Temporal Variations of the Marine Geoid

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Abstract

The effects of temporal changes in the marine geoid on estimates of the ocean dynamic topography is being investigated. Influences from changes of land hydrology, ice sheets, Post-Glacial Rebound (PGR), and ocean and atmospheric dynamics are considered and the associated crustal deformation is included. The strongest signals are associated with the seasonal cycle caused by changes in terrestrial water content and ice sheets as well as the redistribution of atmospheric mass. Second to this is the importance of an overall trend caused by PGR and decreasing ice sheets over Greenland and Antarctica. On long spatial scales, the amplitude of regional trends estimated for the geoid height have a sizable fraction of those from Sea Level Anomaly (SLA) for the period 1993–2016, also after subtraction of steric height of the upper 1000m to analyze trends in deep ocean geostrophic currents. The estimated strong negative geoid height trend south of Greenland for the period 1993–2016 opposes changes in dynamic sea level for the same period thereby affecting past studies on changes of both the strength of the Subpolar Gyre based on SLA and the meridional overturning circulation on a section between Cape Farewell and Portugal applying ocean dynamic topography and hydrography. We conclude that temporal geoid height trends should be considered in studies of (multi-)decadal trends in sea level and circulation on large spatial scales based on altimetry data referenced to a geoid field.

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Key Points:

- The strongest geoid height changes are associated with regionally pronounced seasonal signals and secular trends
- In the Subpolar North Atlantic the geoid height trend biases circulation trend estimates based on altimetry
- Altimetry data needs correction for geoid height change when long term variations in ocean dynamics are studied

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Abstract

The effects of temporal changes in the marine geoid on estimates of the ocean dynamic topography is being investigated. Influences from changes of land hydrology, ice sheets, Post-Glacial Rebound (PGR), and ocean and atmospheric dynamics are considered and the associated crustal deformation is included. The strongest signals are associated with the seasonal cycle caused by changes in terrestrial water content and ice sheets as well as the redistribution of atmospheric mass. Second to this is the importance of an overall trend caused by PGR and decreasing ice sheets over Greenland and Antarctica. On long spatial scales, the amplitude of regional trends estimated for the geoid height have a sizable fraction of those from Sea Level Anomaly (SLA) for the period 1993–2016, also after subtraction of steric height of the upper 1000m to analyze trends in deep ocean geostrophic currents. The estimated strong negative geoid height trend south of Greenland for the period 1993–2016 opposes changes in dynamic sea level for the same period thereby affecting past studies on changes of both the strength of the Subpolar Gyre based on SLA and the meridional overturning circulation on a section between Cape Farewell and Portugal applying ocean dynamic topography and hydrography. We conclude that temporal geoid height trends should be considered in studies of (multi-)decadal trends in sea level and circulation on large spatial scales based on altimetry data referenced to a geoid field.

Plain Language Summary

Changes in ocean surface currents are routinely obtained from satellite altimetry data. A correction for changes in the geoid, the equipotential surface of gravity closest to sea level, is considered small and thus usually neglected. We investigate temporal geoid height changes and potential implications on ocean circulation studies using space-borne gravity data and results from ocean and atmosphere models to discover the individual processes of mass redistribution in the climate system causing thereby changes in the geoid height. We found the largest signals in the seasonal cycle for terrestrial hydrology in the Amazone basin, and in negative trends for the Greenland and West Antarctic Ice sheets. For the period 1993-2016 and on spatial scale larger than 1000 km or so the magnitude of the negative marine geoid height trend south of Greenland is similar to the strength of the negative trend in geometric sea level from altimetry. This outcome affects past studies on changes in the strength of the Subpolar Gyre and the Atlantic meridional overturning circulation that neglect geoid height variations. We conclude that temporal geoid height trends should be considered in studies of (multi-)decadal trends in sea level and circulation on large spatial scales based on altimetry data.

1 Introduction

In oceanography the marine geoid represents an important reference surface relative to which the ocean dynamic topography (ODT) can be inferred from geometric sea surface height (SSH) measurements obtained by satellite altimetry. In the past temporal variations of the geoid were presumed to be small compared to other error sources and were neglected. Under this assumption, any difference from the underlying geoid model of the sea surface height field (SSH) as measured, e.g., by an altimeter, would only result from a moving ocean (neglecting any atmospheric contributions).

Over the last decades much effort was devoted by the geodetic and oceanographic community to obtain detailed knowledge about the geometric sea surface shape relative to the marine geoid down to 100 km spatial scale so as to allow a continuous monitoring of ocean geostrophic surface currents from space. After considerable advances in technologic and scientific devotion, the geometry of the sea surface is now measured routinely by satellite altimetry with a remarkable precision of the order of 1 cm over a horizontal scale of 100 km. Equally remarkable progress has also been achieved in improving geoid models, especially through modern precise space-borne gravity field observations obtained by the U.S./German satellite GRACE (Tapley et al., 2004) and the ESA GOCE (Rummel et al., 2002) satellite missions. Both developments together now allow to compute accurate absolute dynamic surface topography slopes and from this geostrophic surface current on horizontal scales down to about 100 km with an accuracy of 5 cm/s in midlatitudes (Bruinsma et al., 2014).

With its high precision, the GRACE observations also impressively documented broad-band (in time) geoid variations associated with mass movements within the Earth and climate system, involving all climate components. Causes for those movements are manifold, ranging from changes in the ocean and atmospheric circulations, changes in the terrestrial hydrology, to changes in the cryosphere, and, especially melting polar ice sheets and changes of mass distribution in the solid earth due to, e.g., tectonics, Post-Glacial Rebound (PGR) and earthquakes.

In the past, temporal geoid variations were usually neglected while deriving SLA and temporal Mean Dynamic Topography (MDT) from satellite data. However, already in the years leading to the GRACE mission the geoid effect of atmospheric masses and mass movements were discussed at length in the literature (Ecker & Mittermayer, 1969; Rummel & Rapp, 1976; Christodoulidis, 1979). In preparation of the GRACE mission several studies were therefore carried out to estimate the geoid effect of mass changes in the earth system (Dickey et al., 1997; Wahr et al., 1998; Chao, 2003). If neglected while deriving the ODT this can cause problems as it would lead to distorted dynamical sea surface slope estimates and thus distorted ocean currents. Since the sea surface shape adjusts quasi-instantaneously to the time-varying geoid, neglecting temporal geoid changes would project all geoid changes directly into the ocean ODT. However, details of the resulting uncertainty in space-based geostrophic velocity observations were never quantified.

With recent progress about geoid changes and their amplitude, it is now timely to re-think the concept of computing sea surface currents from the difference between time-varying altimetry and static geoid models. Most of past studies dealing with temporal gravity field changes were dedicated to the detection and analysis of the associated mass redistribution in the climate system. Only few publications address temporal geoid height variations over the ocean. As an example, studies in the pre-GRACE (Wahr et al., 1998; Nerem et al., 2000) and early GRACE period (Moore et al., 2005, 2006) infer mass redistributions in the Earth climate system from observed changes in gravity. Frederikse et al. (2017) consider geoid height changes in the Northwest Atlantic.

A number of recent publications deal with self-attraction and loading (SAL) effects caused by the coupled process of mass redistribution, crustal deformation and change in gravity (Tamisiea et al., 2010; Vinogradova et al., 2010, 2011). These publications investigate effects on relative sea level and ocean bottom pressure. Results from these studies support the analysis of observations from tide gauges and bottom pressure sensors, especially if comparing to ocean circulation models, since those usually don't include temporal changes neither in topography nor the geoid. Variations in the geoid changes, though implicit in these studies, were not considered explicitly. The Gauss–Listing geopotential value W_0 temporal variability is discussed in Dayoub et al. (2012), while we here concentrate on geoid gradients. We refer to Wahr et al. (1998) as a prominent example of the latter category.

In this paper, we revisit the question of what causes the geoid to vary as function of time scales to address three questions:

1. What kind of mass redistributions in the climate system are the main contributors to ocean geoid variations for time scales from weeks to decades;

2. How large are geoid height changes arising from mass variations associated with the ocean and atmospheric circulations as a function of geographic position, specifically: Do they matter (for specific time scales) compared to terrestrial signals; and
3. Do we need to account for temporal geoid changes in oceanographic applications?

The goal is to find the dominant contributions to ocean geoid height change for different time scales from weeks to decades, and to adress the geophysical processes involved. The focus is here on significance of geoid height changes for possible impact on ocean circulation studies. While the overall variations in geoid height are obtained from a state-of-the-art GRACE level-3 product, additionaly the contributing effects associated with ocean and atmospheric circulation are investigated separately. This is done to separate contributions from overlapping components of the earth system. More specifically, contributions from atmospheric circulation are separated either from those of the terrestrial system or the ocean circulation.

The spatial solution of geoid height changes available from GRACE products is considered sufficient for our study, though the application of dynamic modeling would allow higher resolution for changes associated with ocean and atmosphere circulation. However, since the geoid height change is proportional to the spatial scale of the associated mass change (see Eqs. 4–6 below), the sensitivity to small scale mass changes is low and the magnitude of geoid height variations diminishes with decreasing scale. In addition, on short spatial scales rather large variability in sea surface height is observed from altimetry caused by mesoscale eddy activity. Thus no significant impact of geoid height changes onto oceanographic applications is expected on these scales.

Our study considers geoid variations arising from (i) water mass cycling between the earth system components, including terrestrial water content (TWC), formation and melting of ice, global atmosphere and ocean mass, (ii) the atmospheric dynamics, (iii) the ocean dynamics, and (iv) PGR. Post-Little-Ice-Age adjustment processes, earth quakes and long term mantle-core processes also lead to changes in the marine geoid. However, their effect is neglected here. Respective unaccounted temporal geoid variations might cause errors on all time scales, specifically the annual cycle due to cycling of water mass through all components of the earth climate system. PGR and the mass redistribution associated with non-steric global sea level rise might cause considerable trends in the geoid.

The structure of the remaining paper is as follows: Section 2 introduces the data sets and the methodology used. For terrestrial water content and ice mass changes state-of-the-art GRACE solutions are applied, and no further validation is provided here. For geoid changes caused by atmospheric and ocean dynamics, however, we compare with external data in Section 3. Net geoid variations observed over the period 2003 through 2012 are described in Section 4. These variations are then split into fast (Section 5) and seasonal variations (Section 6), and linear trends (Section 7). In Section 8, the observed geoid trends are compared to trends in (sterically corrected) SLA from altimetry to discuss the potential impact of neglecting geoid height trends when studying long term changes of surface (deep ocean) currents. Concluding remarks are provided in Section 9.

2 Methodology and Approach

2.1 Components of temporal geoid variations

Temporal geoid variations are usually inferred from a series of states of various static geoid models typically inverted over 1-month periods. Changes between these states represent temporal geoid variations, $\Delta N(t)$, resulting from various mass changes in the earth system occurring on time scales longer than a month. Shorter-time scale variations need to be accounted for during the inversion process to not be aliased. See Dahle et al. (2013) for details on the generation of time-varying GRACE geoid models.

For our purposes we subdivide the change $\Delta N(t)$ between these geoid states into the following contributions:

$$\Delta N = \Delta N^{dyn}_{atmos} + \Delta N^{dyn}_{ocean} + \Delta N_{comp} + \Delta N_{PGR}. \quad (1)$$

Here, ΔN^{dyn}_{atmos} and ΔN^{dyn}_{ocean} are the contribution of air mass and ocean mass redistribution due to atmosphere and ocean dynamic, respectively while total - globally averaged - mass variations are not considered.

The remaining geoid changes

$$\Delta N_{comp} = \Delta N^{gl}_{atmos} + \Delta N^{gl}_{ocean} + \Delta N_{hydro} + \Delta N_{ice} \quad (2)$$

originate from mass fluxes between the different components of the Earth system by specifying the total mass change in the different terms, namely variations in Terrestrial Water Content (TWC, ΔN_{hydro}), land ice mass changes, including glaciers, ice caps and the Greenland and Antarctic ice sheets (ΔN_{ice}), and total changes in atmospheric (ΔN^{gl}_{atmos}) and ocean mass (ΔN^{gl}_{ocean}). ΔN_{PGR} represents the contribution of post-glacial rebound to the total geoid variations.

For each of the above contributions, crustal deformation due to loading and its effect on the geoid is also considered. However, the dynamic models used to compute ΔN^{dyn}_{ocean} and ΔN^{dyn}_{atmos} assume both the geoid and the topography not to vary with time; thus, water and air mass redistribution due to variations in the geoid or crustal deformation due to loading effects are not considered.

The models applied to identify the various contributions to geoid variations provide mass (re)distribution in form of bottom pressure BP or Equivalent Water Height (EWH), with $EWH = \Delta\sigma/\rho_w = BP/g/\rho_w$, where $= \Delta\sigma$ is horizontal mass density change, and gravity acceleration $g = 9.81ms^{-1}$ and sea water density ρ_w=1000 kg m^{-3}. Beside the changes in the geoid, we also show the corresponding variations in EWH where we found it supports the understanding of the physical processes behind the mass redistribution.

To obtain geoid height changes caused by mass redistribution near the Earth's surface we follow the methodology described in Wahr et al. (1998). Especially we apply the standard practice of a thin shell approximation assuming all mass changes at a fixed distance a from the center of earth. With this simplification the change in geoid height ΔN is described as a scale dependent weighting of the corresponding horizontal mass density change $\Delta\sigma$. To apply the weighting, first, the spatial mass change distribution is analyzed to obtain changes in the Spectral Harmonic (SH) coefficients Δc_{lm} and Δs_{lm} for degree l and order m as

$$\left\{ \begin{array}{c} \Delta c_{lm} \\ \Delta s_{lm} \end{array} \right\} = \frac{1}{4\pi a} \int_0^{2\pi} d\Phi \int_0^{\pi} \sin\Theta \; d\Theta \; \Delta\sigma(\Theta,\Phi) \; P^m_l(cos\Theta) \; \left\{ \begin{array}{c} \cos(m\Phi) \\ \sin(m\Phi) \end{array} \right\} \quad (3)$$

where P^m_l are the normalized associated Legendre functions (see e.g. Hofmann-Wellenhof and Moritz (2006)), and Θ and Φ are colatitude and eastern longitude, respectively.

Second, the SH coefficients are transformed back to physical space to obtain the change in geoid height as

$$\Delta N(\Theta,\Phi) = a \sum_{l=0}^{\infty} w_l \sum_{m=0}^{l} P^m_l(cos\Theta) \times (\Delta c_{lm}\cos(m\Phi) + \Delta s_{lm}\sin(m\Phi)) \quad (4)$$

applying the degree-dependent weighting

$$w_l = \frac{3\rho_w}{\rho_{ave}} \frac{1+k_l}{2l+1}, \quad (5)$$

with ρ_{ave} the average density of the Earth ($= 5517$ kg m^{-3}). The Love numbers k_l are taken from Tab. 1 in Wahr et al. (1998). For degrees l not listed there, linear interpolation between neighboring values is applied. The SH coefficients of horizontal mass density change (Δc_{lm}, Δs_{lm}) transform to the fully normalized SH coefficients of ΔN (ΔC_{lm}, ΔS_{lm}) as

$$\left\{ \begin{array}{c} \Delta C_{lm} \\ \Delta S_{lm} \end{array} \right\} = \frac{3}{2l+1}(1+k_l) \left\{ \begin{array}{c} \Delta c_{lm} \\ \Delta s_{lm} \end{array} \right\}. \tag{6}$$

The weighting w_l describes how, depending on spatial scale, a mass density change (and thus also EWH) transforms into geoid height change. Since the weighting is reciprocal to spherical harmonic degree, the geoid height is more sensitive to mass changes on large spatial scales. For hemispheric (degree one) mass redistribution the geoid height change will be roughly 20% of the corresponding change in EWH. For degree 100 (200 km spatial scale) the geoid height change is, however, just 2.7% of the mass change.

2.2 Estimating geoid variations by component

For temporal geoid variations on land, both from hydrological processes and the cryosphere, we apply one of the available level-3 gridded land field products based on space-borne observations of the Gravity Recovery and Climate Experiment (GRACE). Estimates of atmospheric mass changes and PGR are already subtracted in those data sets, so that they come as adequate and ready-to-use products for our purpose. Over the ocean and for atmospheric mass change over land we apply results from dynamical models. This allows for higher temporal resolution than available from GRACE, leakage of strong land signals into the ocean is avoided and a separation of ocean and atmospheric signals is provided.

In the following subsections we will explain the models applied and their application. Because we intend to analyze geoid variations ΔN over the period 2003–2012, all anomalies discussed below will therefore be referenced to this period.

2.2.1 ΔN_{PGR} *estimates*

To estimate PGR we apply Glacial Isostatic Adjustment (GIA) rates for a compressible Earth computed using the ICE5G ice history and the VM2 viscosity profile model (Geruo et al., 2012). It has to be stated that, though PGR trends among different GIA models provide robust pattern, the uncertainty of magnitudes in regional PGR trends from different GIA models is rather large (Caron et al., 2018). The total geoid height trend we present here is, however, independent from the PGR model, since for consistency we have added back the same GIA model that has been subtracted during GRACE processing. This model has been applied to remove GIA rates when producing the JPL-Mascon solution we use for geoid height changes from hydrological processes and the cryosphere. Adding this model back is thus indicated to obtain an unbiased estimate of total geoid height change.

2.2.2 $\Delta N_{ice} + \Delta N_{hydro}$ *estimates*

For mass changes due to changes in ice including ice caps, glaciers and the Greenland and Antarctica ice sheets, as well as hydrological changes, we apply the NASA Jet Propulsion Laboratory (JPL)-Mascon solution (Release 5, Version 2; Wiese et al. (2015); Watkins et al. (2015); Wiese et al. (2016)). From the provided 0.5° resolution the data is binned to a 1°× 1° grid, the resolution we use throughout our study.

2.2.3 ΔN^{dyn}_{ocean} estimates from GECCO reanalysis

Geoid variations caused by ocean dynamics are calculated from the GECCO2 reanalysis (Köhl, 2015). As with the atmospheric mass variations we separate ocean mass variations ρ_{ocean} with $\rho_{ocean} = \rho^{dyn}_{ocean} + \rho^{gl}_{ocean}$ into a dynamic and a global mean contribution. The dynamic surface mass density ρ^{dyn}_{ocean} is directly calculated from variations in hydrography and sea level of the dynamic model. Since GECCO2 utilizes the Boussinesq approximation it is however not capable for producing global mass variations. Instead, global ocean mass changes are calculated from variations in global TWC, the ice mass budget and global atmospheric mass by claiming the global water mass budget to be closed.

2.2.4 ΔN^{dyn}_{atmos} estimates from NCEP reanalysis

Geoid variations caused by atmospheric mass redistribution are calculated in this study applying daily mean surface pressure from the National Centers for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) Reanalysis 1 project (Kalnay et al., 1996). We utilize the hydrostatic approximation assuming changes in surface mass density $\Delta\rho_{atmos}$ are proportional to changes in surface pressure Δp_{atmos}, resulting in $\Delta\rho_{atmos} = 1/g * \Delta p_{atmos}$, with g=9.81 m s^{-2} the acceleration of gravity.

Over the ocean, $\Delta\rho_{atmos}$ is spatially averaged to take into account the Inverse Barometer (IB) effect. Due to this effect, regional variations in surface pressure are compensated by inverse variations in sea level so that only the spatial mean variations over the global ocean are detected in the Ocean Bottom Pressure (OBP) signal, whose variations measure the sum of variations of atmosphere and ocean mass above sea floor. This approach is justified by the long temporal scales considered. For our subsequent analyses the surface pressure is binned to weekly and monthly data sets. Finally, to separate the global mass variations from spatial redistribution, we define $\Delta\rho_{atmos} = \Delta\rho^{dyn}_{atmos} + \Delta\rho^{gl}_{atmos}$, with $\Delta\rho^{gl}_{atmos}$ the global mean of $\Delta\rho_{atmos}$. While $\Delta\rho^{dyn}_{atmos}$ is transformed to ΔN^{dyn}_{atmos} applying Eqs. 3–5, $\Delta\rho^{gl}_{atmos}$ is transformed to ΔN^{gl}_{atmos} used in Eq. 2 to compute ΔN_{comp}.

2.3 Splitting into disjunct frequency bands

To analyze signal strength and temporal evolution on different time scales we describe all mass density and geoid height time series as a sum of mutually uncorrelated terms on different frequency bands. Thus, the variances of the different terms add up to the total variance of the time series. For all but the atmospheric component the time series are separated into an annual and monthly mean term, where the annual mean is defined as anomaly over the reference period (2003–2011) and the monthly mean is defined as anomaly over the year it belongs to. For the atmospheric component in addition high frequency weekly variations are considered. These are based on daily fields with monthly means subtracted before computing weekly averages.

2.4 Meridional Overturning from altimetry and hydrography

In Section 8.3 the impact of the geoid height trend onto estimates of temporal changes in the Meridional Overturning Circulation (MOC) for a specific section over the North Atlantic is investigated. The magnitude of the upper branch of the MOC is estimated by combining the ODT with upper ocean hydrography information to map pressure on the section and derive geostrophic currents from horizontal pressure differences. Ageostrophic wind-driven transports are computed from wind stress data obtained from the NCEP reanalysis project (Kalnay et al., 1996).

To focus on interannual and longer time scales all applied input data is averaged to annual means. The distance between neighboring grid points on the section is set to 25 km while the depths are specified by the hydropgraphy data set selected. MDT, SLA, geoid height trend, wind stress and hydrography maps are interpolated to the grid points on the section, respectively. An additional depth $z = 0$ is defined by applying temperature and salinity of the uppermost level also for the surface. Surface pressure differences are determined from differences in ODT along the section, where the ODT is determined as the sum of MDT, SLA and, when indicated, the geoid height anomaly derived from the trend. With the density defined on the depth levels by the hydrography and expecting linearity of density in-between, the density as well as the pressure is known for arbitrary depths.

The transport for each neighboring grid point pair is then determined as integral from the surface to a selected potential density and accumulated over the entire section. We follow here the approach favoured e.g. by Mercier et al. (2015) to define density- rather than depth-dependend transports since northward warm waters and southward cold waters reside at overlapping depths and partially cancel each other out when defining depth-dependend transports. The magnitude of the MOC is then defined as the maximum of the (density dependent) transport. To allow for an integration down to a selected potential density and to determine the threshold density of maximum transport, potential density profiles are defined centered between the grid points of the setion. Potential density referenced to 1000m depth are determined as averages of the potential density profiles of the neighboring grid points. For each pair of neighboring grid points, depth levels for 0.02 kg m^{-3} potential density bins are determined and transports are calculated by integrating down to each of those density levels.

Ekman volume transports are calculated from the wind stress data for each grid point and are expected to evolve linearly between the grid points. Projection perpendicular to the section provides the required transport across the section.

To test the uncertainty of the input data and their impact on the MOC (variations) we apply two different MDT models,

1. The CNES-CLS18 MDT (Rio et al., 2014) and
2. a geodetic MDT based on the GECO (Gilardoni et al., 2016) geoid model and the DTU15 MSS (Andersen et al., 2016). Both, the geoid model and the MSS are developed until spherical harmonic degree and order 480. The MDT, derived as deviation of the MSS from the geoid, is spatially filtered applying a 0.3° truncated Gaussian filter. The full methodology is explained in Siegismund (2020).

and two different hydrography databases,

1. EN.4.2.1 (Good et al., 2013) and
2. ISAS-15 (Kolodziejczyk et al., n.d.; Gaillard et al., 2016)

3 Comparison with GRACE De-aliasing Products

To analyze geoid height changes based on hydrological and ice mass changes we use a state-of-the-art GRACE product and refer to available validation in the literature (Wiese et al., 2015; Watkins et al., 2015; Wiese et al., 2016). For the composite of atmospheric and ocean dynamical components (sum of the first two components in Eq. 1) we provide here a comparison with the GRACE Atmosphere-Ocean De-aliasing (AOD) product. The AOD product is based on results from dynamic atmosphere and ocean models and intended to serve as background model for the removal of high-frequency non-tidal variability in the production of GRACE level-2 data sets. We use here the version provided by Geoforschungszentrum Potsdam (GFZ; Flechtner et al. (2015); Dobslaw et al. (2013)).

We apply here the GAC product, which contains the sum of variations caused by ocean and atmosphere mass redistribution. This product is provided as spherical harmonic potential. The term c_{00}, which contains overall mass change, is not considered, since the atmosphere-ocean composite in Eq. 1 considers only mass redistribution with the total mass kept unchanged. After adding the loading effect the coefficients are transformed to 1°× 1° gridded geoid height anomalies applying the GOCE User Toolbox (GUT).

In Fig. 1a,b the Root Mean Square (RMS) values of both our ocean-atmosphere composite and the AOD product are shown, respectively. The global means are very close (GFZ AOD and composite: 1.4 mm), the same holds for the spatial patterns. For the region of high variability over Asia our composite shows slightly higher amplitudes than the AOD product. This might be caused by the effect of vertical atmospheric mass distribution on the gravitational potential which is taken into account in the AOD product, while in our composite the simple thin shell approximation (according to Wahr et al. (1998)) is applied. Fig. 1c shows the correlation of the two data sets considered here. While the global average is 0.78, two regions of low correlation are observed: One in the Atlantic and, to a lesser extent, another one in the western Pacific. However, variability in these regions is very low and possible mismatches in geoid height changes are negligible for our study.

4 Net Geoid Variations between 2003 and 2011

We will start our analysis by quantifying the net geoid variation as obtained from the sum of independent estimates of individual components using Eq. (1). The RMS of monthly mean values of this composite is shown in Fig. 2a, including all variations on time scales longer than 1 month and shorter than the 9 year long time series; also included is a trend over the 9-year period resulting from PGR. Enhanced variability can be found over the Amazon basin, and especially Greenland and West Antarctica. RMS values exceed 5 mm also over Siberia, South East Asia and Alaska. On the monthly to interannual time scales considered here, mass changes in these six regions are the major contributors to ocean geoid changes, while geoid variations over the ocean, away from the dominant sources over land, are relatively small.

Fig. 2b shows seasonal and sub-seasonal variations, while Fig. 2c displays interannual and longer time scale variations. A visual comparison of both panels with the top panel reveals that most variations in the original fields reside on the seasonal and subseasonal time scales. On these times scales the largest geoid variations are found over the Amazon basin; smaller amplitude changes are found in Siberia and South East Asia, Alaska and Northwest Canada along the Pacific Coast, over Southwest Greenland and over Africa. While for Siberia redistribution of atmospheric mass is responsible for the strong signal of approximately 5 mm, for all other regions mentioned we can expect changes in the presence of water mass (either in liquid or in frozen form) to be the primary cause for those changes. In contrast, geoid variations over the ocean are fairly modest, specifically near the equator.

On interannual and longer time scales, enhanced variability or changes can be found over Greenland and West Antarctica associated with the loss of ice masses there on longer time scales; this holds also over Alaska (probably because of glacial melting (Jin et al., 2017)). Less prominent signals are found over all continents, e.g. between the Black and the Caspian Sea (because of the decline of the sea level in the Caspian Sea (Chen et al., 2017)), over the Amazon Basin, southern Africa around 15°S and northeastern Australia.

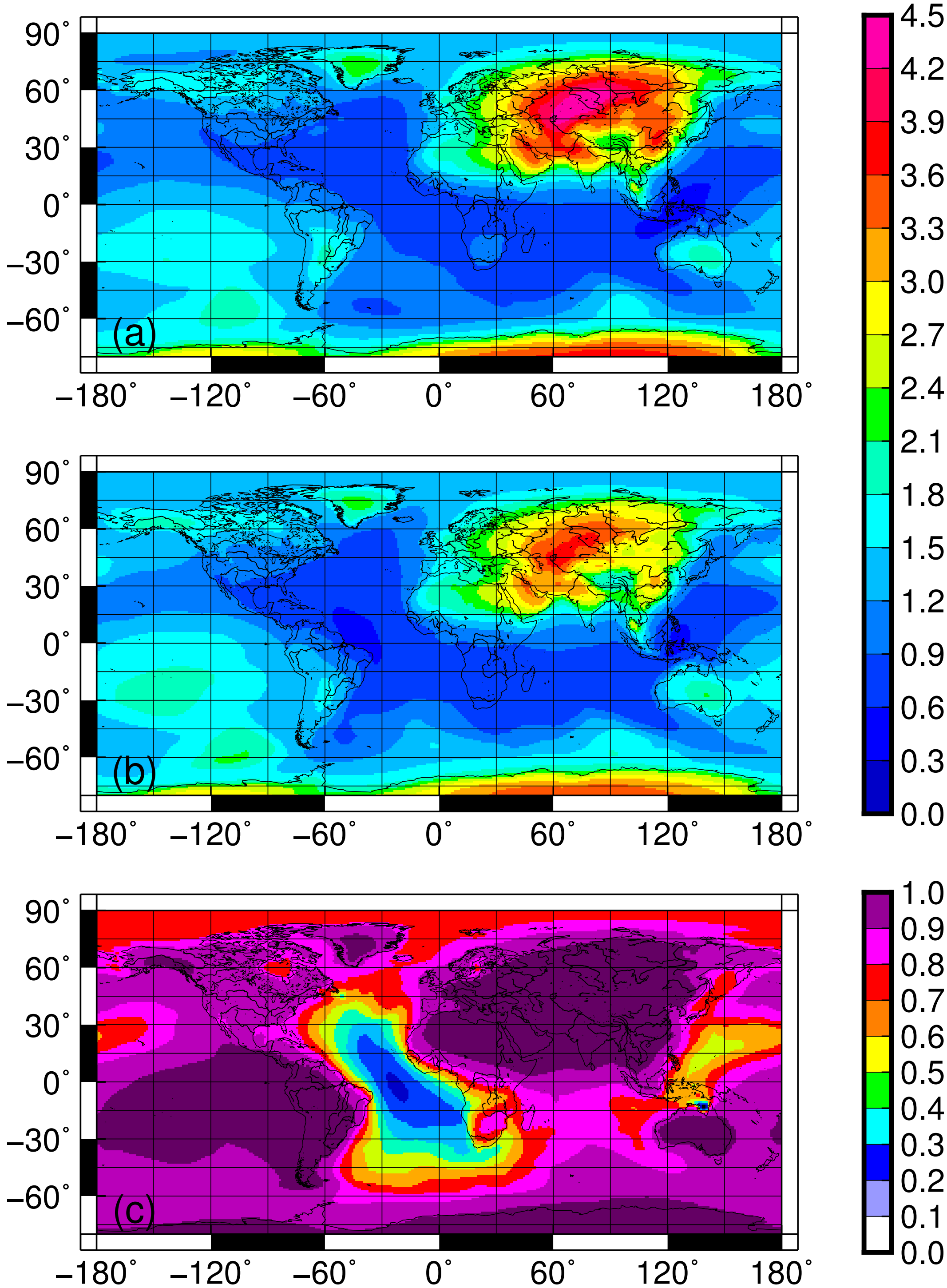

Figure 1. Comparison of geoid changes caused by atmospheric and ocean dynamics (ΔN^{dyn}_{atmos} + ΔN^{dyn}_{ocean}, see Eq. 1) with the GFZ GRACE Atmosphere-Ocean Dealiasing (AOD) product. Displayed are the RMS values [mm] of monthly mean anomalies for (a) ΔN^{dyn}_{atmos} + ΔN^{dyn}_{ocean} and (b) the GFZ AOD product, (c) shows the correlation of the two data sets.

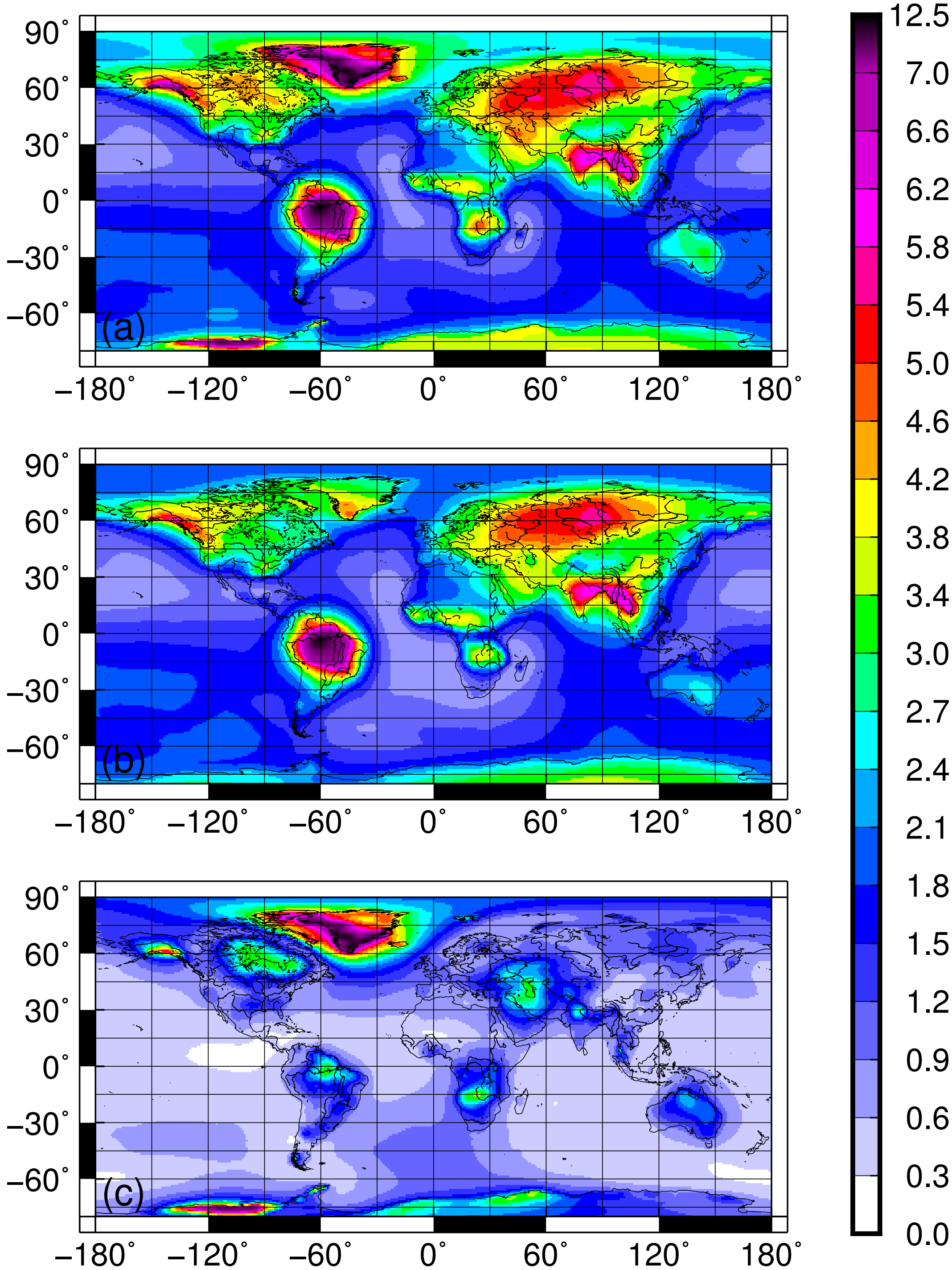

Figure 2. RMS values of geoid variations [mm] based on the composite of individual contributions provided in Eq. 1, from (a) unfiltered monthly mean data (2003 – 2012), (b) monthly mean data with the annual mean subtracted, (c) annual mean data.

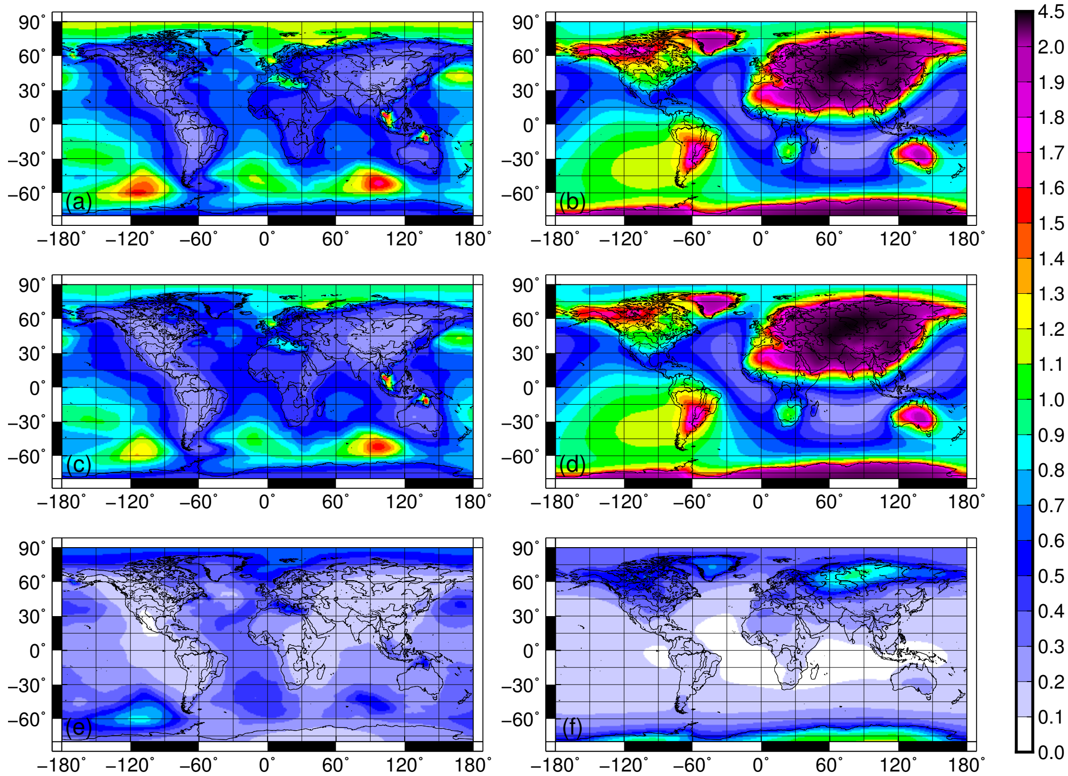

Figure 3. RMS values of geoid variations [mm] from (a,c,e) the ocean dynamic component ΔN^{dyn}_{ocean} and (b,d,f) the atmospheric dynamic component ΔN^{dyn}_{atmos} in Eq. 1, from (a,b) unfiltered monthly mean data (2003 – 2012), (c,d) monthly mean data with the annual mean subtracted, (e,f) annual mean data.

4.1 Contributions from atmospheric and ocean dynamics

The geoid height changes on monthly and longer time scales, as described by the four contributions of Eq. 1 are dominated by ΔN_{comp}, which consists of hydrological processes, mass changes in the cryosphere and mass fluxes between the different components of the Earth system. We do not present ΔN_{comp} here, which is rather similar to the variations seen in Fig. 2 for large part of the globe. Instead we focus on the remaining components of Eq. 1 and want to identify regions and time scales where these components considerably contribute to the net geoid height change presented in Fig. 2.

PGR, expressed as linear GIA trends, is presented in Section 7 were interannual changes are discussed. Here we focus on the remaining contributions ΔN^{dyn}_{atmos} and ΔN^{dyn}_{ocean} from atmospheric and ocean dynamics, respectively. The rows in Fig. 3 are organized the same way as in Fig. 2, with the left (right) column showing the contribution from ocean (atmospheric) dynamics.

(i) Ocean Dynamics: Fig. 3a shows the RMS of monthly geoid variations ΔN^{dyn}_{ocean} [mm] caused by ocean dynamics as they result from the GECCO2 ocean state estimate (see Section 2.2.2 for details). We recall that variations in global ocean mass are not considered in the figure. The figure reveals enhanced geoid height changes ($>$ 1mm) from ocean mass variations in essentially high-latitude oceans. Most prominent are changes in the Arctic revealing that the mass in this basin is changing substantially on the considered time scales. Enhanced variability can also be seen in the North Pacific and the Southern Ocean. Some of those locations are known for their enhanced barotropic variability. As an example, Stammer et al. (2000) describe the variability in the Southern Ocean and relate it to high barotropic variability there in the presence of closed f/H contours, with f the Coriolis parameter and H the ocean depth. Mass variations and their contribution to GRACE signal in the North Pacific have been discussed previously by Chambers and Willis (2008). We note here that since we show only variability on monthly and longer time scales substantial energy is already eliminated since most barotropic changes are on higher frequency.

Most of the RMS variations of monthly variations in ΔN^{dyn}_{ocean} originate from seasonal and sub-seasonal variations (Fig. 3c). Interannual variations (Fig. 3e) remain below 1 mm everywhere and are not further discussed here.

(ii) Atmosphere: Shown in Fig. 3b is the RMS of monthly geoid variations N^{dyn}_{atmos} caused by atmospheric dynamics related mass fluctuations. The fields were derived from NCEP/NCAR surface pressure fields as describe in Section 2.2.3. The surface pressure (not shown) reveals significant variations only over continents while the (spatially constant) variability over the ocean is with 3.9 mm EWH rather small. Strong variations are observed especially over Asia, but also over Greenland and Antarctica. Enhanced atmospheric mass variations but on lower scale can be seen over most of the remaining continents, except the tropical rain forest band.

After conversion to geoid height change (Fig. 3b), due to the scale-dependent weighting (see Eqs. 5,6), the continental signals spread over the ocean. Around the Asian continent RMS of monthly geoid height variations reach 2 mm, but also for Alaska, Greenland, Antarctica, Australia and part of South America near coastal RMS values above 1 mm are reached. Interestingly, the strong surface pressure variations over Asia produce a significant d/o 1 signal in the corresponding geoid height pattern with a second center of variability in the South Pacific west of Chile. Here RMS values around 1.1 mm are observed. As found in case of the ocean also geoid height changes associated with atmospheric dynamics reside almost entirely in seasonal and sub-seasonal time scales (Figs. 3d), while interannual variations (Fig. 3f) remain below 1 mm everywhere and are not further considered.

The analysis of geoid height changes so far should give an overview about temporal variations on intra- as well as interannual time scales broken down into the contributions of individual earth climate components and based on monthly mean data. In the following three sections we want to complete this analyse by (i) including fast changes on sub-monthly time scales and discuss, how well the already presented intra-and interannual variations can be described as a (ii) seasonal cycle and (iii) linear trend, respectively.

5 Geoid Variations on sub-seasonal time scales

Three individual contributions exist to geoid variations on the sub-monthly to intra-annual time scales, originating from (i) atmospheric mass variations, (ii) fast barotropic oceanic motions, and (iii) terrestial hydrological variations. We note that fast geoid motions on sub-monthly time scales are not resolved through GRACE monthly fields and thus are not included in Fig. 2. Significant sub-monthly geoid variations might result from atmospheric mass variations. Analysis of NCEP/NCAR weekly surface pressure variations shows that approximately 40% of the variance is made up by sub-monthly variations. Therefore, we estimated geoid variations caused by atmospheric mass redistribution down to a weekly time scale from external - non-GRACE sources as explained in Section 2.

This time we show the variations of both, mass (Fig. 4, left) and associated geoid height (Fig. 4, right) to visualize the scale-dependent weighting involved in the transformation process (Eqs. 5,6). The left panel of Fig. 4 displays the RMS of surface pres-

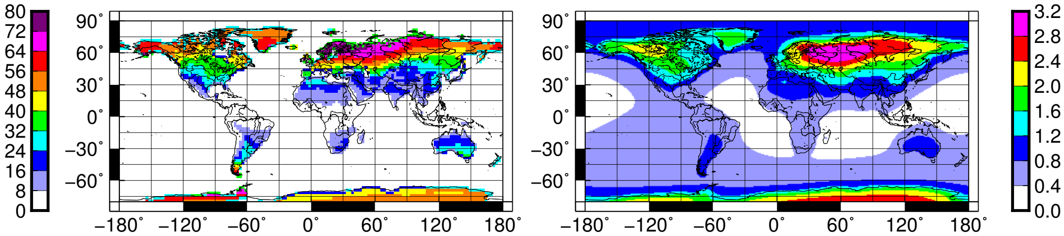

Figure 4. RMS of weekly mean air mass redistribution caused by atmospheric dynamics, after subtraction of the monthly mean. Displayed is the RMS of (left) the surface pressure in terms of Equivalent Water Height [mm], (right) the geoid height [mm], respectively.

sure anomalies associated with air mass redistribution caused by atmospheric dynamics on time scales longer than weekly, after subtraction of the monthly mean. The panel highlights the large pressure fluctuations associated with high-latitude low pressure/storm systems. In contrast, tropical regions are much more "quiet". This holds also in the variations of associated geoid changes shown in the right panel of Fig. 4. For regional patterns of strong air mass variability the transformation to geoid height variations both, flattens and spreads the signal including larger areas, depending on the spatial scale of the pattern. Especially for Siberia and Antarctica the large scale structure of the surface pressure patterns allows the signal to keep substantial magnitude after transformation to geoid height with an RMS of up to 3.2 mm, while the smaller scale pattern with similar amplitude over Alaska refers to a much weaker geoid height signal. Over the ocean the RMS of geoid height variations exceeds 1 mm only close to Antarctica and in the Arctic Mediterranean.

6 Seasonal Geoid Variations

Since the Earth system shows enhanced variability on the seasonal cycle in many of its components, it can be expected that pronounced geoid variations resulting from associated mass shifts in the system exist on the seasonal cycle.

To isolate seasonal geoid variations from what was shown above in the Figs. 2b and 3c,d we estimated seasonal changes in the geoid by fitting an annual harmonic to our monthly composite of contributions from land ice and terrestrial hydrology, the ocean and the atmosphere.

Shown in Figs. 5b,c is the respective amplitude and phase of the total seasonal geoid variations [mm], respectively. The pattern of amplitudes resembles that of total intra-annual variability which we shown again in Fig. 5a, but now with values only over the ocean and the same (but differently scaled) color bar as for Fig. 5b for better comparison. The variations over the Amazon, North and Southeast Asia, Alaska and Southern Greenland are reproduced . Fig. 5d displays the percentage of the variance explained by the seasonal cycle relative to the total intra-annual variability. Regions with strong variability (Fig. 5a) are also those where the variability is explained best so that the bulk of total variability is explained by the seasonal cycle.

Considering ocean dynamics (see Fig. 6a,c,e,g), the seasonal cycle is not as suitable to explain the intra-annual variations in mass distribution in contrast to the total or the other components in our composite. Still, the predominant part of the strong variability around Indonesia and east of the Kerguelen Plateau in the Southern Ocean can be attributed to seasonal variations.

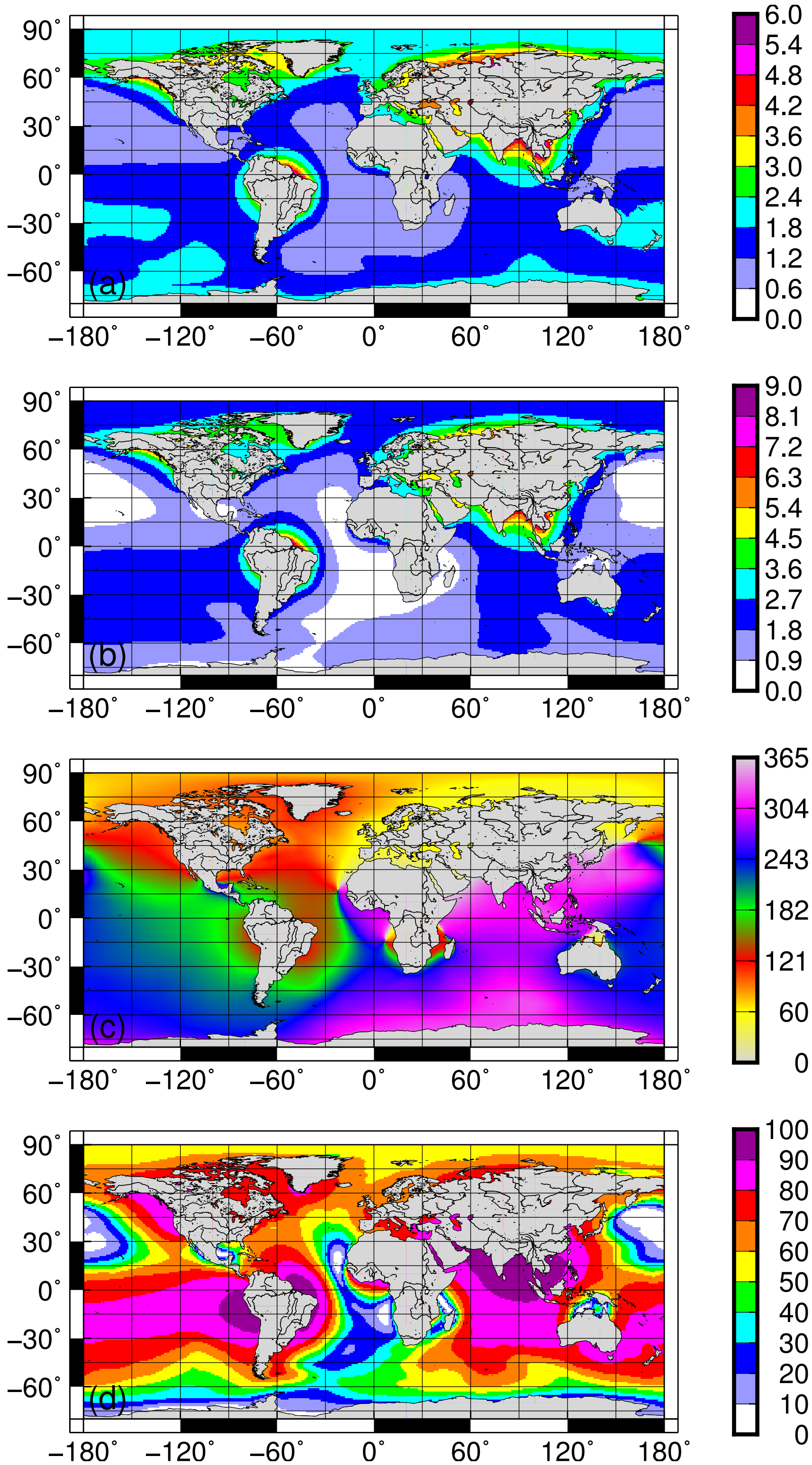

Figure 5. The intra-annual variations in monthly mean geoid height based on the composite of individual contributions provided in Eq. 2. Results of a Least Squares fit to a seasonal cycle $A * cos(\omega t - \lambda)$ are presented. Displayed are (a) RMS of intra-annual variations [mm] (copy of Fig. 2b, but now with land masked out and a different color scale) (b) the amplitude A [mm], (c) the day of maximum λ, (d) the explained variance in %.

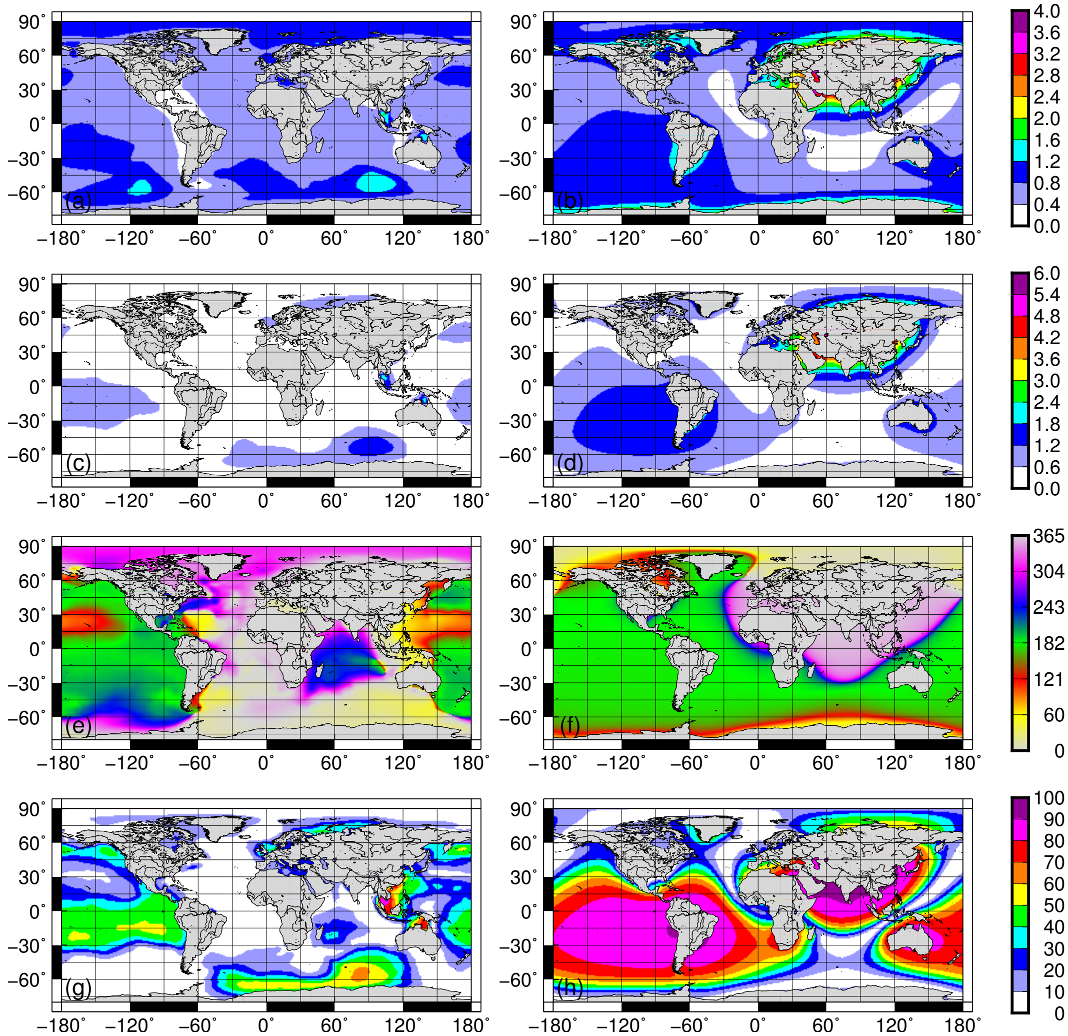

Figure 6. Same as Fig. 5, but for water mass redistribution caused by (a,c,e,g) ocean and (b,d,f,h) atmospheric dynamics.

For inter-annual mass redistribution due to atmospheric dynamics (see Fig. 6b,d,f,h), especially those over Asia, the seasonal cycle is a good approximation that explains most of the variability in all regions with strong variability. Over the ocean, the remaining inter-annual variations in geoid height that cannot be explained by the seasonal cycle are around or below 1 mm.

7 Linear Geoid Trends 2003 - 2012

As above for the intra-annual variations, we checked also if the annual mean variations can be described approximately by a linear trend. Fig. 7 displays the results of this test for geoid variations based on the sum of our composite (including PGR). We stated already above that atmospheric and ocean dynamics only play a minor role for interannual geoid height variations. We just note, that geoid height trends from ocean (atmospheric) dynamics nowhere exceed 0.3 (0.1) mm a^{-1} (not shown) and that annual mean ocean mass increases around 1.4 mm a^{-1} in terms of EWH which is close to what is published recently (e.g. Slangen et al. (2017) and references therein.) Significant geoid height trends are restricted to the cryosphere and terrestrial hydrology, and PGR. We show the total trend in Fig. 7, since this is the important parameter for long-term ocean studies, and PGR in Fig. 8, since this a significant contribution to the trend.

The regionally strong interannual variations around Greenland and parts of Antarctica (see Fig. 7a) are predominantly explained by a linear trend (Fig. 7b,c). The negative trend south and southwest of Greenland up to more than 7 mm a^{-1} near the coast is by far the largest and thus most important signal for long-term ocean studies.

A very prominent contribution to geoid changes on long time scales is known to originate from PGR. To bring the respective signal into context of the observed linear trend we show in Fig. 8 the geoid height trend [mm a^{-1}] as it is caused by PGR; i.e. by viscous mass adjustments in the lithosphere. Positive (uplift) signal is centered around the locations of the Laurentide and Fennoscandian ice sheets. Similar signals can be identified over western Antarctica. Centers of respective rebound signals are located in the western subtropical North Atlantic and the southern Indian Ocean. A comparison of Figs. 7 and 8 suggests that over the continental North America and northern Europe the PGR signal is counter balancing the trend caused specifically by changes in the terrestrial hydrological cycle in these regions.

The PGR trends among different GIA models provide robust pattern, but the uncertainty of magnitudes in regional PGR trends from different GIA models is rather large (Caron et al., 2018). Following the Supporting Information, Fig. 4 of Caron et al. (2018) this might be especially true for the negative trend centered in the tropical North Atlantic with uncertainty in the order of 0.1 mm yr^{-1}.

8 Potential Impact on Dynamic Topography and Transport Estimates

Changes in geoid height are generally ignored when investigating temporal variations in sea level from altimetry data. From the analysis presesented above this approach seems justified for short spatial and temporal scales because of considerably larger variability in sea level anomalies from altimetry than of the geoid. However, when discussing inter-annual and longer term variations and trends the observed sea level amplitudes are usually much smaller. When ignoring geoid height variations, the potential bias in sea level variation studies relative to the investigated signal grows with the time scale considered.

In the following subsections we provide three examples for sea level and circulation studies with potentially large impact from geoid height changes on their findings. Due to the massive mass loss of the Greenland ice sheet and in addition a negative PGR

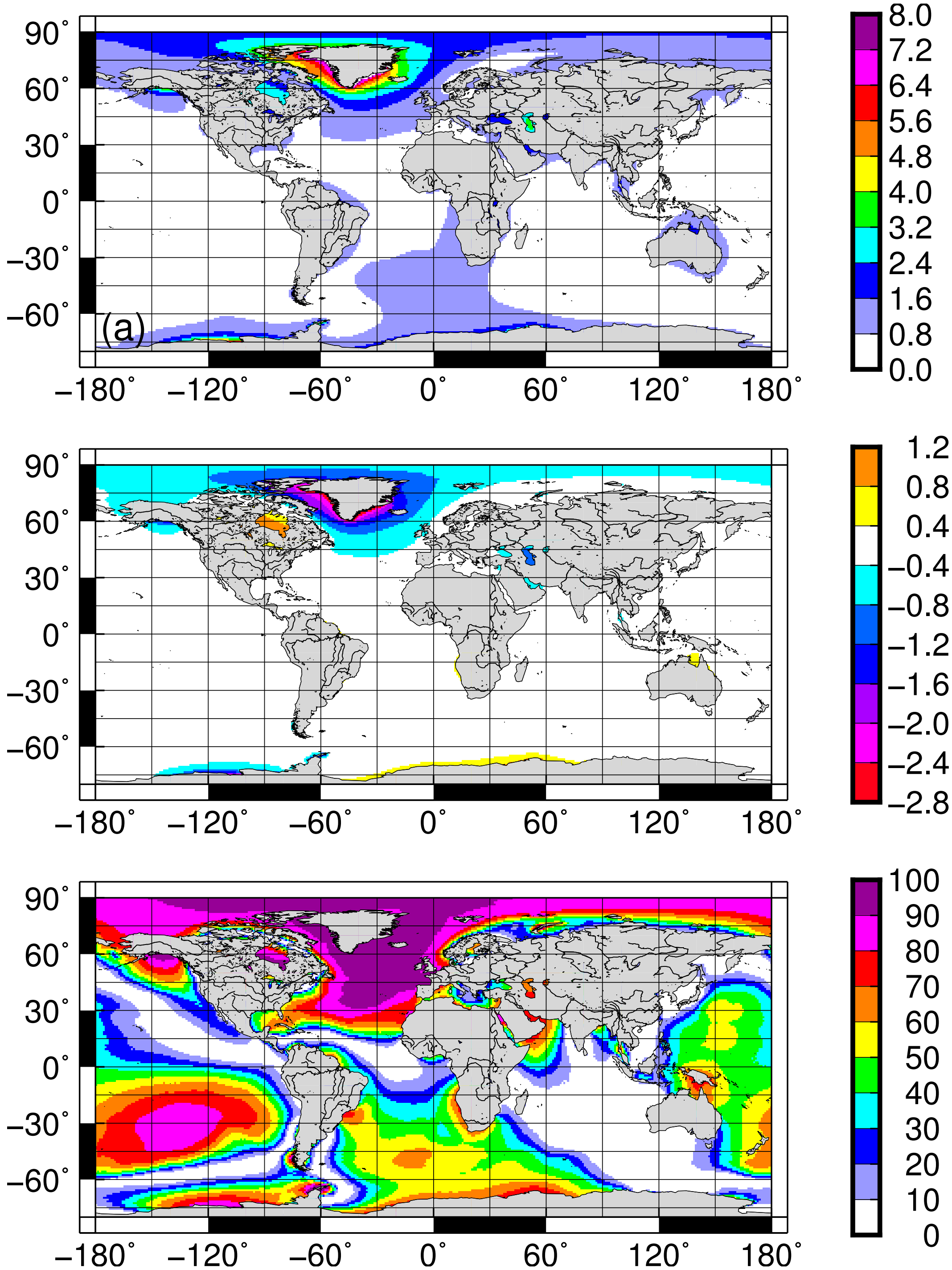

Figure 7. Total geoid height trend (2003–2011) based on the composite of individual contributions provided in Eq. 1. Displayed are (a) RMS of inter-annual variations [mm] (copy of Fig. 2c, but now with land masked out and a different color scale), (b) the trend [mm a^{-1}], (c) the explained variance in %.

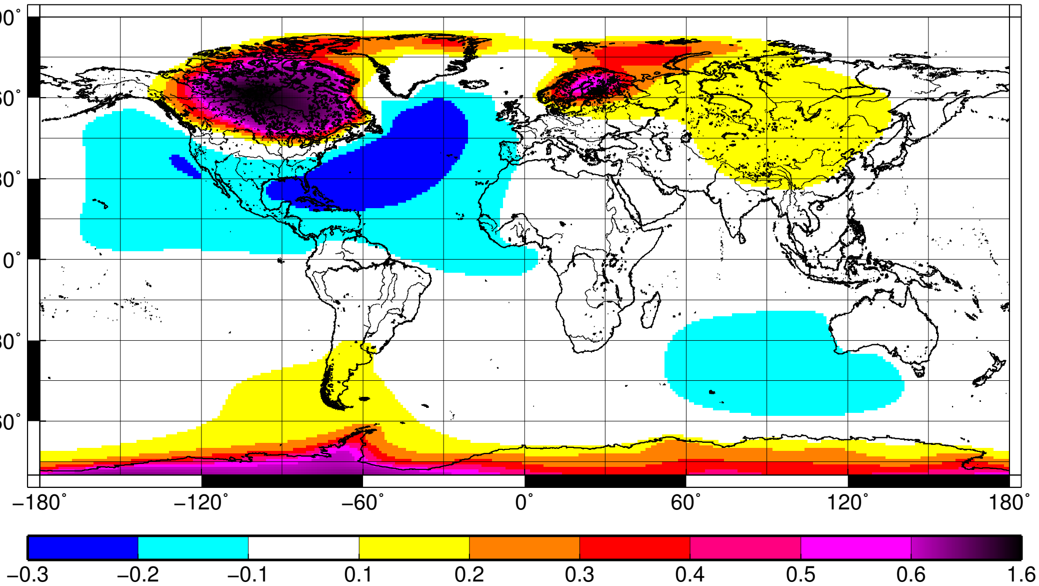

Figure 8. Geoid height trend [mm a^{-1}] caused by post-glacial rebound.

trend south of Greenland, the highest long term geoid changes are found in the North Atlantic. We will therefore focus on this region. In all three cases we investigate the impact of geoid height change by performing two test cases. In the first case geoid height is supposed invariant while in the second case the geoid height change is estimated as the (location-dependend) trend we presented in Section 7 (Fig. 7). The differences of the outcomes are discussed.

8.1 North Atlantic Subpolar Gyre Strength

A prominent example of analysing large-scale ocean circulation changes based on altimetry is the strength estimation of the North Atlantic Subpolar Gyre. For this analysis we apply the delayed-time all-satellite merged altimetry data provided by the Copernicus Marine and Environment Monitoring Service (CMEMS) for the period 1993 to 2016. The data is binned to annual means on a $1° \times 1°$ grid. The standard methods to filter out small scale signals like spatial filtering or spectral methods based on Fourier Transforms or spherical harmonics have the disadvantage to smooth also the patterns of long-term variations we are interested in. Therefore we use instead an Empirical Orthogonal Function (EOF) analysis here which automatically provides modes with long scales of coherency in the leading EOFs while local small scale variations contained in subsequent modes are cut off automatically when using only the leading modes.

The region analyzed (30°N–65°N, 80°W–0°E) is identical to the region often used to investigate variations in the strength of the Subpolar Gyre (SPG, see *Hatun and Chafik* (2018) and references therein). The global mean SLA is subtracted to focus on variations in the dynamics which are related to local sea level changes relative to the global mean rather than the global mean itself. We consider two cases: One applies SLA as described with temporal geoid variations completely included, while in the second case SLA is corrected for the estimated geoid trend from our composite shown in Fig. 7a. Though this trend might overestimate the geoid change in the years before GRACE (1993–2002) when Greenland ice sheet decrease was smaller than in recent years, the rough approach applied here should be sufficient for our impact study, especially in view of future regional

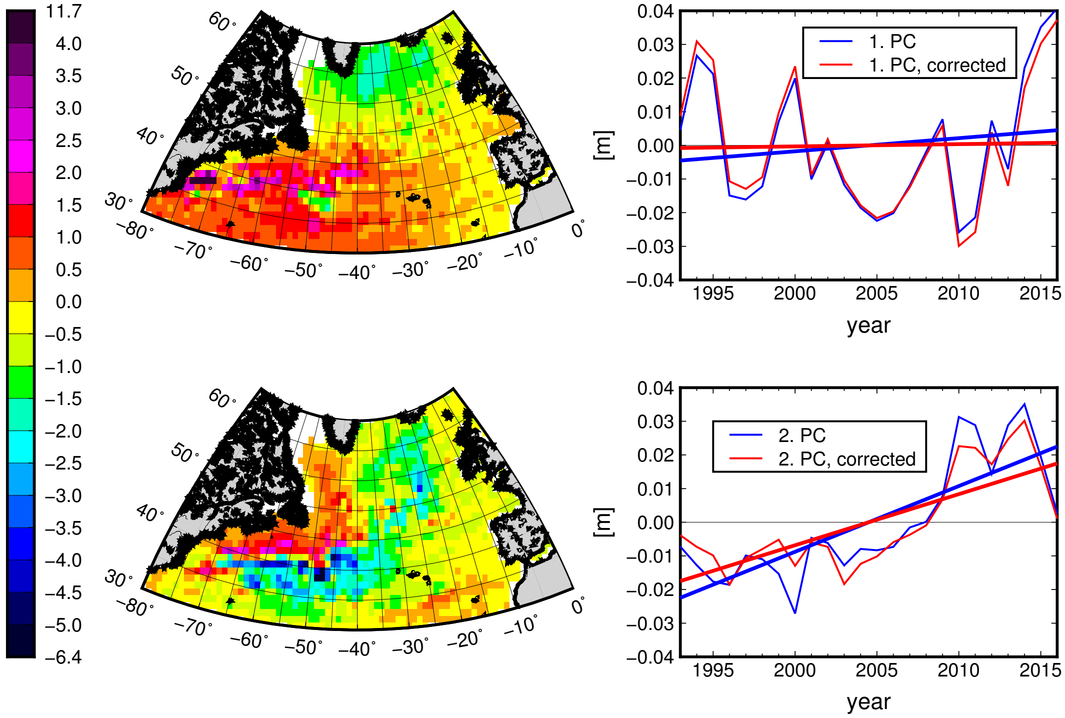

Figure 9. EOF analysis of annual mean SLA from altimetry. Displayed are (left) the EOFs and (right) the PCs for (top) the leading and (bottom) the second mode. For the PCs two time series are provided: (blue) the PC corresponding to the EOF displayed on the left and (red) the PC obtained from a second EOF after subtraction of the geoid trend. For changes in the EOF see text.

sea level studies with larger geoid height trends expected from increased Greenland ice sheet melt.

We concentrate here on the leading two modes of variability from the EOF analysis. These explain 34% (33%) of the variance in annual mean un-corrected (corrected) SLA, respectively. Hatun and Chafik (2018) argue, that these two modes combined are necessary to deduce the strength of the SPG from SLA in recent years, in contrast to the usual description by the (normalized) first PC only, termed as SPG index. In Fig. 9 the results of the EOF analysis are displayed. The PCs carry the units, while the EOFs are normalized. The EOFs reveal robust patterns hardly dependent on the correction with correlations of 0.99 and 0.96 for the first and second mode, respectively. As a simple linear trend is applied as correction, as expected, also the PCs mainly differ with respect to the trends. These trends change significantly for both PCs when correction with respect to the geoid height trend is applied. Though this correction hardly influences interannual variability in SLA from altimetry without correction, trends get biased if only hydrodynamic processes are investigated and not the geometric change in sea level is of interest. This bias will grow with acceleration of melting of the Greenland ice sheet in the future.

8.2 Sterically-Corrected Sea Sevel

Temporal changes in sea level observed in altimetry data is an often used indicator for changes in the upper ocean circulation and hydrography. Theoretically, the combination of SLA with density profiles from in-situ observations allows a determination of temporal variations in volume, heat and salt transports, and deep ocean circulation. Practically, the spatial and temporal density of temperature and salinity observations has to be sufficient as well as the accuracy of both altimetry and hydrography. The improving knowledge of the regionally dependent upper ocean density profile from ARGO floats in the upper ocean starts to offer an alternative to the standard level of no (or known) motion approach. To determine changes in the weak deep ocean currents this does, however, also increase the requirements on the accuracy of sea level gradients. We want thus in a second example analyze the potential impact of the usually neglected geoid height trend on estimates of changes in the deep ocean circulation based on sterically corrected SLA.

The top panel of Fig. 10 shows the linear trend in sea level for the period 1993–2016 after subtraction of the global mean and smoothing the SLA with a 10° spatial Gaussian filter truncated after two standard deviations. A number of trend patterns of both increasing and decreasing sea level emerge, with amplitudes up to approximately 2 mm a^{-1}. This is close to amplitudes we found for the geoid height shown in Fig. 10 (top).

In the middle panel of Fig. 10 steric height from surface to 1000m depth is subtracted from SLA to correct the total dynamic sea level change from the bulk of the steric effects and provide an estimate of the trend in deep circulation as dynamic height at 1000 m. The trends are now generally smaller near the equator and again in the order of the geoid height trends. This points to a significant influence of geoid trends when altimetry is combined with in-situ hydrography to estimate long term changes in deep ocean circulation.

For a closer view in the bottom panel of Fig. 10 we subtracted the geoid height trend as seen from our composite (top panel in Fig. 7) from the sterically corrected SLA. Outside of the North Atlantic the spatial structure of positive and negative trend patches is hardly changed, though significant changes in amplitudes are found. Due to the strong negative geoid height trend south of Greenland the low negative trend in sterically corrected SLA changes to an increase of approximately 1 mm a^{-1}. As above for the surface circulation in the subpolar North Atlantic also investigations of the deep circulation and associated heat and salt transports based on altimetry data are affected by geoid height trends.

8.3 Overturning from Combining Altimetry with In-Situ Hydrography

In Section 8.1 we have already discussed the impact on estimating changes in the surface circulation of the Subpolar North Atlantic from altimetry when disregarding geoid height trends. Performing the investigation on a coast-to-coast section and adding density information from hydrography allows to extend this analysis to temporal variability of the Meridional Overturning Circulation (MOC) of the Atlantic, which is a crucial element in climate research. Specifically with the increased availability of insitu hydrography from ARGO floats the combination of insitu hydrography, altimetry and (optionally) wind speed data/models becomes an alternative or complement to the elaborate and expensive section-based measurements from ship cruises or moored instruments. Specifically long-term trends are an important issue and this is where geoid height trends might essentially bias the results.

The section where MOC variability is investigated is defined as two connected geodetic lines, starting from Cape Farewell and ending at the coast of Portugal (see Fig. 10, top), and is close to the OVIDE and FOUREX sections (see e.g. Mercier et al. (2015)). The methodology described in Section 2.4 follows largely the methodology described by Mercier et al. (2015). Since we are interested in interannual to longer-term changes all computations here work with annual mean data. To investigate the impact of the geoid height in comparison to uncertainties in other datasets, we compute an ensemble of MOC estimates by combining two MDT models (CNES-CLS18 and a geodetic MDT) with two

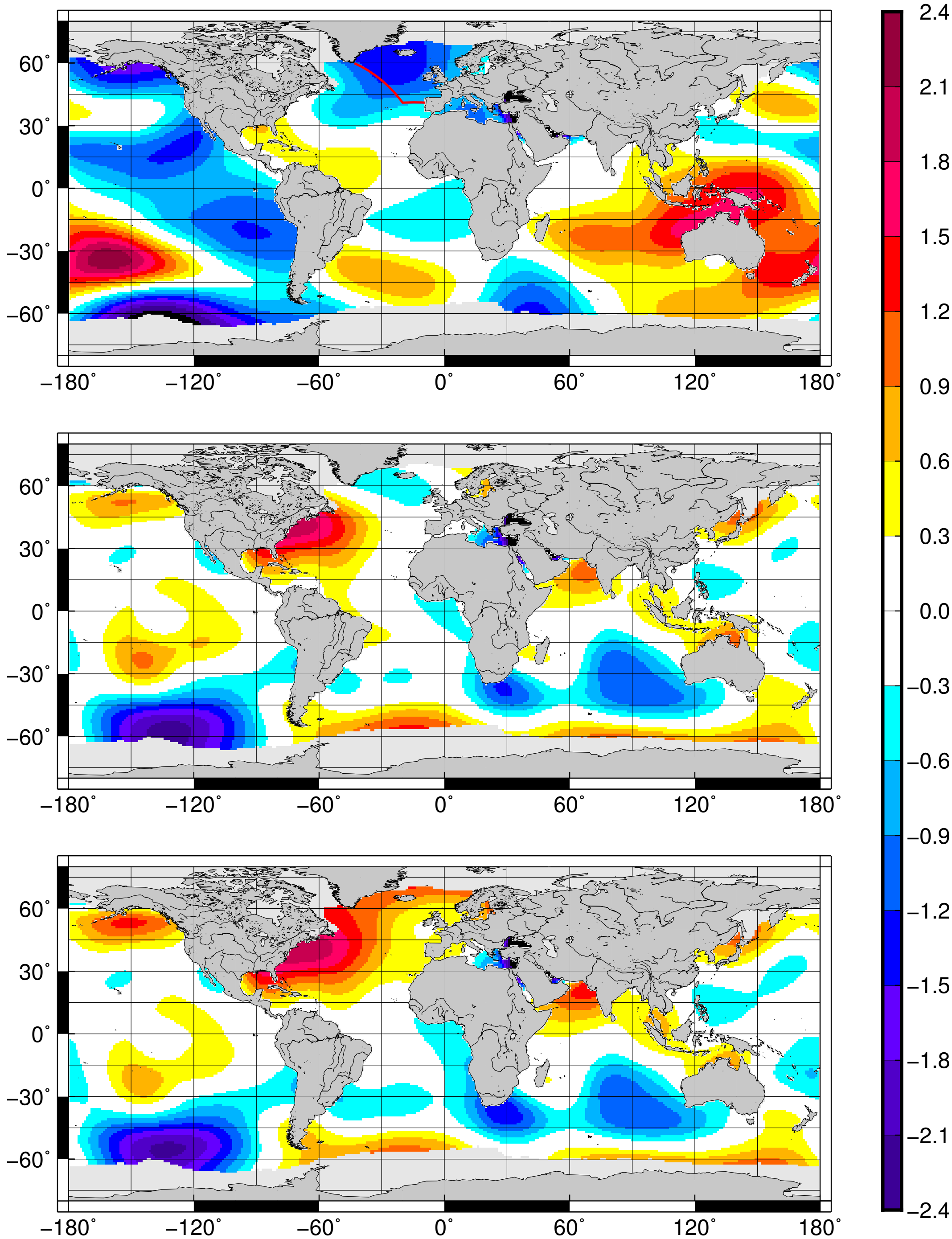

Figure 10. Trend in sea level anomalies from altimetry (1993–2016, mm a^{-1}) after subtraction of (top) the global mean, (middle) in addition the steric height for the upper 1000 m, and (bottom) in addition the linear trend in geoid height (see Fig. 10). In addition, in the top panel the section used for MOC strength estimates in Section 8.3 is displayed.

hydrography databases (ISAS and EN4). Utilizing all possible combinations results in an ensemble of four members. Since ISAS is only available for 14 years from 2002–2015, which is too short for our purpose, the years 1993–2001 and 2016–2018 are filled with data from EN4. We found, however, a large variance for the threshhold potential density that marks the lower bound of the upper branch of the MOC (see Fig. 11). Those unrealistic variations are probably caused by mesoscale variability in SLA, where the corresponding steric effects cannot be resolved by the hydrography data base and will thus cause unrealistic geostrophic currents. We have thus, in a second ensemble, fixed the threshold potential density to 32.16 kgm^{-3}, the value found by Mercier et al. (2015) for 1997–2010 based on measurements on the OVIDE and FOUREX sections.

The results are displayed in Fig. 12 for three-years running means. For both ensembles and all members a decreasing MOC during the 1990s, a maximum around 2003 and a weak MOC between 2006–11 is observed. This is generally close to the results of Mercier et al. (2015), though they found a strong MOC around 2010. Fixing the threshold potential density to 32.16 kgm^{-3} lowers the MOC on average by approximately 1 Sv($=10^6 m^3 s^{-1}$), has some impact on the magnitude of extremes and increases the linear trend for all ensemble members. Interestingly, apart from the case when both fixing the threshold density and applying EN4 hydrography data, trends are negative when ignoring geoid height variations, but zero or positive when the geoid height trend is considered. The mean MOC difference for 1994–2017 following the trend is increased by 1.2 Sv if the geoid height trend is included in the investigation. The standard deviation of MOC trends for the ensemble ignoring geoid height changes is only 0.7 Sv. Though given the small number of members (and the partly dependence of the data) a robust estimation of uncertainty of the MOC estimates can hardly be provided, it can be stated that the geoid height trend at least has a significant impact on the MOC trend that will increase with extended periods considered and an accelerated geoid height trend for Greenland in the future.

9 Conclusions

This paper discusses the weekly to interannual variations in geoid height as a whole and subdivided into the contributions from water mass cycling between the earth system components (terrestrial water content and land ice, global atmosphere and ocean mass), the atmospheric and ocean dynamics, and PGR. The analysis is performend on different time-scales and the regional patterns should allow investigators working with SLA or ODT to decide whether temporal variations in the geoid have to be considered significant in their study or are negligible.

Our main conclusions are:

1. Submonthly geoid height variability over the ocean due to redistribution of atmospheric mass is everywhere small because of the IB effect (RMS below 2 mm) , specifically away from the coast (below 1 mm).
2. Monthly geoid height variations are between 0.5 mm and 5 mm over the oceans. Larger RMS values are found only along the Greenland coast.
3. After subtraction of the annual mean the monthly variations are predominantly caused by changes in terrestrial water content and atmospheric mass redistribution with only minor contributions from ocean dynamics (below 2 mm). These intra-annual variations are generally well represented by a seasonal cycle defined as a trigonometric function.
4. For the geoid height trend over the period considered (2003–2011), only GIA and terrestrial water content changes play a role with the largest, negative, signals around Greenland and in the Pacific section of the Southern Ocean due to decreasing ice sheets and a positive signal south of the African continent.

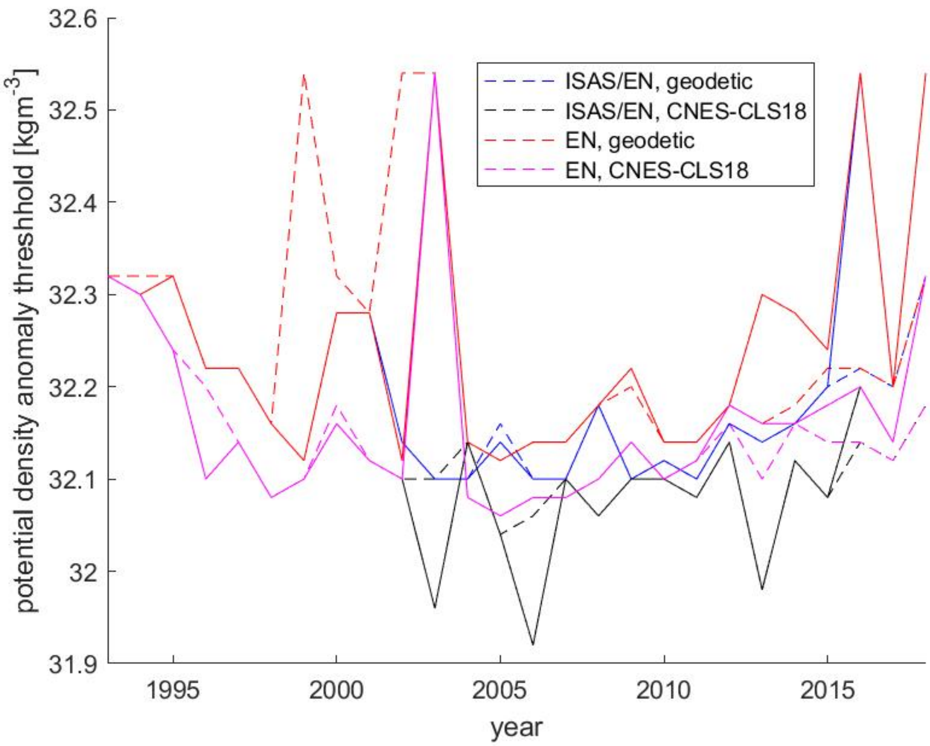

Figure 11. For the computation of the upper-limb transport MOC over the Greenland-Portugal section, the threshold potential density anomaly σ_{th} is displayed. σ_{th} defines the density level where integration of section-accumulated transports, which is started from the surface, maximizes. The applied hydrography (EN or combined ISAS/EN) and MDT model (CNES-CLS18 or geodetic) is indicated in the inset. For solid (dashed) lines geoid height trend is included (excluded) in Dynamic Topography computation.

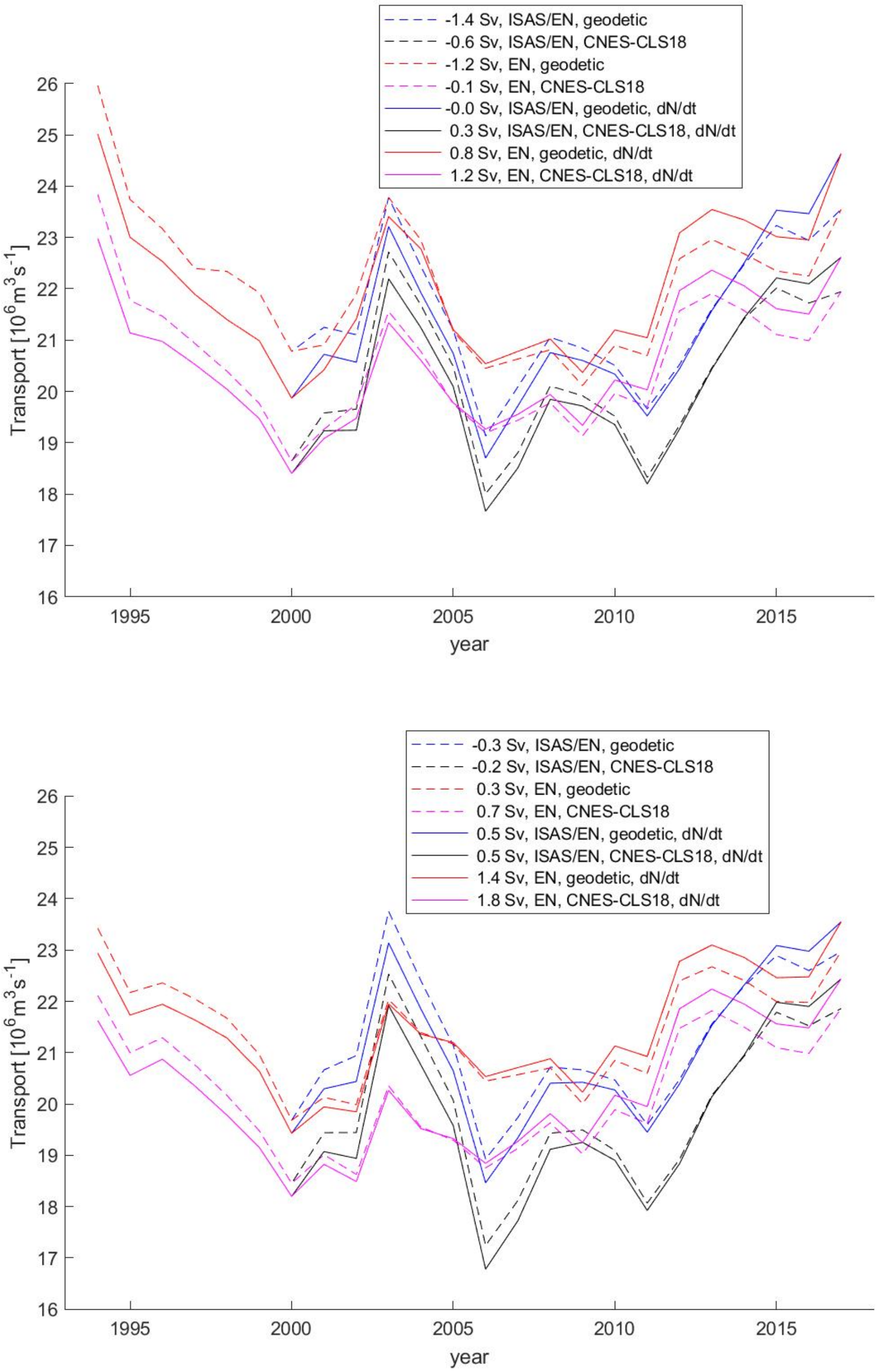

Figure 12. Running three-years-mean MOC for years 1994–2017. The change in MOC within the period, according to the linear trend, is provided in the inset, together with the applied hydrography (EN or combined ISAS/EN) and MDT model (CNES-CLS18 or geodetic). (Top) Integration is performed down to the threshhold potential density anomaly as shown in Fig. 15, (bottom) integration is always performed down to potential density anomaly $\sigma = 32.16 kgm^{-3}$.'

5. Geoid height variations are usually not considered when altimetry data is applied to investigate changes in ocean dynamics. From our study this seems justified if short temporal (up to inter-annual) or spatial scales (up to 1000 km or so) are considered. However, for the subpolar North Atlantic, due to the melt of the Greenland ice sheet, the associated strong geoid height trend is biasing long term changes in surface and deep ocean currents based on (sterically corrected) altimetry data. A correction for the geoid height change is necessary in that region when investigating long term changes in sea level based on altimetry data.

Acknowledgments

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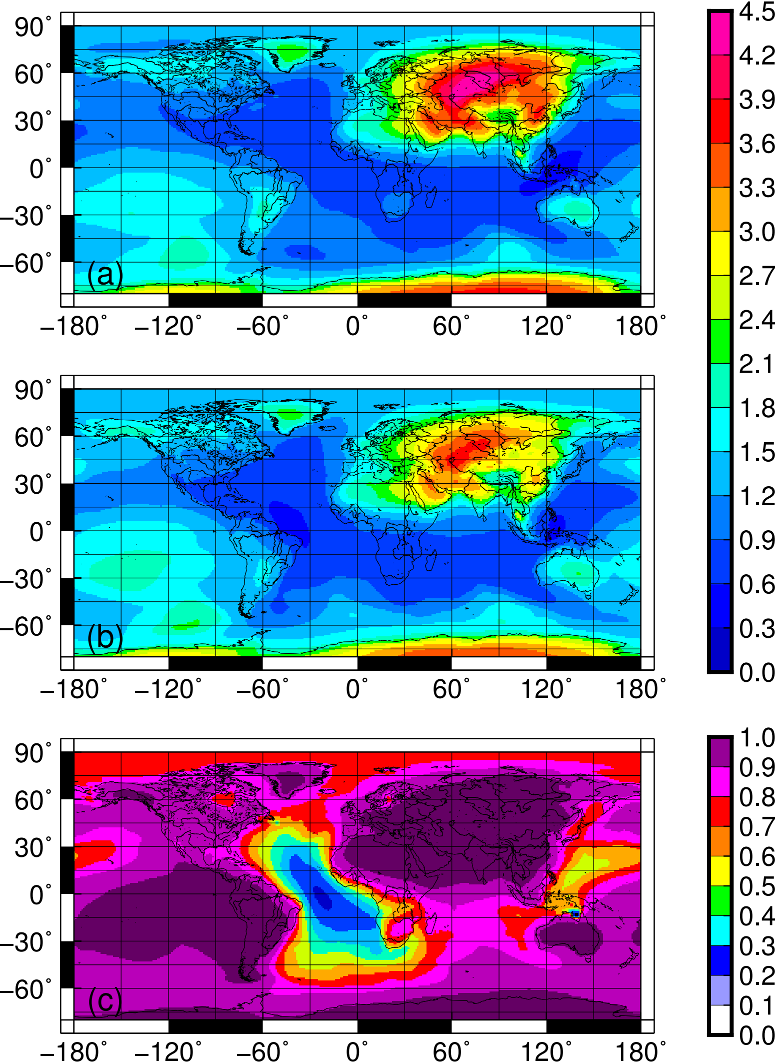
Figure 1.



90˚
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–60˚
(a)
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0˚
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4.5
4.2
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3.6
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3.0
2.7
2.4
2.1
1.8
1.5
1.2
0.9
0.6
0.3
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1.0
0.9
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0.5
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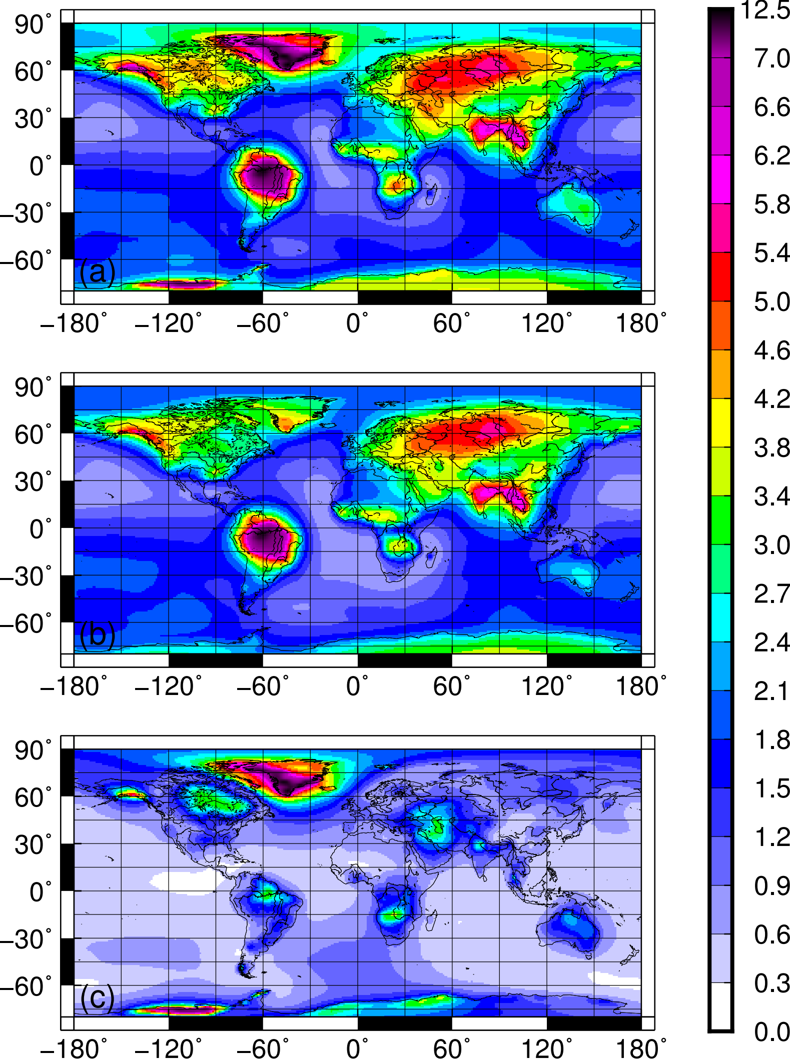
Figure 2.



(a)
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60°
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180°
12.5
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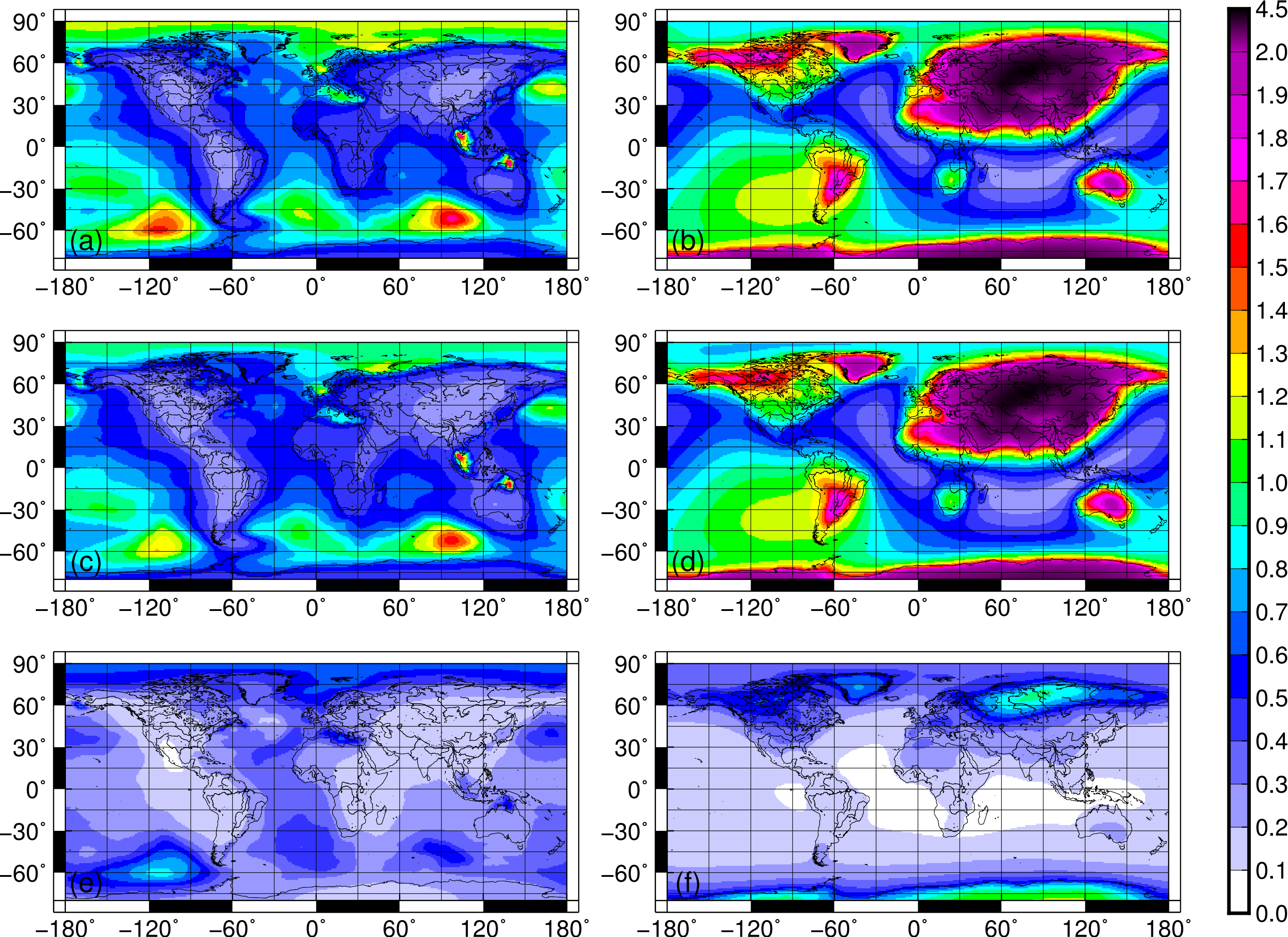
6.2
5.8
5.4
5.0
4.6
4.2
3.8
3.4
3.0
2.7
2.4
2.1
1.8
1.5
1.2
0.9
0.6
0.3
0.0

Figure 3.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1.3
1.2
1.1
1.0
0.9
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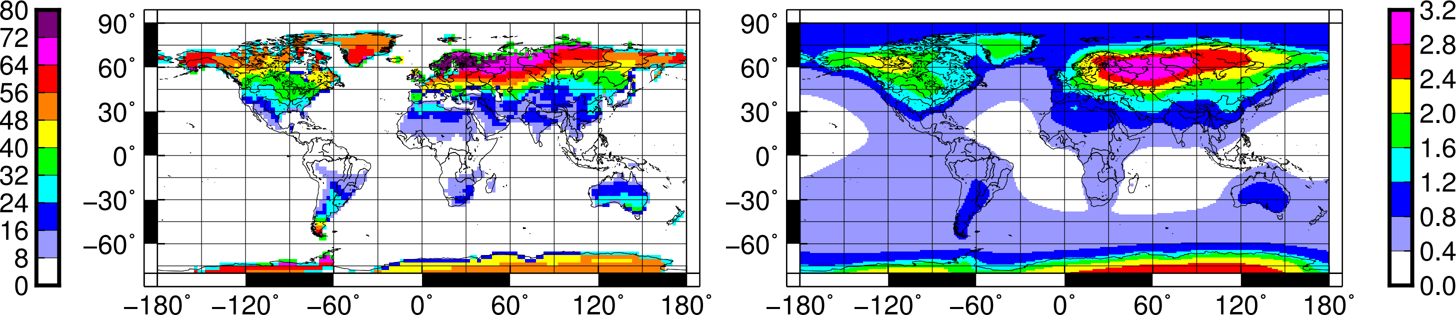
Figure 4.



80
72
64
56
48
40
32
24
16
8
0
90°
60°
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0°
−30°
−60°
−180°
−120°
−60°
0°
60°
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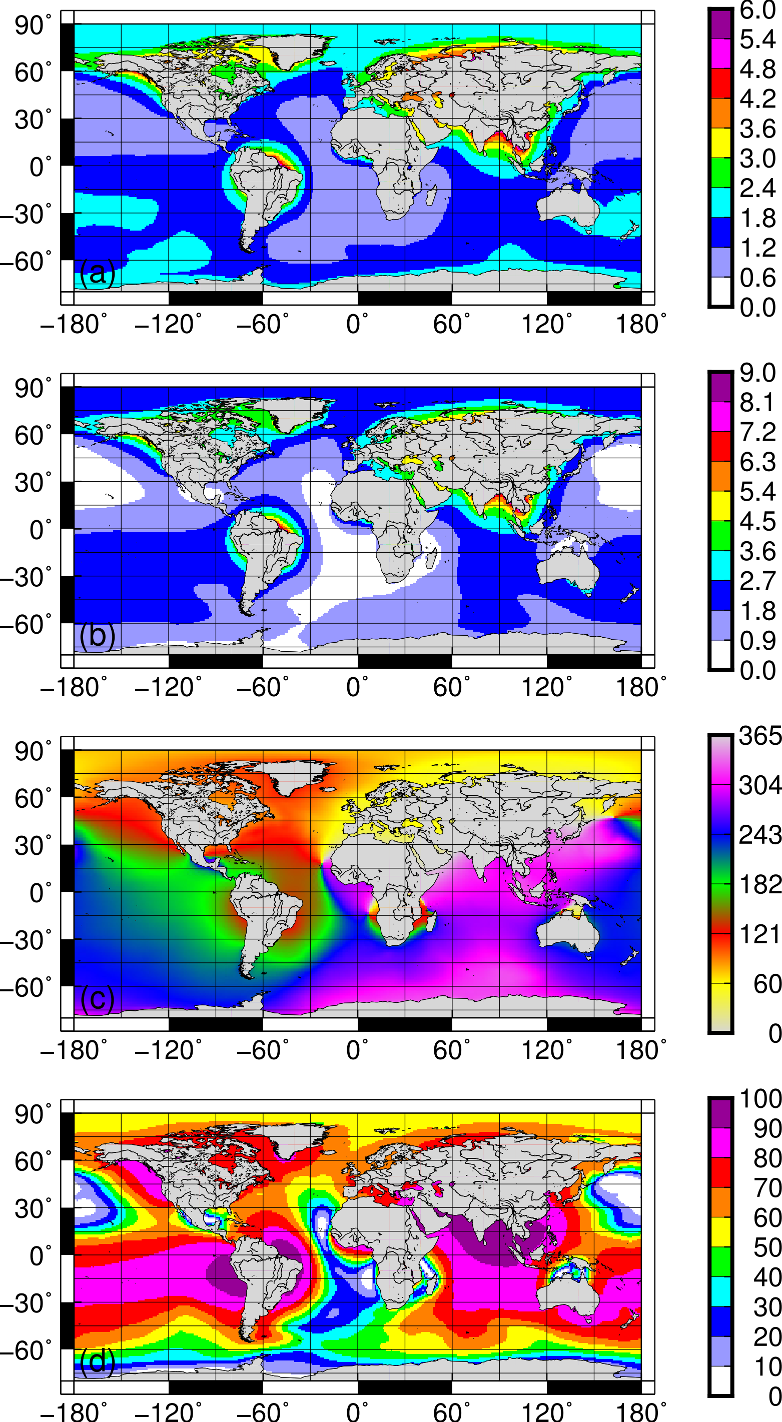
3.2
2.8
2.4
2.0
1.6
1.2
0.8
0.4
0.0

Figure 5.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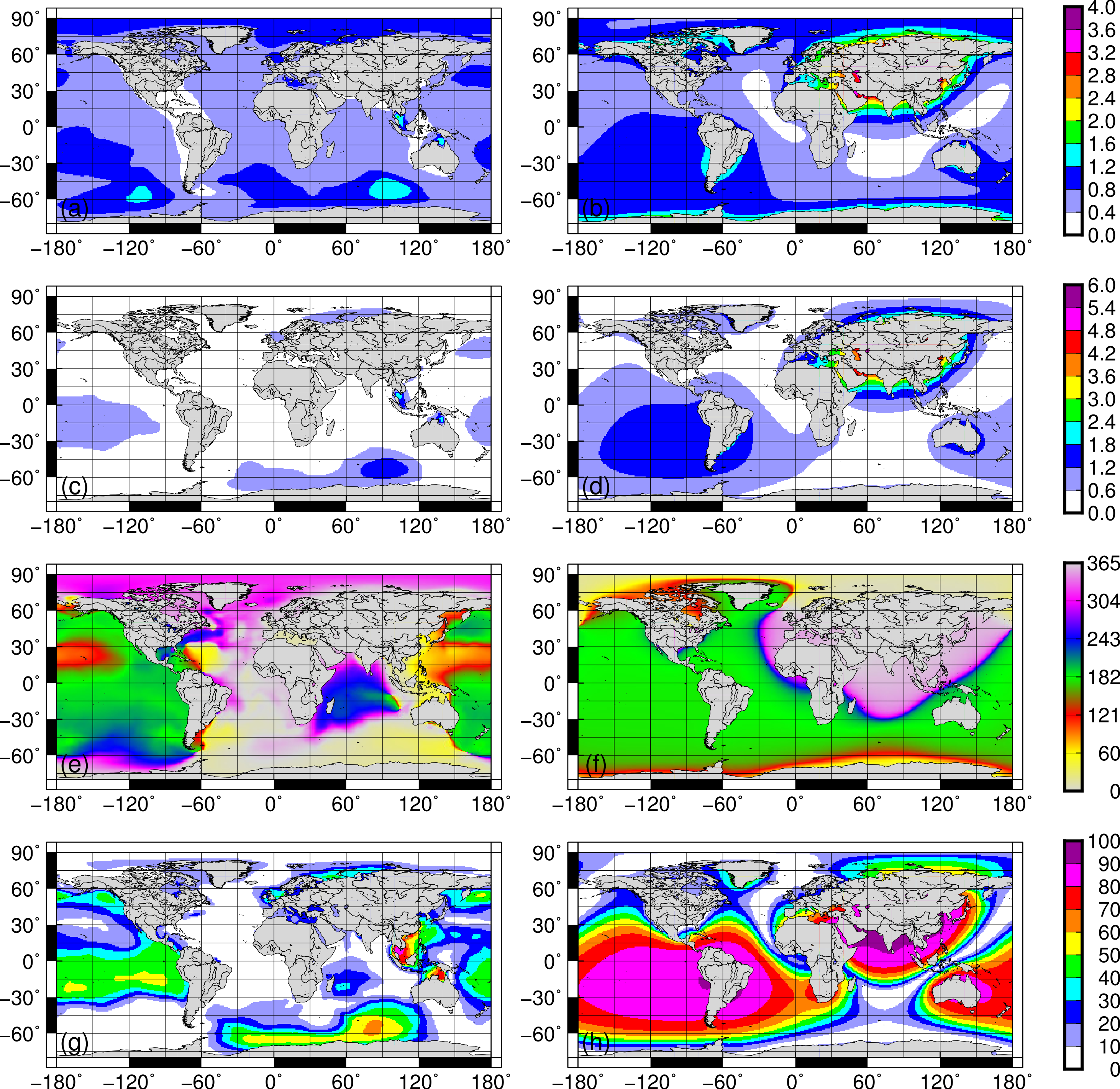
6.0
5.4
4.8
4.2
3.6
3.0
2.4
1.8
1.2
0.6
0.0
(b)
9.0
8.1
7.2
6.3
5.4
4.5
3.6
2.7
1.8
0.9
0.0
(c)
365
304
243
182
121
60
0
(d)
100
90
80
70
60
50
40
30
20
10
0

Figure 6.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0.4
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1.2
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2.0
2.4
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0.0
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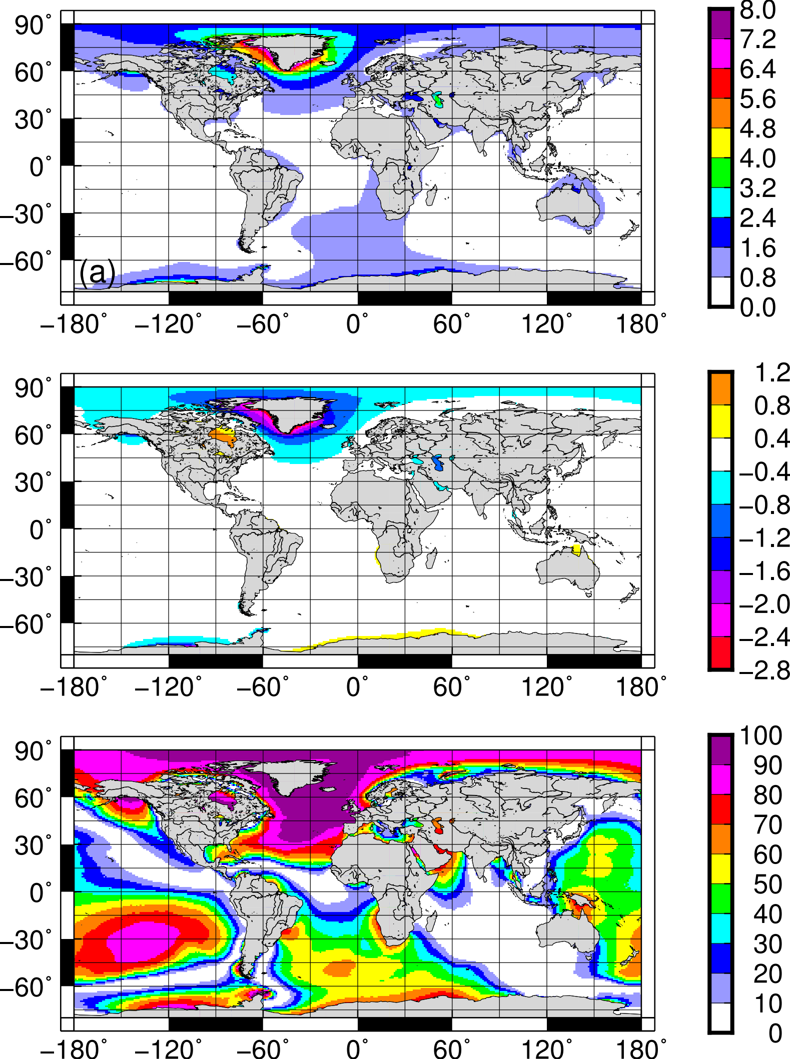
100

Figure 7.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4.0
3.2
2.4
1.6
0.8
0.0
1.2
0.8
0.4
−0.4
−0.8
−1.2
−1.6
−2.0
−2.4
−2.8
100
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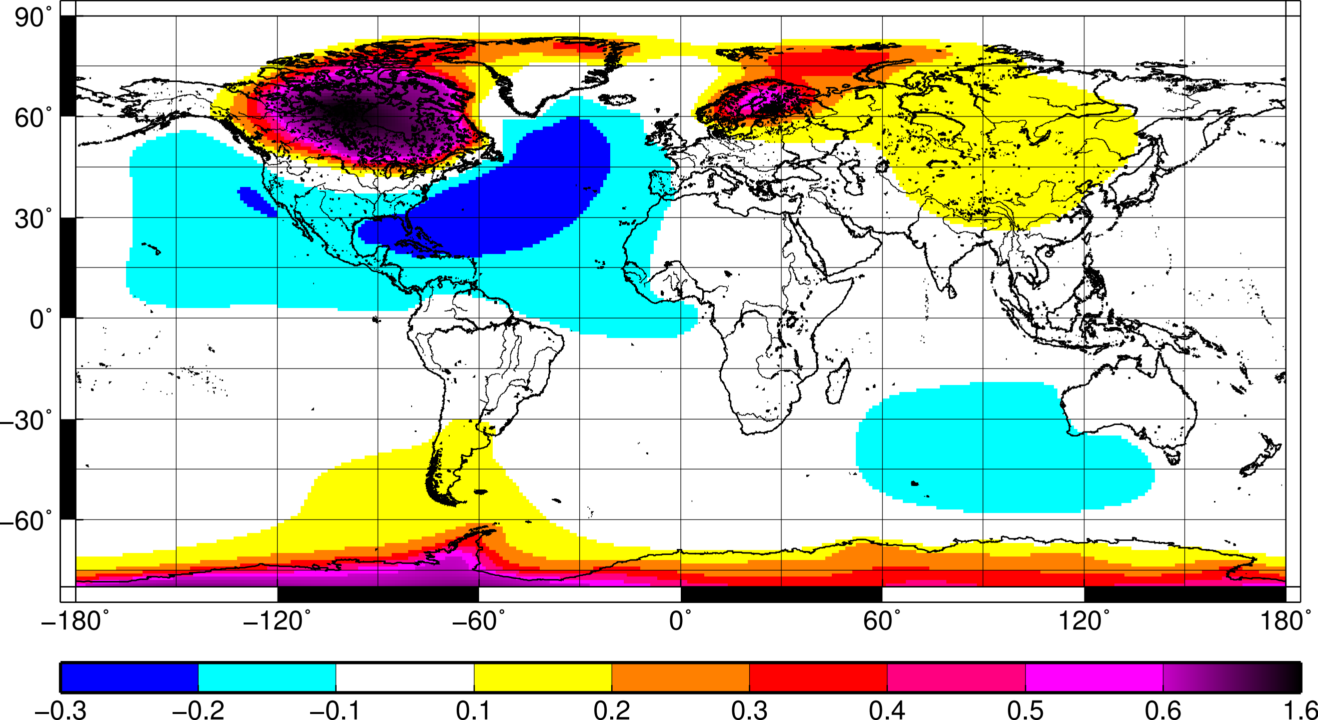
Figure 8.



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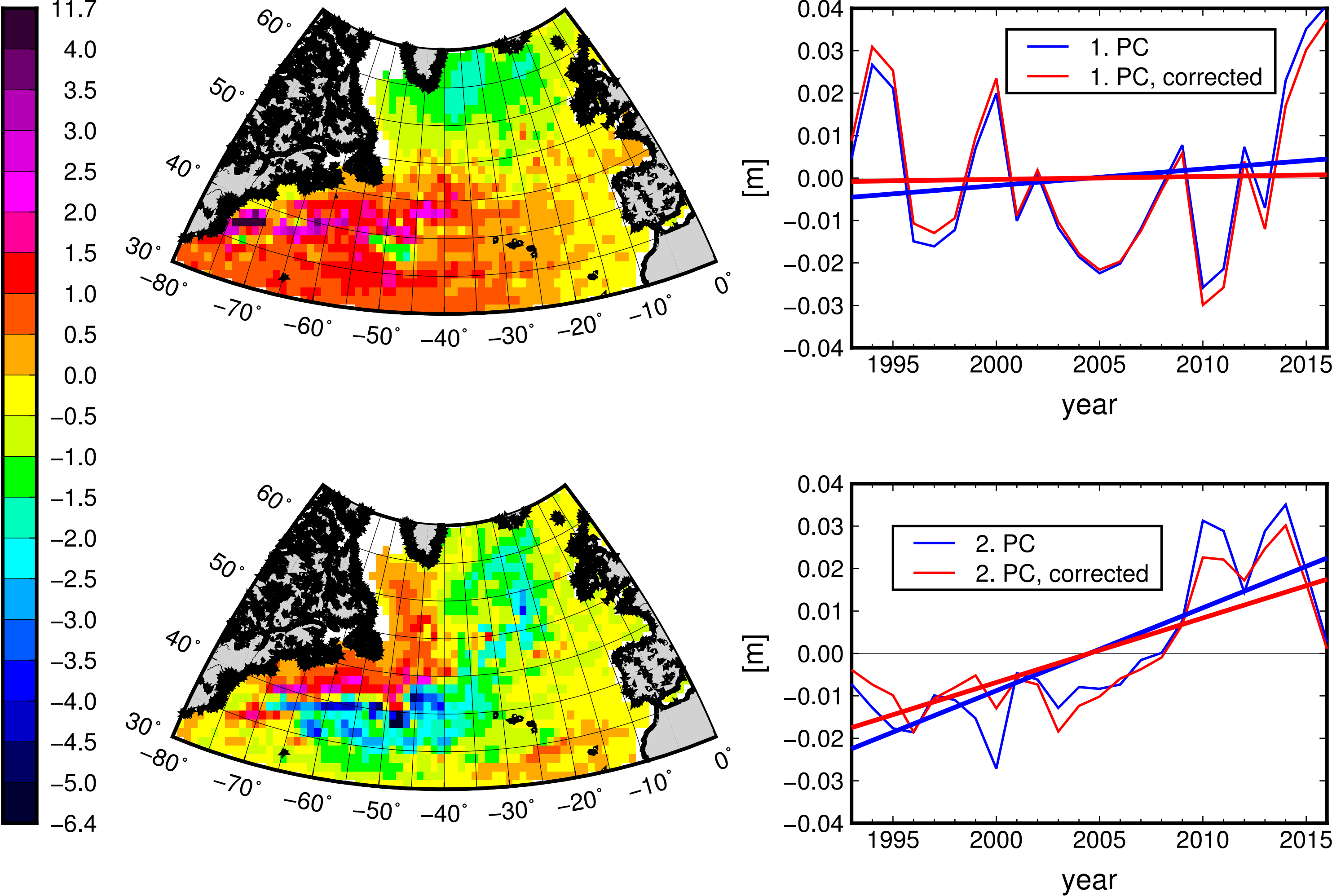
−0.1
0.1
0.2
0.3
0.4
0.5
0.6
1.6

Figure 9.



11.7
4.0
3.5
3.0
2.5
2.0
1.5
1.0
0.5
0.0
−0.5
−1.0
−1.5
−2.0
−2.5
−3.0
−3.5
−4.0
−4.5
−5.0
−6.4
60°
50°
40°
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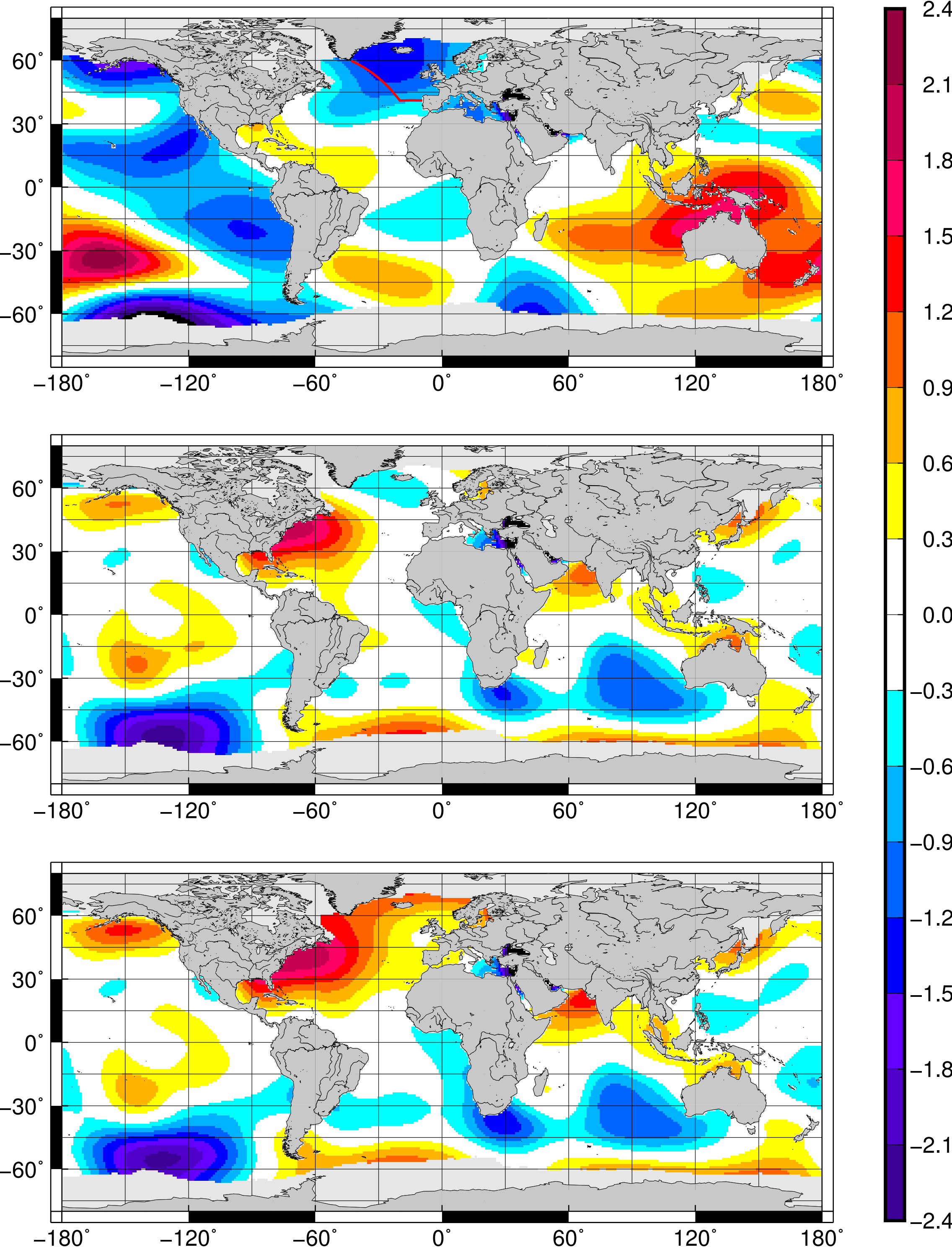
−80°
−70°
−60°
−50°
−40°
−30°
−20°
−10°
0°
1. PC
1. PC, corrected
2. PC
2. PC, corrected
[m]
year
1995
2000
2005
2010
2015
0.04
0.03
0.02
0.01
0.00
−0.01
−0.02
−0.03
−0.04

Figure 10.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2.1
1.8
1.5
1.2
0.9
0.6
0.3
0.0
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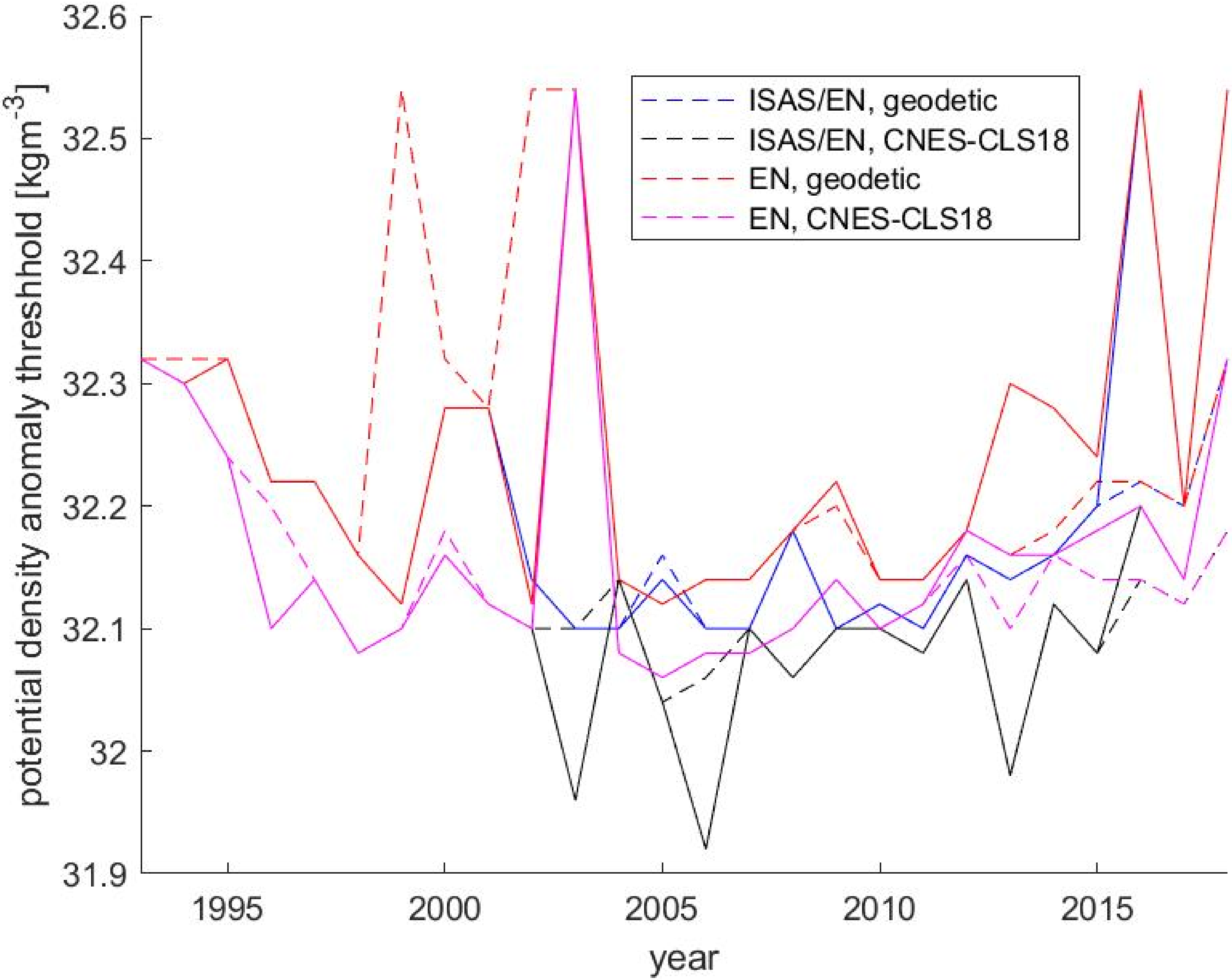
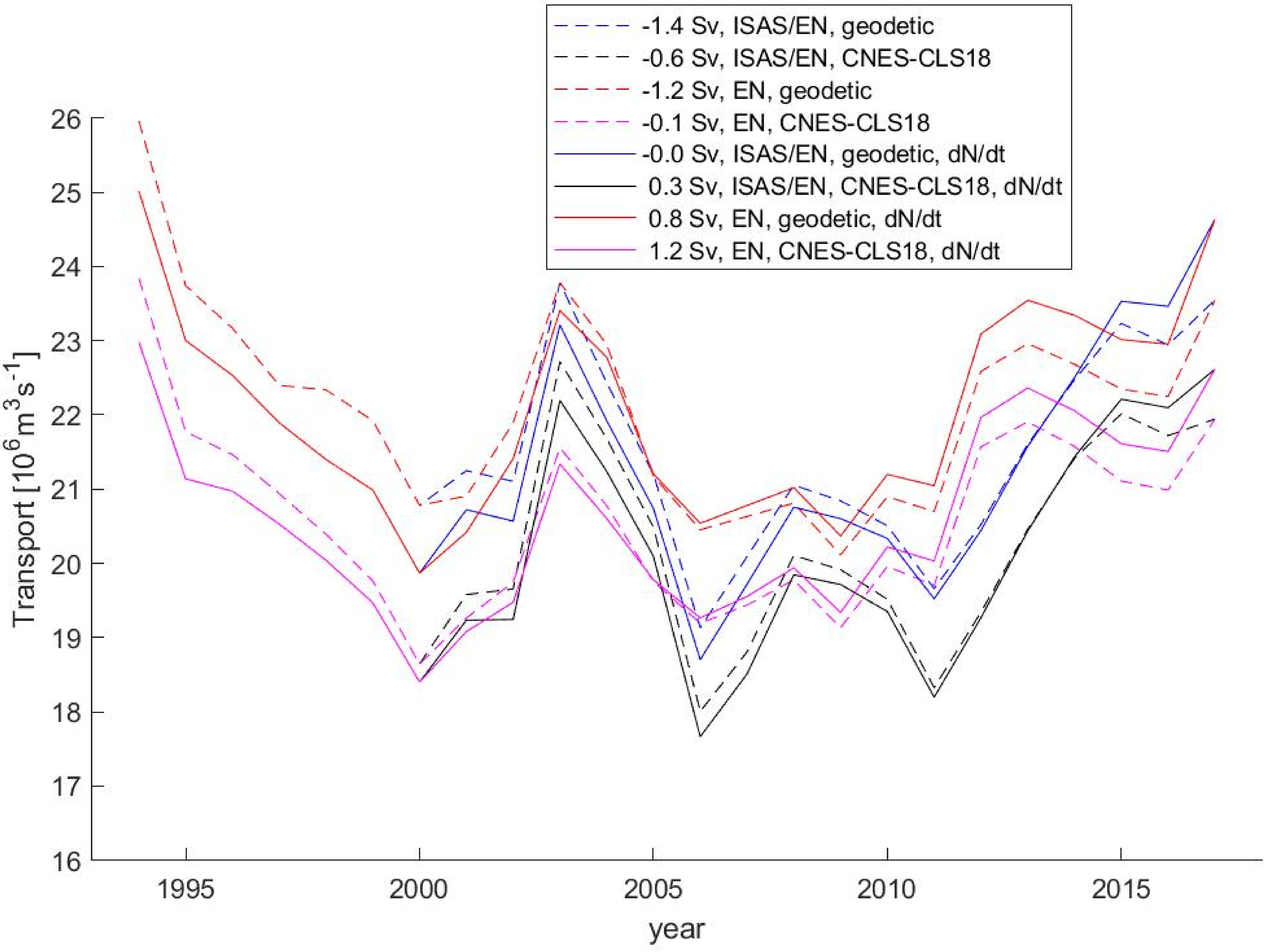
−0.6
−0.9
−1.2
−1.5
−1.8
−2.1
−2.4

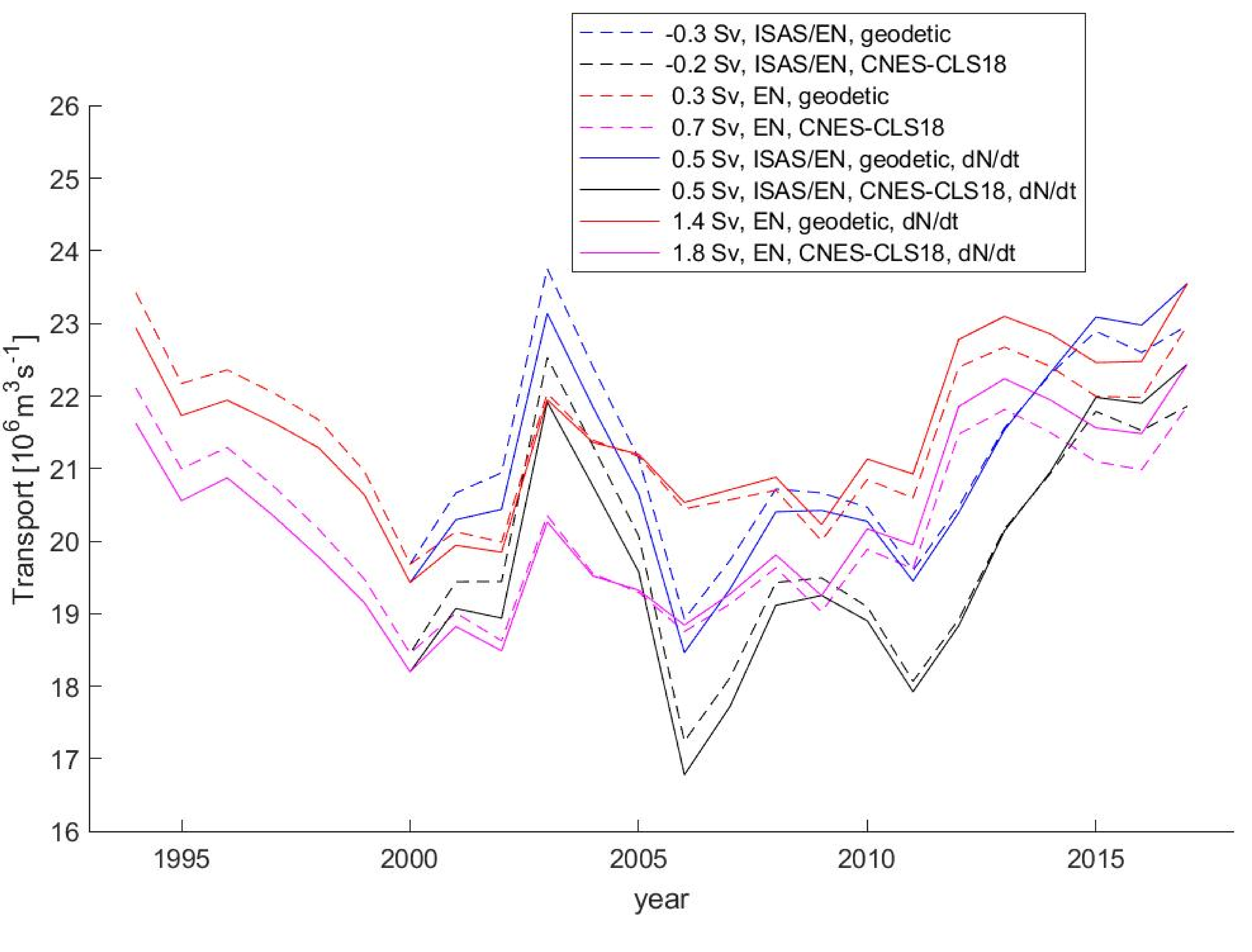
Figure 11.



ISAS/EN, geodetic
ISAS/EN, CNES-CLS18
EN, geodetic
EN, CNES-CLS18
potential density anomaly threshold [kgm^{-3}]
32.6
32.5
32.4
32.3
32.2
32.1
32
31.9
1995
2000
2005
2010
2015
year

Figure 12.


-1.4 Sv, ISAS/EN, geodetic
-0.6 Sv, ISAS/EN, CNES-CLS18
-1.2 Sv, EN, geodetic
-0.1 Sv, EN, CNES-CLS18
-0.0 Sv, ISAS/EN, geodetic, dN/dt
0.3 Sv, ISAS/EN, CNES-CLS18, dN/dt
0.8 Sv, EN, geodetic, dN/dt
1.2 Sv, EN, CNES-CLS18, dN/dt
Transport [10^6m^3s^-1]
year


-0.3 Sv, ISAS/EN, geodetic
-0.2 Sv, ISAS/EN, CNES-CLS18
0.3 Sv, EN, geodetic
0.7 Sv, EN, CNES-CLS18
0.5 Sv, ISAS/EN, geodetic, dN/dt
0.5 Sv, ISAS/EN, CNES-CLS18, dN/dt
1.4 Sv, EN, geodetic, dN/dt
1.8 Sv, EN, CNES-CLS18, dN/dt
Transport [10^6m^3s^-1]
year