

Using GRACE to explain variations in the Earth's oblateness

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Key Points:

- The proposed approach allows to both estimate and interpret the J_2 time series.
- The obtained data-driven time series of J_2 variations is consistent with SLR solutions.
- Inter-annual J_2 variations are primarily explained by mass variations in atmosphere, oceans and land waters.

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Abstract

We present a new approach to estimate time variations in J_2 . Those variations are represented as a sum of contributions from individual sources. This approach uses solely GRACE data and the geoid fingerprints of mass redistributions that take place both at the surface and in the interior of the solid Earth. The results agree remarkably well with those based on satellite laser ranging, while estimates of the sources explain the observed variations in J_2 . Seasonal variations are dominated by terrestrial water storage (TWS) and by mass redistribution in the atmosphere and ocean. Trends, however, are primarily controlled by the Greenland and Antarctic ice sheets and by glacial isostatic adjustment (GIA). The positive trend from surface mass variations is larger than the negative trend due to GIA and leads to an overall rising trend during the GRACE period (2002 - 2017).

Plain Language Summary

J_2 variations indicate changes in the flattening of the Earth, which are mainly due to the Earth’s response to large-scale mass redistribution at its surface and related to ongoing climate change. Though monitored over four decades by Earth observations satellites, the contributing sources to J_2 variations have not yet been accurately constrained, mostly due to deficiencies in geophysical models. In this study, we propose an approach to simultaneously estimate and interpret J_2 temporal variations based on gravity observations from the GRACE satellite mission. We reconstruct the observed gravity changes by a superimposition of spatial patterns characteristic of individual sources. We find that the seasonal and inter-annual variations are well explained by mass changes in atmosphere, oceans and land water storage. The secular trend, on the other hand, is mainly caused by ice-sheet melt, which has a positive effect (causing the Earth to be flatter), and by the ongoing solid Earth response to past glaciations, which has a negative trend (causing the Earth to be rounder). The trend due to ice-sheet melting during 2002-2017 has a larger absolute value, so that the overall trend is rising, and the Earth is currently becoming flatter.

1 Introduction

J_2 , also known as the Earth’s dynamic oblateness, is a fundamental parameter affecting the Earth’s rotational behavior. Its time variations, ΔJ_2 (hereafter denoted as J_2 for simplicity), are driven by latitudinal large-scale mass redistributions in the Earth system, which are intimately associated with ongoing climate change [e.g., *Cox and Chao*, 2002; *Dickey et al.*, 2002], as well as with geodynamic processes inside the solid Earth [*Benjamin et al.*, 2006].

For more than four decades, J_2 has been monitored by satellite laser ranging (SLR) [*Cheng et al.*, 2013]. In the last 20 years, new data types, platforms and methodologies have become available, and this enables comparisons between different solutions [*Meyrath et al.*, 2017]. Under the assumption that the Earth deforms elastically [*Blewitt*, 2003], seasonal J_2 variations can be derived from GPS-detected solid Earth deformations [*Gross et al.*, 2004; *Lavallée et al.*, 2010] or Earth’s rotation variations [*Chen and Wilson*, 2003]. However, a consideration of only elastic Earth’s properties implies that viscoelastic solid earth signals, such as glacial isostatic adjustment (GIA), are ignored. As a consequence, linear trends in J_2 cannot be reliably estimated. The recently ended Gravity Recovery And Climate Experiment (GRACE) satellite mission monitored changes in the Earth’s gravity field for almost 15 years [*Topley et al.*, 2004]. This includes the observation of J_2 , which is directly related to the degree-2 zonal (C_{20}) coefficient of the gravity field ($J_2 = -\sqrt{5}C_{20}$). Unfortunately, GRACE estimates of J_2 variations are generally less accurate than those based on SLR observations [e.g., *Chen and Wilson*, 2008; *Talpe et al.*, 2017; *Bonin et al.*, 2018],

partly due to thermal-dependent systematic errors in the accelerometers on board the GRACE satellites [Cheng and Ries, 2017]. Sun *et al.* [2016a] have demonstrated that a sufficiently accurate J_2 time series can be obtained from GRACE data supported by an oceanic C_{20} coefficient time series, derived from ocean bottom pressure (OBP) variations by an ocean circulation model [Meyrath *et al.*, 2017]. This approach, which is an extension of the method by Swenson *et al.* [2008], will be from here on referred to as the GRACE-OBP approach. It has been further refined by using better parameter settings [Sun *et al.*, 2016b] and incorporating information about errors in the input datasets [Sun *et al.*, 2017]. It is important to note that the trend estimates obtained with the GRACE-OBP approach contain, among others, the contribution provided by GIA models. Those models, however, are believed to carry large uncertainties.

The J_2 time series is composed not only of a linear trend, but also a prominent annual cycle, as well as other signals with other periods, including long-term ones. The linear trend is mainly due to a combination of GIA effects [Yoder *et al.*, 1983] and ongoing mass loss of the ice sheets [Nerem and Wahr, 2011], while the annual cycle is primarily caused by mass redistribution within atmosphere, oceans and land hydrology. Other signals, including multiple-year (2-7 yr) and decadal variations, are also recognizable [Cheng and Ries, 2018], but still not fully explained due to deficiencies of existing geophysical models [Chen *et al.*, 2016]. Variations at periods longer than one year are especially interesting as the superposition of such signals can be strong enough to temporally reverse the secular trend [Cheng and Tapley, 2004; Cox and Chao, 2002]. A notable example is the “1998 anomaly” [Cox and Chao, 2002]. This anomaly has been reproduced by a superposition of anomalies of land water, oceans and atmosphere variations [Dickey *et al.*, 2002; Lavallée *et al.*, 2010] and by a combination of decadal variations and El Niño-Southern Oscillation (ENSO) related inter-annual variations [Cheng and Tapley, 2004; Cheng and Ries, 2018]. Alternatively, it can be partly attributed to mis-modeling of the Earth body tide (with a period of 18.6 yr) [Benjamin *et al.*, 2006]. Therefore, a definitive explanation of the “1998 anomaly” is still lacking.

A novel way to interpret temporal variations of the Earth’s gravity field was proposed by Rietbroek *et al.* [2012, 2016]. Their approach employs prescribed normalised geoid change patterns associated to regional mass redistributions as basis functions, also known as fingerprints (see more information in Text S1). As input, variations of the Earth’s gravity and sea level are used. The observation of those variations are provided by GRACE and satellite altimetry, respectively. This approach gives insight not only into gravity field variations, but also into regional sea level variations. Furthermore, the authors have demonstrated the ability of this approach to estimate degree-1 variations of the Earth’s gravity field, which are directly related to geocenter motion [Blewitt, 2003].

In this paper, we extend the work of Rietbroek and colleagues in order to retrieve J_2 variations. We show that this approach allows for not only an estimation, but also for direct interpretation of the observed J_2 variations. Such an approach will be referred as the fingerprint approach (FPA) hereafter.

2 Methodology

2.1 Fingerprint database

We define the base functions in line with Rietbroek *et al.* [2016]. Each base function is defined as a geoid fingerprint associated with a particular spatial mass distribution. To that end, the sea level equation on an elastic Earth is solved [Farrell and Clark, 1976; Kendall *et al.*, 2005; Tamisiea *et al.*, 2010]. First, we define two types of mass distribution over land areas: ice-related and hydrology-related ones. The former account for ice mass variations over the Antarctica Ice Sheet (AIS), the Greenland

Ice Sheet (GIS), and glaciers (GLA). They are obtained by splitting land areas into geographical regions, assuming that a unit mass is distributed uniformly over each of those regions. The hydrology-related mass distributions are associated with terrestrial water storage (TWS) variations, which reflect the global water cycle. At that stage, we use the same mass distributions as described in *Rietbroek et al.* [2016]. They were derived from the WaterGAP Hydrology Model (WGHM) [e.g., *Müller Schmied et al.*, 2014, 2016] by applying the empirical orthogonal function (EOF) analysis [*North*, 1984; *Monahan et al.*, 2009]. A unit mass is then distributed following the spatial pattern prescribed by a given EOF mode. In addition, we define GIA-related mass distributions to account for the solid Earth contribution. We do not estimate the atmosphere and dynamic ocean effects (AO), since they are largely removed from GRAE level-2 data product by means of the Atmosphere and Ocean Dealiasing (AOD) product (RL05 or RL06) [*Flechtner and Dobslaw*, 2013; *Dobslaw et al.*, 2017].

We divided the AIS and the GIS into drainage systems following the definition of the Goddard Ice Altimetry Group [*Zwally et al.*, 2012]. The AIS consists of 27 drainage systems (Fig. S1). The GIS is divided into 8 drainage systems, each being further divided into two regions separated by the 2000-m elevation line, for a total of 16 regions (Fig. S2). Each of the regions defined within the AIS and GIS is associated with one fingerprint. For GLA, we use glacier positions archived by the National Snow and Ice Data Center (NSIDC) through the Global Land Ice Measurements from Space (GLIMS) project [*Kargel et al.*, 2014]. We separate the global set of glaciers into 50 groups (Fig. S3). For each group, the mass variation pattern is proportional to the number of glaciers per $1^\circ \times 1^\circ$ grid cell (Fig. S4). Note that we also model glaciers along the coastlines of Greenland and Antarctica, as many of them are not a part of the ice sheet drainage systems. In particular, large residual signals are seen if fingerprints representing the Greenland peripheral glaciers are ignored [*Rietbroek et al.*, 2016]. Over Antarctica, the effect of considering peripheral glaciers is much smaller. An exception is the Totten Glacier. Though located inside a drainage system, this glacier definitely needs a dedicated fingerprint, due to its large mass loss [*Mohajerani et al.*, 2018]. To simplify the analysis of the obtained results, we categorize all the glaciers into three large groups, the Greenland peripheral glaciers (GLA-GRE), the Antarctic peripheral glaciers (GLA-ANT) and the rest of the continental glaciers (GLA-CON). The estimated changes in GLA-GRE and GLA-ANT should be added to GIS and AIS, respectively, to obtain the total contribution of Greenland and Antarctica to J_2 variations. To compute the hydrology-related fingerprints, we select the 60 leading EOFs (explaining over 90% of the signal variance). Additionally, we create fingerprints for the Caspian Sea and the North India plains assuming the uniform mass distribution in both cases. The former addresses inconsistency between the WGHM predictions and GRACE observations; the latter accounts for large TWS changes due to human activities such as groundwater extraction. Both are not (well) represented by the WGHM model.

Finally, we pay special attention to creating GIA fingerprints. In total, we define 6 GIA fingerprints, which are associated with Laurentide, Fennoscandian, Antarctic ice sheets. Based on the ICE-6G ice history model [*Peltier et al.*, 2015], we create 3 regional GIA fingerprints for the Laurentide ice sheet and 2 fingerprints for the Fennoscandian ice sheet (Fig. S5), in order to allow for the possible presence of regional biases in the input GIA model. For Antarctica, we create only one fingerprint for the entire territory. It is based on the IJ05 ice-history [*Ivins and James*, 2005], which has been shown to provide the closest agreement to a data-driven GIA solution [*Riva et al.*, 2009]. The viscosity model used to generate the GIA fingerprints has a higher value in the lower mantle (10^{22}Pas) than VM5a [*Peltier et al.*, 2015], in line with results from a joint inversion of GIA and mantle convection observables [*Mitrovica and Forte*, 1997]. More information about preparing regional GIA models can be found in supporting materials (Text S2). Further more, we include two fingerprints representing

the C_{21} and S_{21} coefficients to account for any possible errors in modeling the Earth’s rotational feedback [e.g., *Mitrovica and Wahr, 2011; Martinec and Hagedoorn, 2014*]. Altogether, we are using 163 fingerprints (Data set S1).

2.2 Inversion approach

Our approach uses the fingerprints as basis functions to represent the geoid changes observed by GRACE. Once the fingerprint database is built, we can estimate the scaling factors α^{f^i} associated to each fingerprint and needed to reproduce the observed geoid changes via:

$$\underset{(k-5) \times k}{\mathbf{T}} \underset{k \times n}{\mathbf{F}} \underset{n \times 1}{\alpha} = \underset{(k-5) \times 1}{\mathbf{g}}, \quad (1)$$

where \mathbf{F} is a matrix of spherical harmonic analysis with n columns. Each column represents an individual fingerprint expressed in the spectral domain with k spherical harmonic coefficients starting from C_{00} (the unique index of each coefficients is given by $l^2 + i \times l + m + 1$, where l and m represent the degree and order, respectively; i equals 0 if $m=0$ and 1 otherwise). \mathbf{g} is the GRACE-based gravity field for a given month starting from C_{21} . The C_{00} , degree-1 and C_{20} coefficients are omitted, so that the length of vector \mathbf{g} is $(k - 5)$. \mathbf{T} is a truncation matrix used to account for the lacking coefficients in \mathbf{g} :

$$\mathbf{T} = \begin{bmatrix} 0 & 0 & 0 & 0 & 1 & 0 & \cdots & \cdots & 0 \\ 0 & 0 & 0 & 0 & 0 & 1 & \ddots & \ddots & \vdots \\ \vdots & \vdots & \vdots & \vdots & \vdots & \ddots & \ddots & \ddots & \vdots \\ \vdots & \vdots & \vdots & \vdots & \vdots & \ddots & \ddots & 1 & 0 \\ 0 & 0 & 0 & 0 & 0 & \cdots & \cdots & 0 & 1 \end{bmatrix}. \quad (2)$$

An explicit form of Eq. (1) can be found in Text S3.

Vector α contains n unknowns (one for each fingerprint in a particular month) that are estimated by least squares:

$$\alpha = (\mathbf{F}^T \mathbf{T}^T \mathbf{P} \mathbf{T} \mathbf{F})^{-1} \mathbf{F}^T \mathbf{T}^T \mathbf{P} \mathbf{g}, \quad (3)$$

where \mathbf{P} is a diagonal weight matrix with elements inversely proportional to the calibrated error variances of GRACE coefficients (downloaded from <ftp://ftp.csr.utexas.edu/outgoing/grace/>). Eq. (3) gives an individual monthly GRACE solution, a time series of α can ultimately be obtained in this way for each fingerprint. It is worth noting that the scaling factors related to surface mass fingerprints are estimated without constrains, whereas those related to the GIA fingerprints need to be constrained in the time domain so that the resulting GIA-related J_2 components consist of only linear trends. An application of this constraint is facilitated by the fact that all coefficients of all months are estimated simultaneously. Readers are referred to the supporting materials for more information (Text S4 and Fig. S6). Once the scaling factors α^{f^i} have been determined, they are used to scale the C_{20} coefficients of each fingerprint. After that, the contribution of individual fingerprints are added together, allowing us to determine monthly variations in J_2 . Note that the resulting J_2 solution is independent from the GRACE C_{20} coefficients, but consistent with the rest of the GRACE solutions.

There is likely a certain degree of correlations between the GIA and the present-day mass transport (PDMT) trends at sub-regional scales, which could prevent an

accurate separation of the different signals. However, PDMT fingerprints are representative of load changes occurring over much smaller regions than those of the GIA fingerprints, meaning that they are largely constrained by signals that are almost completely unrelated to GIA. This is also the reason why we are using a single GIA fingerprint for the whole of Antarctica. The ability of the inversion in *Rietbroek et al.* [2016] to separate GIA from the PDMT signals in Antarctica was limited. For that reason a constraint was applied which penalized large trend deviations from the apriori model. In this study, the applied regularization merely forces the GIA to a straight trendline, but it does not penalize its incline. Surprisingly, the current inversion does not show an instability to simultaneously estimated GIA and PDMT. The exact reasons are yet unknown but we speculate that the use of diagonal error-covariance matrices in combination with the truncated set of Stokes coefficients acts as an implicit regularization w.r.t. to a full error-covariance model. Preliminary experiments based on significantly different GIA fingerprints show that the resulting J_2 trend varies less than 12% (see Text S5 and Table S1 for more information), which suggests that our inversion is fairly stable.

2.3 GRACE data

The Center for Space Research (CSR) currently provides two releases of GRACE solutions, RL05 and the latest one, RL06. Compared to RL05, RL06 is based on new background models, including an updated AOD data product, and new parameterization strategies. The AOD RL05 is based on the atmosphere model by the European Center for Medium-Range Weather Forecast (ECMWF) [*Dee et al.*, 2011] and the ocean model for circulation and tides (OMCT) [*Thomas*, 2002]. The AOD RL06 [*Dobslaw et al.*, 2017] makes use of the same atmosphere model, but is based on a different ocean model, i.e., the Max-Planck-Institute for Meteorology Ocean model (MPIOM) [*Jungclaus et al.*, 2013]. In this study, we use both the CSR RL05 [*Bettadpur*, 2012] and CSR RL06 [*Save*, 2018] models completed to degree and order 60. In the RL05, the available solutions are from 2002 April to 2017 June, while in the RL06 the available period is approximately two years shorter (2003 January to 2016 August). In the main text, we mainly show the results based on RL06, while the RL05 results are similar and can be found in the Supplemental material. A mean gravity field in the period covered by the GRACE data has been removed to isolate time-variations of the gravity field. We did not apply any filtering/smoothing techniques. The pole tide correction [*Wahr et al.*, 2015] is not applied as it has negligible effects on the resulting J_2 solutions. Also, such a correction is no longer needed for the CSR RL06 (personal communication by Felix Landerer).

3 Results

We show in Fig. 1a the global geoid trend estimated from the original CSR RL06 solutions, in Fig. 1b the trend reconstructed by the fingerprints, and in Fig. 1c the residuals between panel a and b. Additionally, we show in the right column (panels d, e and f) the same results, represented in terms of equivalent water heights (EWH). The residual signals are rather small except for co-seismic signals from megathrust earthquakes, such as the 2004 Sumatra Earthquake and the 2011 Tohoku Earthquake, as well as the signal around the Kamb Ice Stream in West Antarctica. These are expected because no earthquake-related fingerprints are prescribed in the database, and because the large gradients in West Antarctica may not be well represented by uniform mass variations assigned to the drainage systems there. However, these signals only have minor effects on J_2 for their limited spatial scales. The only visible difference at large spatial scales is the band-shaped signal in Fig. 1c, which shows that the inverted C_{20} coefficient is considerably different from the original GRACE C_{20} (RMS

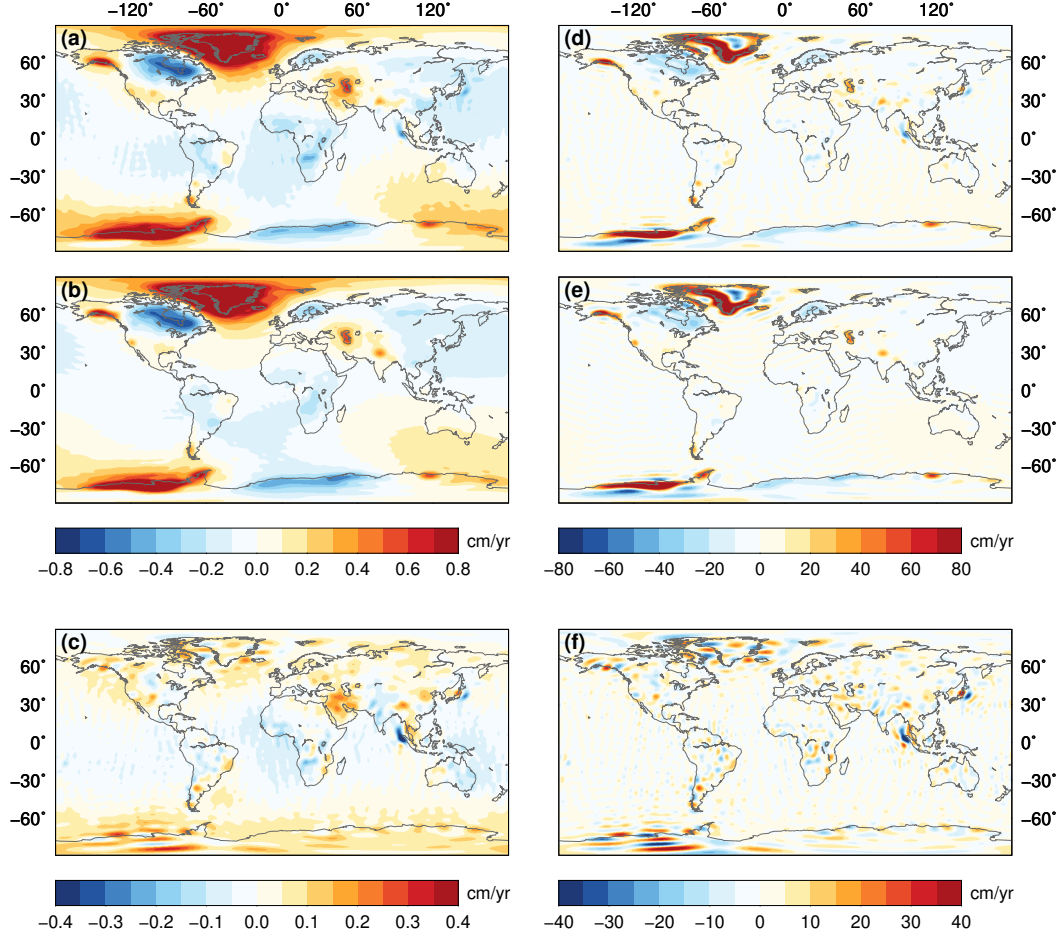


Figure 1. Global trend map in terms of geoid (panels a, b, c) and EWH (panels d, e, f). The top panels (a, d) show the results recovered with the original CSR RL06 solution. The middle panels (b, e) show those based on the fingerprint approach, while the bottom panels (c, f) show the residual signals between the two solutions (a-b and d-e). Notice the different color scales.

of residuals is 1.7×10^{-10} , Fig. S7). The results for CSR RL05 are rather similar, but with more noise (RMS of residuals is 3.0×10^{-10} , Fig. S7 and S8).

In Fig. 2 we plot the FPA and GRACE-OBP [Sun *et al.*, 2016b] estimates of J_2 based on CSR RL06 solutions, as well as the estimates based on SLR data [Cheng *et al.*, 2013]. A plot showing the estimates based on the CSR RL05 can be found in the supporting materials (Fig. S9). We calculate uncertainties of the FPA solution using the method developed by Ditmar *et al.* [2018]. The $1\text{-}\sigma$ interval (1.01×10^{-10}) is shown with black error bars: it is slightly larger than the RL05 version ($\sigma = 0.97 \times 10^{-10}$) and comparable with the SLR uncertainty, which has a mean value of 0.83×10^{-10} (computed from the uncertainties shown in the data file ftp://ftp.csr.utexas.edu/pub/slr/degree_2/RL06/C20_RL06_2018_04.txt). The differences between FPA and GRACE-OBP estimates (RMS: 0.44×10^{-10}) are smaller than those between FPA and SLR solution (RMS: 1.01×10^{-10}), indicating that the two GRACE-based estimates are consistent. Also, we notice that there is a

linear trend left in the FPA/SLR differences from 2002 to 2008. The reasons for it remain unknown.

All three J_2 solutions show dominant annual cycles. The annual amplitude of the FPA solution is virtually identical to that of the SLR solution (Table S2). In contrast, the annual amplitude of the GRACE-OBP solution is significantly larger (about 17%) and statistically different, though larger values are still in line with some other SLR estimates [Cox and Chao, 2002; Chen and Wilson, 2008]. All three solutions also show significant semi-annual variations (Table S2). They are comparable to each other within $2\text{-}\sigma$, as well as with those from some previous studies [Lavallée et al., 2010]. Phase estimates for both annual and semi-annual terms are insensitive to the time interval and consistent with each other within $1\text{-}\sigma$ (annual) and $2\text{-}\sigma$ (semi-annual) range, which is about one and two weeks, respectively.

One of the main advantages of the fingerprint approach is that it allows one to separate the contributions of individual sources to the total signal. Here we separate the ice sheets, the glaciers, global land hydrology, and GIA (Fig. 3). The AO contribution is not estimated by the FPA method, but predicted by the AOD RL06 product. Note that it is almost the same as the one from RL05 (Fig. S10), though they are based on different ocean circulation models. Clearly, the seasonal signals in J_2 are controlled by two components, AO and TWS. AO is the largest one with an annual amplitude of $1.85 \pm 0.10 \times 10^{-10}$, while TWS is slightly smaller with an annual amplitude of $1.27 \pm 0.07 \times 10^{-10}$. The combination of these two components explains almost the entire seasonal variations in J_2 (98% of the total signal variance; Table S3). However, both components have negligible contributions to the trend estimates over the whole time-span (though a significant TWS trend is present over shorter periods).

It is clear that the rising trend in J_2 is mainly due to mass loss from GIS + GLA-GRE (0.44×10^{-10}). AIS + GLA-ANT also shows a positive trend of 0.07×10^{-10} . GLA-CON, on the other hand, does not provide a significant contribution to the trend in J_2 . Consistently with the fact that present-day GIA is causing large-scale uplifts at high latitudes, we see negative trends in all GIA-related sources, with the Laurentide GIA providing the largest contribution (-0.16×10^{-10}), followed by the Antarctica GIA (0.08×10^{-10}) and Fennoscandian GIA (-0.03×10^{-10}). In addition, the scaling factors for GIA fingerprints are provided in Data set S2, the linear trend in those scaling factors are listed in Table S4.

In order to investigate inter-annual signals, we have computed 11-month running averages of the SLR, GRACE-OBP and FPA detrended solutions. In Fig. 4a, we see that the two GRACE-based solutions are close. The SLR solution is also generally aligned with the other two except for discrepancies during years 2009-2011. The AO component (Fig. 3) explains a considerable part of the observed variability, which is particularly strong during 2008-2010 due to a major ENSO event [Cheng and Ries, 2018], but certainly cannot account for all the variations. In Fig. 4b, we show the residual J_2 time series (FPA-AO), i.e., the one obtained prior to the restoration of the AO contribution, which is common to all solutions. Removing the AO contribution highlights even further the discrepancy between SLR and GRACE-based solutions during years 2009-2011. Most of the residual variability is well explained by the TWS component (Fig. 4c); the correlation coefficient between the FPA-AO and the TWS is as high as 0.85. We further compare the FPA TWS component with predictions based on three hydrological models: the Global Land Data Assimilation System NOAH model (GLDAS-NOAH V1.0) [Rodell et al., 2004], the Climate Prediction Center model (CPC) [Dool et al., 2003; Fan and Dool, 2004] and the WGHM [Müller Schmied et al., 2014, 2016] (Fig. 4d, Fig. S11). Note that the TWS fingerprints are based on an EOF analysis of WGHM, while the resulting amplitude is derived from GRACE data. But none of the hydrological models reproduces some features of the FPA solution, such as the low in 2005.

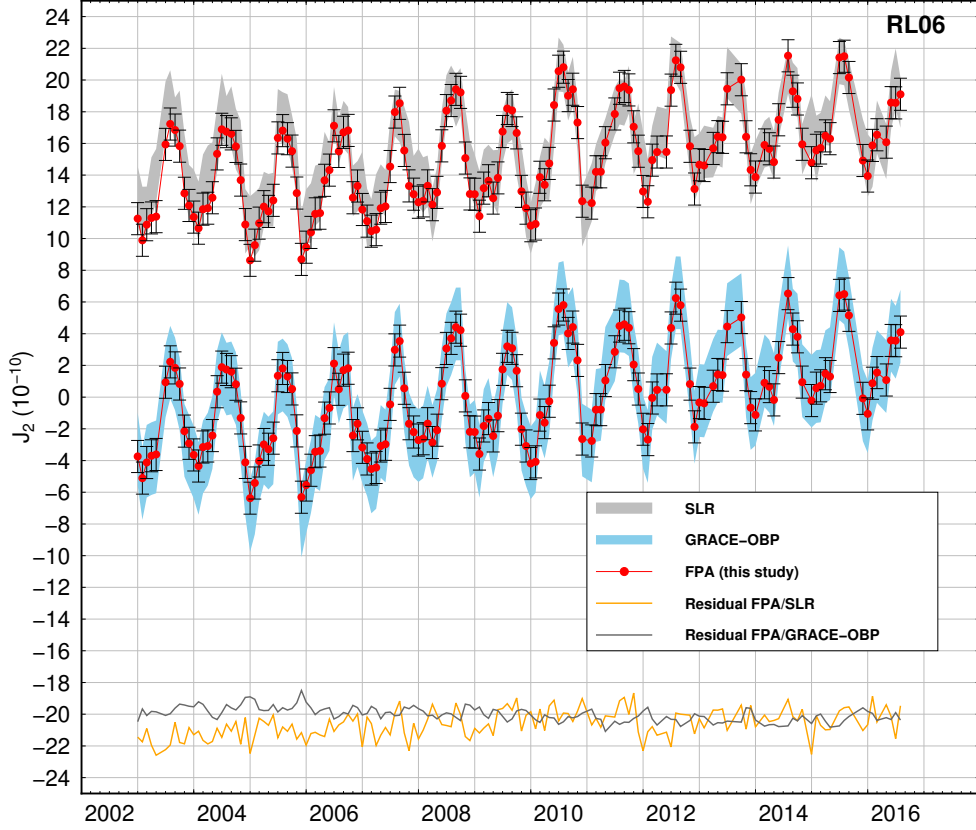


Figure 2. A comparison of J_2 estimates obtained from CSR RL06 data using the FPA approach (this study, red curve with dots) with those obtained using the GRACE-OBP approach *Sun et al.* [2016b] (blue band), as well as with those based on SLR data *Cheng et al.* [2013] (gray band). An offset of (15×10^{-10}) is applied to the top pair of curves. Note that the SLR solution is represented by a gray band where the upper and lower bounds are defined by the $\mu \pm 2\sigma$ (μ denotes the mean values), respectively. The same is also applied to the GRACE-OBP solution, where the uncertainty ($\sigma = 1.13 \times 10^{-10}$) is calculated based on *Ditmar et al.* [2018]. The error bars are also shown for the FPA solution. All three solutions represent the full J_2 signals, which means that the J_2 contributions from the atmosphere and dynamic ocean effects have been restored in the GRACE-based solutions. The differences between FPA and SLR as well as between FPA and GRACE-OBP are shown at the bottom of the plot with an offset of -20×10^{-10} being applied for clarity. The time series are based on the RL06 solutions, meaning that AOD RL06 product is adopted, the SLR solution uses AOD RL06 as background model. A RL05 version is available in the supporting materials (Fig. S9).

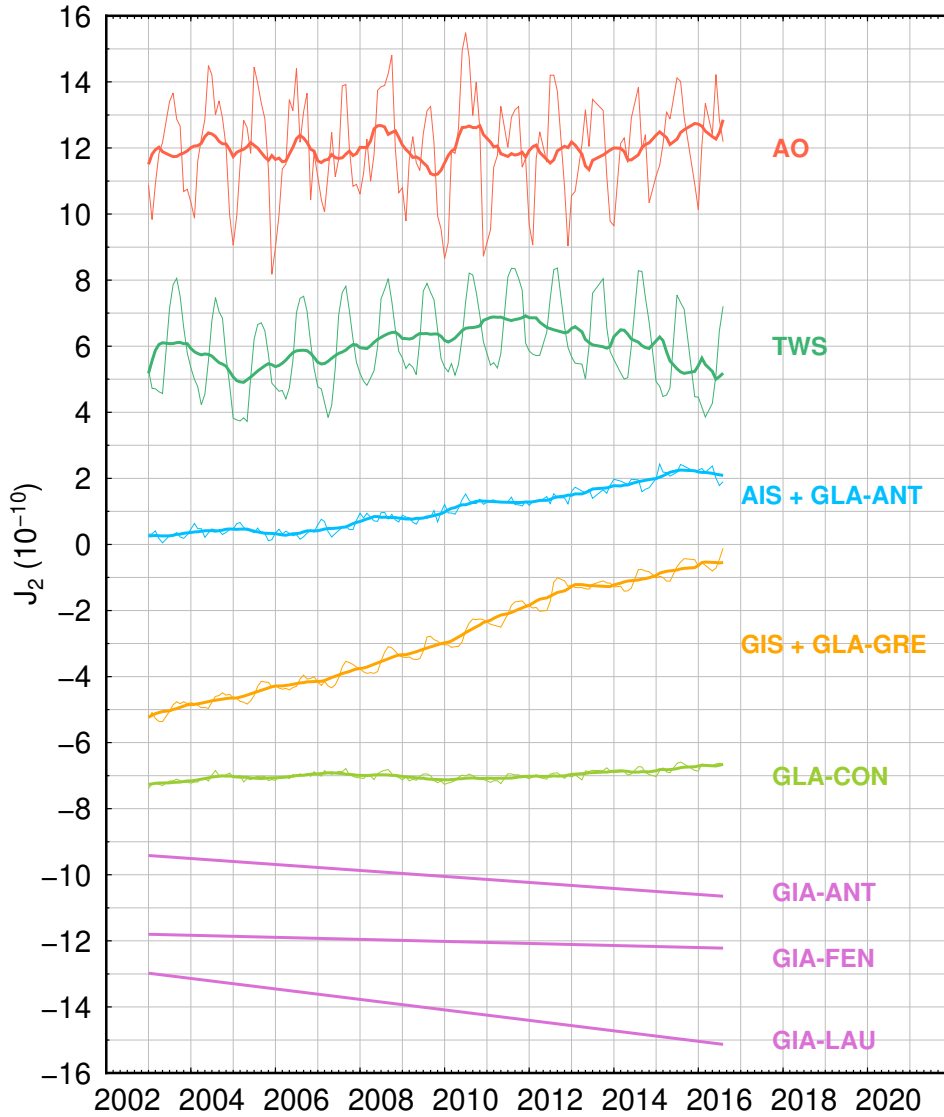


Figure 3. Contributions to the RL6 J_2 time-series estimated via the fingerprint approach. Offsets are applied for clarity. Note that the AO component is not estimated but directly taken from the AOD product. AIS + GLA-ANT and GRE + GLA-GRE represent the combined contributions of ice sheets and peripheral glaciers of Antarctica and Greenland, respectively. GLA-CON denotes the other continental glaciers. GIA-ANT, GIA-FEN and GIA-LAU refer to the GIA effects due to Antarctica, Fennoscandian and Laurentide ice sheets, respectively. Thick lines on top of the non-GIA components are the 11-month running means.

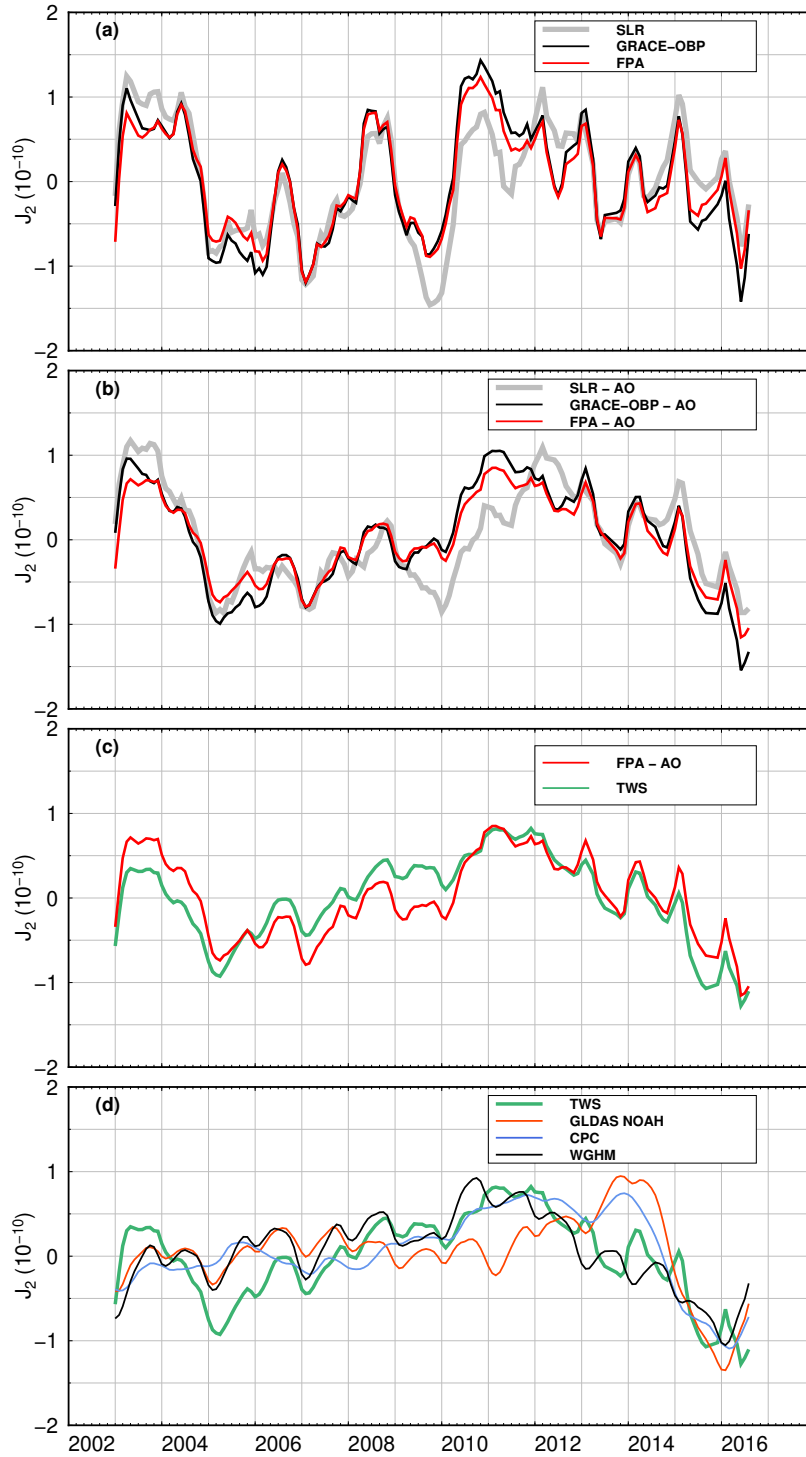


Figure 4. Inter-annual variations of the J_2 RL06 time series. Panel (a) shows the comparison of J_2 solutions based on SLR data, GRACE-OBP approach and this study. Note that these are the full J_2 time series with the AO component restored. Panel (b) shows the residual J_2 signals without restoring the AO component. Panel (c) compares the residual J_2 of this study with one of its contributing sources, i.e., TWS. Panel (d) compares the J_2 TWS component with those based on 3 hydrology models (a comparison with these three hydrological models at the seasonal time scale can be found in Text S5 and Fig. S11).

4 Discussions and Conclusions

The FPA is an effective approach to both explain and retrieve J_2 time series. According to the method proposed by *Ditmar et al.* [2018], the error in the FPA solution (approximately $\pm 1.0 \times 10^{-10}$) is comparable to that of the SLR solution [*Cheng et al.*, 2013]. Our solution suggests that the primary contributors to seasonal and inter-annual variations in J_2 over the GRACE period (2002 - 2017) are the AO and TWS component. The positive trend during this particular time-interval is mainly due to mass loss from the ice sheets, particularly from Greenland, while contributions from mountain glaciers are minor.

The uniqueness of the FPA compared to other methods is that it helps understanding the driving mechanisms of the observed time series. The traditional way of interpreting J_2 variations relies on the reconstruction of the observed signal by means of combining various of geophysical models. Unfortunately, some of the Earth's components, e.g., mountain glaciers, are difficult to model due to rather sparse data constraints. Models for other components, such as GIA, often carry significant and unknown uncertainties. Also, there is no consensus on the choice of a hydrological model. Though many available, they typically lack some signals (particularly, of an anthropogenic origin). This limits their ability to explain all the mass variations observed with independent methods. Also, as noted by *Lavallée et al.* [2010] and by *Scanlon et al.* [2018], geophysical models may be particularly weak in reconstructing multiple-year variations. Our study also confirms a limited accuracy of existing hydrological models. We found that those models cannot fully explain the TWS contribution to our J_2 solution. Therefore, the estimated J_2 time series of different components from this study may be used to validate geophysical models in the future studies.

One may argue that our TWS component is estimated using base functions extracted from a hydrological model and thus should inherently carry uncertainties of that model. However, the FPA approach does not need a very accurate hydrology model to work because it re-scales the original signals on a monthly basis. As long as a hydrology model can predict the major mass change patterns correctly, it should result in quite good J_2 solutions. The fact that we are only using the 60 leading EOFs is equivalent to a low pass filtering the hydrology model. Therefore, the impact of missing small-scale mass changes is likely minor. We have verified, for instance, that removing the fingerprint for India plains, which is known to suffer from serious groundwater depletion, does not significantly alter the J_2 time series (RMS difference for the TWS-induced J_2 is 0.02×10^{-10}).

Finally, we have co-estimated the geocenter motion (i.e., variations of degree-1 coefficients) simultaneously with C_{20} coefficient time series and compared it with the GRACE-OBP solution (Table S5 and Fig. S12). A high accuracy of the GRACE-OBP solution has been proved by a few previous studies [*Sun et al.*, 2016b, 2017; *Zhang and Sun*, 2018]. Therefore, the close agreement of the two solutions provides a further support for the FPA approach.

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<https://disc.gsfc.nasa.gov/datasets?keywords=GLDAS>). Fingerprints used in this study as well as their corresponding scaling factors for each months are submitted as data set in the supplementary materials. All results from this study will be made publicly available upon publication through the 4TU.Centre for Research Data (<https://researchdata.4tu.nl/en/home/>).

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References

- Benjamin, D., J. Wahr, R. D. Ray, G. D. Egbert, and S. D. Desai (2006), Constraints on mantle anelasticity from geodetic observations, and implications for the J_2 anomaly, *Geophysical Journal International*, *165*(1), 3–16, doi:10.1111/j.1365-246X.2006.02915.x.
- Bettadpur, S. (2012), UTCSR Level-2 Processing Standards Document, *Technical Version 4*, Univ. Texas, Austin.
- Blewitt, G. (2003), Self-consistency in reference frames, geocenter definition, and surface loading of the solid Earth, *Journal of Geophysical Research: Solid Earth*, *108*(B2), 2103, doi:10.1029/2002JB002082.
- Bonin, J. A., D. P. Chambers, and M. Cheng (2018), Using satellite laser ranging to measure ice mass change in Greenland and Antarctica, *The Cryosphere*, *12*(1), 71–79, doi:10.5194/tc-12-71-2018.
- Chen, J. L., and C. R. Wilson (2003), Low degree gravitational changes from earth rotation and geophysical models, *Geophysical Research Letters*, *30*(24), 2257, doi:10.1029/2003GL018688.
- Chen, J. L., and C. R. Wilson (2008), Low degree gravity changes from GRACE, Earth rotation, geophysical models, and satellite laser ranging, *Journal of Geophysical Research: Solid Earth*, *113*(B6), B06402, doi:10.1029/2007JB005397.
- Chen, J. L., C. R. Wilson, and J. C. Ries (2016), Broadband assessment of degree-2 gravitational changes from GRACE and other estimates, 2002–2015, *Journal of Geophysical Research: Solid Earth*, *121*(3), 2112–2128, doi:10.1002/2015JB012708.
- Cheng, M., and J. Ries (2017), The unexpected signal in GRACE estimates of C_{20} , *Journal of Geodesy*, *91*(8), 897–914, doi:10.1007/s00190-016-0995-5.
- Cheng, M., and J. C. Ries (2018), Decadal variation in Earth’s oblateness (J_2) from satellite laser ranging data, *Geophysical Journal International*, *212*(2), 1218–1224, doi:10.1093/gji/ggx483.
- Cheng, M., and B. D. Tapley (1999), Seasonal variations in low degree zonal harmonics of the Earth’s gravity field from satellite laser ranging observations, *Journal of Geophysical Research: Solid Earth*, *104*(B2), 2667–2681, doi:10.1029/1998JB900036, 00059.
- Cheng, M., and B. D. Tapley (2004), Variations in the Earth’s oblateness during the past 28 years, *Journal of Geophysical Research: Solid Earth*, *109*(B9), B09402, doi:10.1029/2004JB003028.
- Cheng, M., B. D. Tapley, and J. C. Ries (2013), Deceleration in the Earth’s oblateness, *Journal of Geophysical Research: Solid Earth*, *118*(2), 740–747, doi:10.1002/jgrb.50058, 00011.
- Cox, C. M., and B. F. Chao (2002), Detection of a Large-Scale Mass Redistribution in the Terrestrial System Since 1998, *Science*, *297*(5582), 831–833, doi:10.1126/science.1072188.
- Dee, D. P., S. M. Uppala, A. J. Simmons, P. Berrisford, P. Poli, S. Kobayashi, U. Andrae, M. A. Balmaseda, G. Balsamo, P. Bauer, P. Bechtold, A. C. M. Beljaars, L. van de Berg, J. Bidlot, N. Bormann, C. Delsol, R. Dragani, M. Fuentes, A. J. Geer, L. Haimberger, S. B. Healy, H. Hersbach, E. V. Hólm, L. Isaksen, P. Kållberg, M. Köhler, M. Matricardi, A. P. McNally, B. M. Monge-Sanz, J.-J. Morcrette, B.-K. Park, C. Peubey, P. de Rosnay, C. Tavolato, J.-N. Thépaut, and F. Vitart (2011), The ERA-Interim reanalysis: configuration and performance of the data assimilation system, *Quarterly Journal of the Royal Meteorological Society*, *137*(656), 553–597, doi:10.1002/qj.828.
- Dickey, J. O., S. L. Marcus, O. de Viron, and I. Fukumori (2002), Recent Earth oblateness variations: unraveling climate and postglacial rebound effects, *Science (New York, N.Y.)*, *298*(5600), 1975–1977, doi:10.1126/science.1077777.

- Ditmar, P., N. Tangdamrongsub, J. Ran, and R. Klees (2018), Estimation and reduction of random noise in mass anomaly time-series from satellite gravity data by minimization of month-to-month year-to-year double differences, *Journal of Geodynamics*, *119*, 9–22, doi:10.1016/j.jog.2018.05.003.
- Dobslaw, H., I. Bergmann-Wolf, R. Dill, L. Poropat, M. Thomas, C. Dahle, S. Esselborn, R. König, and F. Flechtner (2017), A new high-resolution model of non-tidal atmosphere and ocean mass variability for de-aliasing of satellite gravity observations: AOD1b RL06, *Geophysical Journal International*, *211*(1), 263–269, doi:10.1093/gji/ggx302.
- Dool, H. v. d., J. Huang, and Y. Fan (2003), Performance and analysis of the constructed analogue method applied to U.S. soil moisture over 1981–2001, *Journal of Geophysical Research: Atmospheres*, *108*(D16), doi:10.1029/2002JD003114.
- Fan, Y., and H. v. d. Dool (2004), Climate Prediction Center global monthly soil moisture data set at 0.5 resolution for 1948 to present, *Journal of Geophysical Research: Atmospheres*, *109*(D10), doi:10.1029/2003JD004345.
- Farrell, W. E., and J. A. Clark (1976), On Postglacial Sea Level, *Geophysical Journal of the Royal Astronomical Society*, *46*(3), 647–667, doi:10.1111/j.1365-246X.1976.tb01252.x.
- Flechtner, F., and H. Dobslaw (2013), AOD1b Product Description Document for Product Release. 05, *GFZ German Research Centre for Geosciences*.
- Gross, R. S., G. Blewitt, P. J. Clarke, and D. Lavallée (2004), Degree-2 harmonics of the Earth’s mass load estimated from GPS and Earth rotation data, *Geophysical Research Letters*, *31*(7), L07,601, doi:10.1029/2004GL019589.
- Ivins, E. R., and T. S. James (2005), Antarctic glacial isostatic adjustment: a new assessment, *Antarctic Science*, *17*(04), 541–553, doi:10.1017/S0954102005002968.
- Jungclaus, J. H., N. Fischer, H. Haak, K. Lohmann, J. Marotzke, D. Matei, U. Mikolajewicz, D. Notz, and J. S. v. Storch (2013), Characteristics of the ocean simulations in the Max Planck Institute Ocean Model (MPIOM) the ocean component of the MPI-Earth system model, *Journal of Advances in Modeling Earth Systems*, *5*(2), 422–446, doi:10.1002/jame.20023.
- Kargel, J. S., G. J. Leonard, M. P. Bishop, A. Kääb, and B. H. Raup (Eds.) (2014), *Global Land Ice Measurements from Space*, Geophysical Sciences, Springer-Verlag, Berlin Heidelberg.
- Kendall, R. A., J. X. Mitrovica, and G. A. Milne (2005), On post-glacial sea level – II. Numerical formulation and comparative results on spherically symmetric models, *Geophysical Journal International*, *161*(3), 679–706, doi:10.1111/j.1365-246X.2005.02553.x.
- Lavallée, D. A., P. Moore, P. J. Clarke, E. J. Petrie, T. van Dam, and M. A. King (2010), J2: An evaluation of new estimates from GPS, GRACE, and load models compared to SLR, *Geophysical Research Letters*, *37*(22), doi:10.1029/2010GL045229, 00007.
- Martinec, Z., and J. Hagedoorn (2014), The rotational feedback on linear-momentum balance in glacial isostatic adjustment, *Geophysical Journal International*, *199*(3), 1823–1846, doi:10.1093/gji/ggu369.
- Meyrath, T., P. Rebeschung, and T. van Dam (2017), GRACE era variability in the Earth’s oblateness: a comparison of estimates from six different sources, *Geophysical Journal International*, *208*(2), 1126–1138, doi:10.1093/gji/ggw441.
- Meyssignac, B., J. M. Lemoine, M. Cheng, A. Cazenave, P. Gégout, and P. Maisongrande (2013), Interannual variations in degree-2 Earth’s gravity coefficients C2,0, C2,2, and S2,2 reveal large-scale mass transfers of climatic origin, *Geophysical Research Letters*, *40*(15), 4060–4065, doi:10.1002/grl.50772.
- Mitrovica, J. X., and A. M. Forte (1997), Radial profile of mantle viscosity: Results from the joint inversion of convection and postglacial rebound observables, *Journal of Geophysical Research: Solid Earth*, *102*(B2), 2751–2769, doi:10.1029/96JB03175.

- Mitrovica, J. X., and J. Wahr (2011), Ice Age Earth Rotation, *Annual Review of Earth and Planetary Sciences*, *39*(1), 577–616, doi:10.1146/annurev-earth-040610-133404.
- Mohajerani, Y., I. Velicogna, and E. Rignot (2018), Mass Loss of Totten and Moscow University Glaciers, East Antarctica, Using Regionally Optimized GRACE Mascons, *Geophysical Research Letters*, *45*(14), 7010–7018, doi:10.1029/2018GL078173.
- Monahan, A. H., J. C. Fyfe, M. H. P. Ambaum, D. B. Stephenson, and G. R. North (2009), Empirical Orthogonal Functions: The Medium is the Message, *Journal of Climate*, *22*(24), 6501–6514, doi:10.1175/2009JCLI3062.1.
- Moore, P., Q. Zhang, and A. Alothman (2005), Annual and semiannual variations of the Earth’s gravitational field from satellite laser ranging and CHAMP, *Journal of Geophysical Research: Solid Earth*, *110*(B6), doi:10.1029/2004JB003448, 00011.
- Müller Schmied, H., S. Eisner, D. Franz, M. Wattenbach, F. T. Portmann, M. Flörke, and P. Döll (2014), Sensitivity of simulated global-scale freshwater fluxes and storages to input data, hydrological model structure, human water use and calibration, *Hydrology and Earth System Sciences*, *18*(9), 3511–3538.
- Müller Schmied, H., L. Adam, S. Eisner, G. Fink, M. Flörke, H. Kim, T. Oki, F. T. Portmann, R. Reinecke, C. Riedel, and others (2016), Variations of global and continental water balance components as impacted by climate forcing uncertainty and human water use, *Hydrology and Earth System Sciences*, *20*(7), 2877–2898.
- Nerem, R. S., and J. Wahr (2011), Recent changes in the Earth’s oblateness driven by Greenland and Antarctic ice mass loss, *Geophysical Research Letters*, *38*(13), L13,501, doi:10.1029/2011GL047879.
- North, G. R. (1984), Empirical Orthogonal Functions and Normal Modes, *Journal of the Atmospheric Sciences*, *41*(5), 879–887, doi:10.1175/1520-0469(1984)041<0879:EOFANM>2.0.CO;2.
- Peltier, W. (2004), Global glacial isostasy and the surface of the ice-age Earth: the ICE-5g (VM2) model and GRACE, *Annual Review of Earth and Planetary Sciences*, *32*(1), 111–149, doi:10.1146/annurev.earth.32.082503.144359.
- Peltier, W. R., D. F. Argus, and R. Drummond (2015), Space geodesy constrains ice age terminal deglaciation: The global ICE-6g.c (VM5a) model, *Journal of Geophysical Research: Solid Earth*, *120*(1), 2014JB011,176, doi:10.1002/2014JB011176.
- Rietbroek, R., S. E. Brunnabend, J. Kusche, and J. Schröter (2012), Resolving sea level contributions by identifying fingerprints in time-variable gravity and altimetry, *Journal of Geodynamics*, *59–60*, 72–81, doi:10.1016/j.jog.2011.06.007, 00013.
- Rietbroek, R., S.-E. Brunnabend, J. Kusche, J. Schröter, and C. Dahle (2016), Revisiting the contemporary sea-level budget on global and regional scales, *Proceedings of the National Academy of Sciences*, *113*(6), 1504–1509, doi:10.1073/pnas.1519132113.
- Riva, R. E. M., B. C. Gunter, T. J. Urban, B. L. A. Vermeersen, R. C. Lindenbergh, M. M. Helsen, J. L. Bamber, R. S. W. van de Wal, M. R. van den Broeke, and B. E. Schutz (2009), Glacial Isostatic Adjustment over Antarctica from combined ICESat and GRACE satellite data, *Earth and Planetary Science Letters*, *288*(3–4), 516–523, doi:10.1016/j.epsl.2009.10.013.
- Rodell, M., P. R. Houser, U. Jambor, J. Gottschalck, K. Mitchell, C.-J. Meng, K. Arsenault, B. Cosgrove, J. Radakovich, M. Bosilovich, J. K. Entin, J. P. Walker, D. Lohmann, and D. Toll (2004), The Global Land Data Assimilation System, *Bulletin of the American Meteorological Society*, *85*(3), 381–394, doi:10.1175/BAMS-85-3-381.
- Save, H. (2018), GRACE RL06 Reprocessing and Results from CSR, *EGU General Assembly, Vienna, Austria*.
- Scanlon, B. R., Z. Zhang, H. Save, A. Y. Sun, H. Müller Schmied, L. P. H. van Beek, D. N. Wiese, Y. Wada, D. Long, R. C. Reedy, L. Longuevergne, P. Döll, and M. F. P. Bierkens (2018), Global models underestimate large decadal declining and rising water storage trends relative to GRACE satellite data, *Proceedings of the*

- National Academy of Sciences*, doi:10.1073/pnas.1704665115.
- Seo, K.-W., J. Chen, C. R. Wilson, and C.-K. Lee (2015), Decadal and quadratic variations of Earth's oblateness and polar ice mass balance from 1979 to 2010, *Geophysical Journal International*, *203*(1), 475–481, doi:10.1093/gji/ggv312.
- Sun, Y., P. Ditmar, and R. Riva (2016a), Observed changes in the Earth's dynamic oblateness from GRACE data and geophysical models, *Journal of Geodesy*, *90*(1), 81–89, doi:10.1007/s00190-015-0852-y.
- Sun, Y., R. Riva, and P. Ditmar (2016b), Optimizing estimates of annual variations and trends in geocenter motion and J_2 from a combination of GRACE data and geophysical models, *Journal of Geophysical Research: Solid Earth*, *121*(11), 8352–8370, doi:10.1002/2016JB013073.
- Sun, Y., P. Ditmar, and R. Riva (2017), Statistically optimal estimation of degree-1 and C_{20} coefficients based on GRACE data and an ocean bottom pressure model, *Geophysical Journal International*, *210*(3), 1305–1322, doi:10.1093/gji/ggx241.
- Swenson, S., D. Chambers, and J. Wahr (2008), Estimating geocenter variations from a combination of GRACE and ocean model output, *Journal of Geophysical Research: Solid Earth*, *113*(B8), B08,410, doi:10.1029/2007JB005338.
- Talpe, M. J., R. S. Nerem, E. Forootan, M. Schmidt, F. G. Lemoine, E. M. Enderlin, and F. W. Landerer (2017), Ice mass change in Greenland and Antarctica between 1993 and 2013 from satellite gravity measurements, *Journal of Geodesy*, *91*(11), 1283–1298, doi:10.1007/s00190-017-1025-y.
- Tamisiea, M. E., E. M. Hill, R. M. Ponte, J. L. Davis, I. Velicogna, and N. T. Vinogradova (2010), Impact of self-attraction and loading on the annual cycle in sea level, *Journal of Geophysical Research: Oceans*, *115*(C7), C07,004, doi:10.1029/2009JC005687.
- Tapley, B. D., S. Bettadpur, M. Watkins, and C. Reigber (2004), The gravity recovery and climate experiment: Mission overview and early results, *Geophysical Research Letters*, *31*(9), L09,607, doi:10.1029/2004GL019920.
- Thomas, M. (2002), Ocean induced variations of Earth's rotation results from a simultaneous model of global circulation and tides, Ph.D. thesis, Univ. of Hamburg, Germany.
- Wahr, J., R. S. Nerem, and S. V. Bettadpur (2015), The pole tide and its effect on GRACE time-variable gravity measurements: Implications for estimates of surface mass variations, *Journal of Geophysical Research: Solid Earth*, *120*(6), 2015JB011,986, doi:10.1002/2015JB011986.
- Yoder, C. F., J. G. Williams, J. O. Dickey, B. E. Schutz, R. J. Eanes, and B. D. Tapley (1983), Secular variation of Earth's gravitational harmonic J_2 coefficient from Lageos and nontidal acceleration of Earth rotation, *Nature*, *303*(5920), 757–762, doi:10.1038/303757a0.
- Zhang, H., and Y. Sun (2018), Climate-driven seasonal geocenter motion during the GRACE period, *Acta Geophysica*, *66*(2), 223–232, doi:10.1007/s11600-018-0130-5.
- Zwally, H. J., M. B. Giovinetto, M. A. Beckley, and J. L. Saba (2012), Antarctic and Greenland drainage systems, *GSFC Cryospheric Sciences Laboratory*.