-varying atmospheric modeling error computed along GRACE orbit for 30 days in terms of potential difference (mapped with respect to the leading satellite).

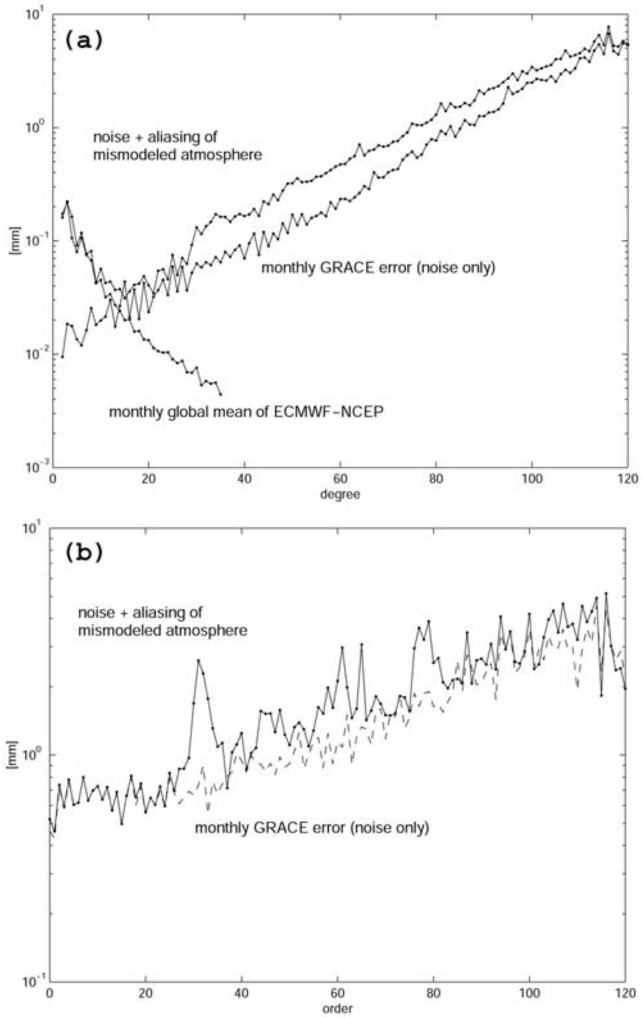

Figure 5. (a) Degree RMS and (b) order RMS of recovered coefficients in the case of noise only and noise combined with residual atmospheric perturbation.

GRACE sensitivity (noise only). However, the realistic (time-variable residual atmosphere) error curve shows significant deviations beyond degree 30 from the noise-only error curve. Therefore we can conclude that these larger errors over higher degrees (n > 30) are due to sampling of the time-varying residual atmosphere with an equivalent temporal resolution of one month in the monthly solution of the gravity field.

[23] Even though most of the power of the residual atmosphere is limited to low degree and order (say, less than 20) in this simulation, we see that the orbital sampling of its temporal variations for a month caused errors over the entire spectrum of estimates. Note that the temporal aliasing is the effect caused by sampling a signal less frequently than needed to discriminate its entire spectrum. The residual atmosphere is not periodic thus its aliasing period can not be computed, unlike the ocean tidal constituents. However, it was shown that the coefficient errors beyond degree 30 are due to a monthly sampling of low degree and order residual atmosphere varying with shorter periods. This temporal aliasing effect could be larger than the measurement noise. It is not limited to low degree and order harmonics, even though the input residual atmosphere was limited to degree 50. Therefore the atmospheric mismodeling can degrade the monthly GRACE mean gravity product at all spatial frequencies. In terms of the geoid height (degree and order up to 120), the atmosphere model error and its temporal aliasing introduce 30% more error. Considering the initial altitude of 400 km in this simulation, GRACE is performing 15.6 revolutions per day and resonance occurs at orders, m, close to 16, 31, 47, 62, 78, etc. Figure 5b shows that the resonant and near resonant orders are most severely affected by the temporal variation.

[24] Some ad hoc methods to mitigate the temporal aliasing effects especially in the resonant orders (secondary resonant orders, m = 31 or 32) could be considered. Even though there is nothing to reduce the inherent modeling error except using a more accurate model, the effect of its high frequency temporal variation can be reduced by solving resonant coefficients more frequently, e.g., every day, or adding more (nongravitational) stochastic parameters to absorb the aliasing effects. The latter might be dangerous, because the additional parameters can absorb the gravitational effects too. Thus we tried to solve for resonant and near resonant coefficients (m = 30, 31, 32, and 33) every day and others every month. The daily estimates for resonant and near resonant coefficients were averaged to produce monthly mean estimates of daily solutions. The errors of the monthly mean of daily solutions and the original monthly solution were compared in terms of the geoid degree and order variances. Figure 6a shows *partial* degree RMS of errors in the recovered resonant coefficients (just m = 30, 31, 32, and 33) in three ways; (1) monthly solution in the presence of measurement noise only, (2) monthly solution in the presence of measurement noise and atmospheric modeling error, (3) one month average of daily solutions in the presence of measurement noise and atmospheric modeling error. By comparing the cases (2) and (3), we see some improvement in the recovered resonant coefficients below degree 90. However, the average of daily solutions is not better than monthly mean solution at degrees higher than 90. It might be due to the fact that one day is too short a time span to recover such relatively high degrees ($n \geq 90$). Figure 6b shows the order RMS of errors for the above three cases. They were computed from degree 30 to degree 90. Each order (degrees lumped) shows slight improvement (submillimeter) in the geoid height.

5. Continental Surface Water

[25] The monthly mean WSA and the equivalent water height were computed for January, 2001. No model and data were used over Antarctica and Greenland. The overall magnitude is at the decimeter level. The monthly mean WSA is the anomalous quantity with respect to the 5-year mean field. During a month WSA is not static, of course, hence its short period temporal variability will contaminate the monthly mean WSA estimate. In order to quantify how much the temporal variability of WSA affects the GRACE potential difference observable, one month of daily WSA data and the monthly mean WSA data were used to compute the corresponding geopotential difference. Figure 7 shows the difference between one month of daily WSA's and monthly mean WSA in terms of the potential difference

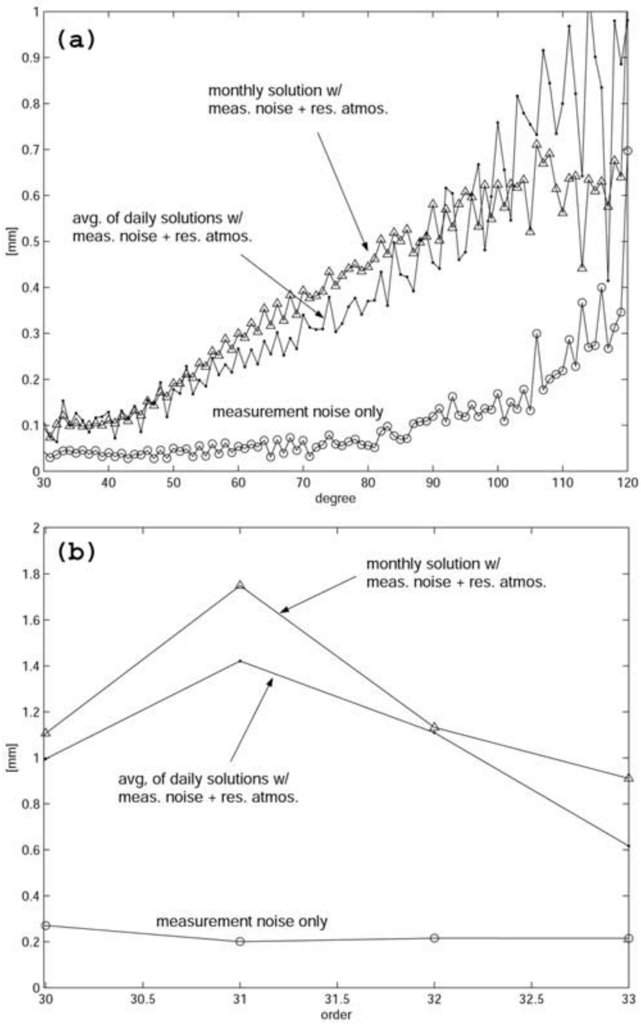

Figure 6. (a) Partial degree RMS (m = 30, 31, 32, and 33) of errors in the geoid height; (b) order RMS ($30 \leq n \leq 90$) of errors in the geoid height: coefficients error (monthly mean solution) in the presence of measurement noise only; coefficients error (monthly mean solution) in the presence of measurement noise and the residual atmosphere; coefficients error (one month average of 30 daily solutions) in the presence of measurement noise and the residual atmosphere.

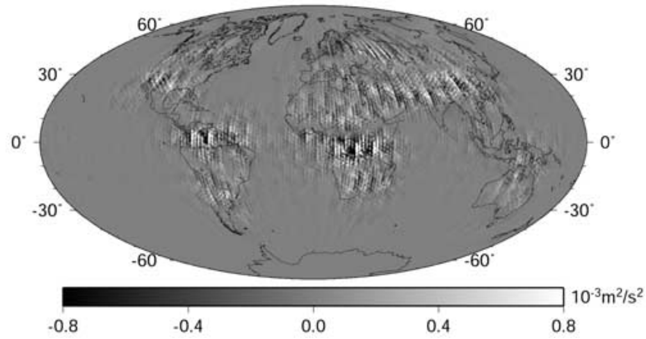

Figure 7. The perturbation in the GRACE potential difference measurements due to daily variability of WSA for a month (mapped with respect to the leading satellite).

along the GRACE orbit for a month. If GRACE samples the WSA globally and instantaneously, Figure 7 would show only zero. However, the orbital sampling from GRACE causes nonzero perturbations due to the temporal variability of WSA. Clearly, the significant perturbations are limited to the continental regions and their magnitudes are in the level of several 1.0×10^{-3} m^2/s^2.

[26] Two simulations were performed in order to recover the continental water mass redistribution in the presence of measurement noise. The input for the first computation was based on MWSA data, hence the only thing to prohibit the perfect recovery is the measurement noise. The input for the second computation was based on one month of DWSA data, thus the measurement noise as well as the temporal aliasing effect due to the short period (daily) variability would remain in the recovered coefficients. Figure 8 shows the degree RMS of the "truth" MWSA (no measurement noise) and two recovered coefficients in the presence of noise with MWSA and noise with one month DWSA. Note that the power of the recovered coefficients is larger than "truth" MWSA, because of contamination due to the noise over high degrees. To highlight the short period aliasing effect, the difference between results from the first and second computation was computed and its degree RMS was depicted in the same figure. The effect of the daily varying WSA seems to be less than the effect of measurement noise for low degree and orders harmonics (n, m $\leq$ 30). However, it is as significant as the measurement noise beyond degree and order 30. By checking the degree RMS over higher degrees (not drawn here), it was found that hydrological aliasing affects the entire spectrum of coefficients covering the higher degrees and orders, as well; however, it was still less than the effect of measurement noise. It is emphasized that the degree RMS values may not be a good indicator showing the realistic temporal aliasing effects of local (continental) hydrology. The continental water mass redistribution is not a global phenomenon but a local one. Its temporal variation is limited to continental area, which is just 30% of the Earth. Therefore the spectrum based on the global spherical harmonic representation may not present

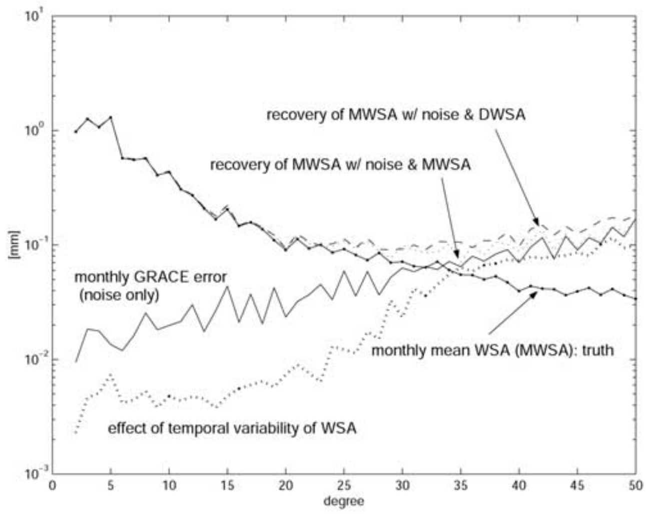

Figure 8. The degree RMS of the "truth" MWSA, its recoveries, and the aliasing effect.

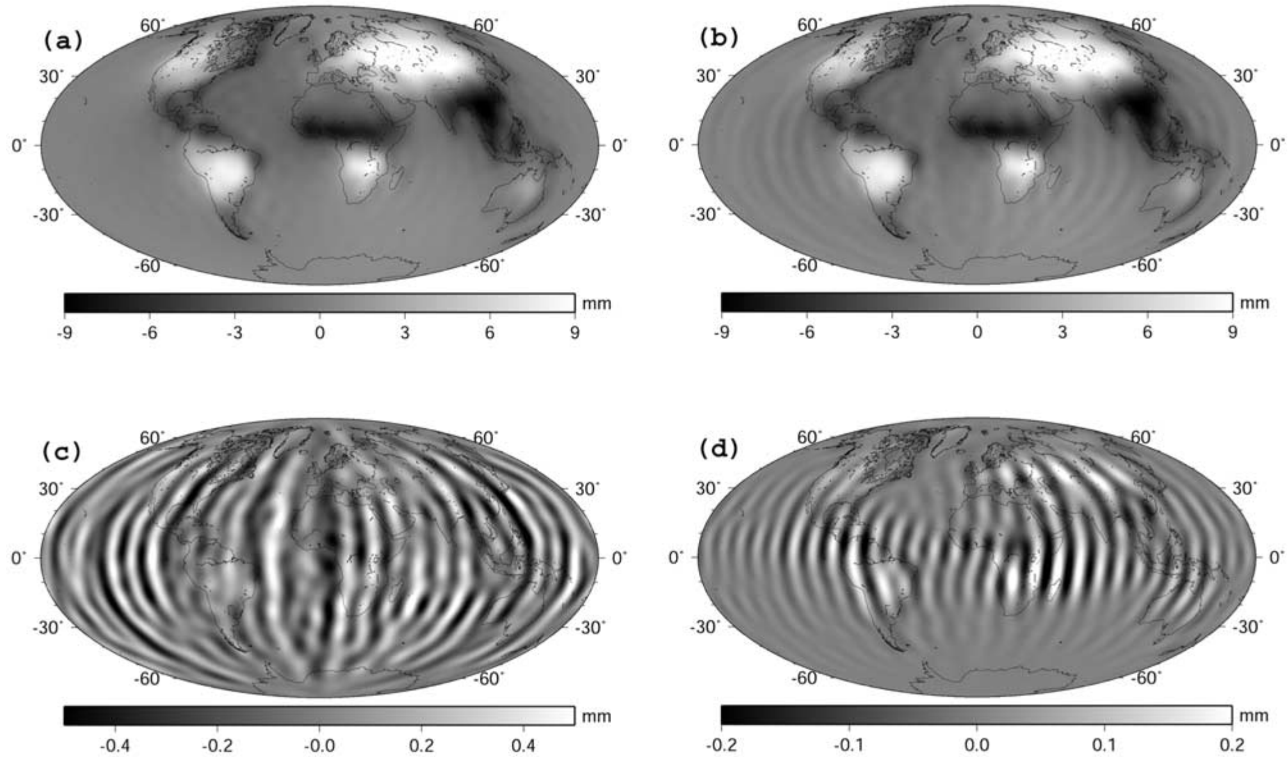

Figure 9. (a) The "truth" geoid change; (b) the recovered geoid change; (c) the effect of noise and aliasing; and (d) the aliasing effect only.

the actual spectral contents of the high-frequency time-varying effects of local hydrology.

[27] To quantify the aliasing effect on the geoid, the geoid changes using the "truth" MWSA coefficients and the recovered coefficients from the second test were computed the degree and order up to 30, because the error and signal spectra cross at degree 30. Figure 9a shows the "truth" geoid change due to the monthly mean WSA, and Figure 9b shows the recovered geoid change in the presence of the measurement noise and daily WSA. Figure 9c shows the effect of noise and daily WSA, i.e., difference between (a) and (b), and (d) shows the effect of daily WSA (aliasing effect), i.e., difference between the recovered coefficients from the first and second tests. The global RMS values of the geoid change signal, the effect of noise, and the effect of aliasing are 2.49, 0.16, and 0.05 mm, respectively. Even though the global RMS of the geoid error due to the temporal aliasing of daily hydrology seems to be very small (several hundredth of a mm), it should be noted that the effect over the continental regions could be as strong as the effect of measurement noise. For example, the geoid error could reach up to a few 0.1 mm over the central Asia and South Africa, as shown in Figure 9d. The small RMS is simply due to the fact that the local effects were averaged globally. Figure 9d also indicates that the effect of temporal variability of continental surface water corrupt the gravity field over the ocean area around the continents.

6. Conclusion

[28] The time-variable effects of ocean tidal and atmospheric modeling errors and continental water on the monthly mean GRACE gravity field were analyzed by simulating realistic GRACE orbits and estimating the gravity field. The analysis is based on the current geophysical fluid models. Even though the error based on the model differences may underestimate the actual error level, the characteristics of the temporal aliasing shown here would remain the same.

[29] The analysis based on CSR4.0-NAO99 indicates that the constituents (K_1, O_1, and M_2) have aliasing period shorter than one month and correspondingly corrupt mostly the harmonic coefficients of orders 30 $\sim$ 36 and all corresponding degrees. Their effect on coefficients less than degree and order 30, where most of the power lies for the temporal gravity field like ocean mass and groundwater redistribution, is less than 40% of the effect of measurement noise. However, the error of tidal constituent, S_2, has a long aliasing period and does not average out in the monthly mean gravity field. It significantly distorts the low degree and order harmonic coefficients in the recovered gravity field solution. The effect of S_2 modeling error is 3 times larger than the effect of measurement noise at degrees less than 15.

[30] Using ECMWF-NCEP with 6-hour resolution, we found that the residual atmosphere can considerably contaminate the monthly mean gravity coefficients over all degrees and orders. The errors in the recovered coefficients less than degree 20 due to the atmospheric mismodeling are large enough to corrupt the monthly temporal gravity (geoid) estimates significantly. The errors beyond degree 30 have larger magnitude than the measurement noise over all degrees, thus the temporal aliasing effect degrades the overall (all degrees) performance of monthly mean GRACE

gravity products. By looking at the order RMS, it was found that resonant orders (even m = 78) were especially contaminated by the residual atmosphere, even though the input residual atmosphere was limited in degree and order up to 50. As an alternative solution strategy, we demonstrated that a more frequent (e.g., daily) estimation of resonant coefficients could help decrease the effects coming from the temporal variability of the residual atmosphere.

[31] On the basis of NCEP/NCAR daily CDAS-1 data, it was found that the daily variability with respect to the monthly mean are strong as the effect of measurement noise over the continental region. All recovered coefficients are affected by the daily variability, even though the input of the hydrology signal was limited in spatial degree to 50. However, the effect over the continental region only (30% of the Earth) is significant as much as the measurement noise. In terms of the geoid height, for example, the effect of the daily variability over the continental region is at the level of 0.1 $\sim$ 0.2 mm comparable to the effects of noise, while its global RMS is about 3 times smaller. Therefore it can be said that the effects of the high frequency temporal (daily) variability of the continental water mass are significant over the continental regions and they corrupt the continental (local) hydrology recovery as much as the measurement noise, assuming the current CDAS-1 data represent the variability reasonably well.

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