The Global Ocean Mass Budget in 1993–2003 Estimated from Sea Level Change

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Abstract

The mass budget of the ocean in the period 1993–2003 is studied with a general circulation model. The model has a free surface and conserves mass rather than volume; that is, freshwater is exchanged with the atmosphere via precipitation and evaporation and inflow from land is taken into account. The mass is redistributed by the ocean circulation. Furthermore, the ocean’s volume changes by steric expansion with changing temperature and salinity. To estimate the mass changes, the ocean model is constrained by sea level measurements from the Ocean Topography Experiment (TOPEX)/Poseidon mission as well as by hydrographic data. The modeled ocean mass change within the years 2002–03 compares favorably to measurements from the Gravity Recovery and Climate Experiment (GRACE), and the evolution of the global mean sea level for the period 1993–2003 with annual and interannual variations can be reproduced to a 0.15-cm rms difference. Its trend has been measured as 3.37 mm yr$^{-1}$ while the constrained model gives 3.34 mm yr$^{-1}$ considering only the area covered by measurements (3.25 mm yr$^{-1}$ for the total ocean). A steric rise of 2.50 mm yr$^{-1}$ is estimated in this period, as is a gain in the ocean mass that is equivalent to an eustatic rise of 0.74 mm yr$^{-1}$. The amplitude and phase (day of maximum value since 1 January) of the superimposed eustatic annual cycle are 4.6 mm and 278° respectively. The corresponding values for the semiannual cycle are 0.42 mm and 120°. The trends in the eustatic sea level are not equally distributed. In the Atlantic Ocean (80°S–67°N) the eustatic sea level rises by 1.8 mm yr$^{-1}$ and in the Indian Ocean (80°S–30°N) it rises by 1.4 mm yr$^{-1}$, but it falls by −0.20 mm yr$^{-1}$ in the Pacific Ocean (80°S–67°N). The latter is mainly caused by a loss of mass through transport divergence in the Pacific sector of the Antarctic Circumpolar Current (−0.42 Sv; Sv = 10$^{6}$ kg s$^{-1}$) that is not balanced by the net surface water supply. The consequence of this uneven eustatic rise is a shift of the oceanic center of mass toward the Atlantic Ocean and to the north.

1. Introduction

The mass budget of the ocean is a key problem in the global hydrological cycle as well as in the understanding of sea level change. Recent ocean volume changes are monitored successfully by altimetry. However, the corresponding mass changes—or bottom pressure variations—can be estimated only by using secular changes in the geoid provided, for example, from the Gravity Recovery and Climate Experiment (GRACE) mission since 2002. In the past many authors, for example, Chen et al. (1998), Minster et al. (1999), Cazenave et al. (2000), and Chambers et al. (2000), attempted to estimate the annual cycle of the global ocean mass balance (eustatic sea level) from altimetric measurements in combination with climatological hydrographic data. Recently Chambers et al. (2004) published the first global ocean mass variations from GRACE. But these data are still too inaccurate and cover a period too short to tell anything about the spatial or temporal variability (other than the annual cycle).

To find a consistent reanalysis of the measured sea level rise and its regional distribution, it is insufficient to apply local corrections in temperature or sea surface height or vertical adjustment (heave). Only an optimization of the forcing of the ocean that leads to sustained circulation changes and thus indirectly to sea level changes can be successful. In the present paper the ocean-state estimation technique employed constrains an ocean general circulation model (OGCM) by data. This offers the possibility of combining altimetric measurements with hydrographic data in a dynamically consistent manner and of looking at the oceans mass balance in more detail, in space as well as in time. Using altimetric and hydrographic data for the period 1993–2003, the regional and global trends in the mass balance...
will be the main topic discussed, thus supplementing the papers mentioned above. The model used and the data are introduced in the next section followed by a comparison of the model results with the data (section 3), that is, to demonstrate how well the model fits the constraints and how it compares in relation to independent data. Section 4 then deals with the mass balance of the ocean itself, and concluding remarks are given in section 5.

2. Model and data

For our purposes, we use the Hamburg Large Scale Geostrophic model (LSG; Maier-Reimer and Mikolajewicz 1991). This model was originally designed for climate studies with time scales of thousands of years (e.g., Maier-Reimer et al. 1993), but in conjunction with its adjoint it has also been used successfully for ocean-state estimation (e.g., Wenzel et al. 2001; Wenzel and Schröter 2002; Hellmer et al. 2005). The model version used in this paper has \(2^o \times 2^o\) horizontal resolution, 23 vertical layers (varying from 20-m thickness for the top layer to 750 m for the deepest ones), and the implicit formulation in time allows for a time step of 10 days.

The model is very suitable for our purposes because it has a free surface and conserves mass rather than volume. The usefulness of the model is further improved by adding the steric effects explicitly to the original coding. The temporal evolution of the sea surface height \(\xi\) is determined as

\[
\frac{\partial}{\partial t} \xi = P - E + R + \nabla \cdot \int_{-H}^{\xi} \mathbf{v} \, d\xi \quad \text{divergence}
\]

\[
+ \int_{-H}^{\xi} \frac{\partial}{\partial z} \left( \frac{1}{\alpha} \right) S_p \frac{\partial}{\partial t} T \, d\xi \quad \text{thermosteric effect}
\]

\[
+ \int_{-H}^{\xi} \frac{\partial}{\partial z} \left( \frac{1}{\alpha} \right) T_p \frac{\partial}{\partial t} S \, d\xi \quad \text{halosteric effect},
\]

where \(\xi\) represents the sea level, \(H\) is the depth, \(P\) is precipitation, \(E\) is evaporation, \(R\) is river runoff, \(T\) is the temperature, \(S\) is salinity, \(p\) is the pressure, \(\alpha = 1/p\) is the specific volume, and \(\mathbf{v}\) is the horizontal velocity.

This offers the ability to estimate the individual contributions to sea level change, and the steric (thermosteric, halosteric) and the eustatic effects (local freshwater balance, mass redistribution) separately. Please note that the model cannot distinguish between the individual contributions to the surface freshwater flux.

The datasets used in the assimilation experiment include the following:

- monthly sea surface temperatures (SST) for the period 1993–2003 (Reynolds et al. 2002);
- gridded fields of 10-day averages of sea surface height anomalies (SSHA) as measured by the Ocean Topography Experiment (TOPEX)/Poseidon altimetric mission for the period 1993–2003, provided by Geoforschungszentrum (GfZ) Potsdam (S. Esselborn 2004, personal communication); these anomalies are combined with the Service Hydrographique et Océanographique de la Marine model (SHOM98.2) mean sea surface height (MSSH; available online at the time of writing at http://www.cls.fr/html/oceano/projets/mss/clshom_en.html) referenced to the EIGEN-GRACE01S geoid (available online at the time of writing at http://www.gfz-potsdam.de/grace/results/grav/g001_eigen-grace01s.html) to give absolute sea surface height values;
- temporal mean transports of mass, freshwater, and heat as obtained by different authors and as they are summarized by, for example, Bryden and Imawaki (2001) and Wijffels (2001); the transport constraints are not applied for the Antarctic Circumpolar Current (ACC);
- the mean annual cycle of temperatures, salinities, and horizontal velocities on two sections in the Weddell Sea area—one running across the inner part of the sea and the other on the crest of the South Scotia Ridge; these data are taken from a high-resolution model of the Weddell Sea (Schodlok et al. 2002) whose water mass characteristics and circulation are in good agreement with local observations; and
- the climatological mean temperatures and salinities from the World Ocean Circulation Experiment (WOCE) Global Hydrological Climatology (WGHC; Gouretski and Koltermann 2004) in combination with the mean annual cycle from the most recent World Ocean Atlas 2001 (WOA01; Conkright et al. 2002); these data are supplied to the assimilation procedure with small weights and thus serve only as background information.

In the present paper we do not use any subsurface data other than the aforementioned background climatology. This work is a first step toward the use of subsurface data. The assimilation of actual temperatures, for example, from Willis et al. (2004), will be done in the next experiment. At present we retain these data for verification.

There is no direct constraint on the surface freshwater flux in our model. A direct constraint on this flux seems to be inadequate because any data on \(P - E +\)
are orders of magnitude more uncertain than what is required here. In this experiment the surface freshwater flux largely plays the role of a residual that is constrained indirectly by the use of the SSHA and SST data. In general, the freshwater flux would be able to compensate for any change in the SSHA that cannot be explained by steric effects. But this flux not only changes the volume of the model’s uppermost layer but the surface salinity as well by keeping the amount of salt in this layer constant. This sets a rough limit on the estimated freshwater fluxes because salinity is constrained to climatology. In addition, any eustatic sea level change would cause a halosteric change of the same sign that cannot be neglected locally (e.g., Antonov et al. 2002; Wenzel and Schröter 2002). This gives a further limitation to the estimated flux.

To adjust the model to the data, the adjoint method is employed, which is a variational optimization method. The control parameters of this optimization are the model’s initial temperature and salinity state as well as the forcing fields (wind stress, air temperature, and surface freshwater flux), whereas the first-guess forcing is taken from the monthly National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis fields. In summary, the experiment WEDD as analyzed in this paper is an update of the WEDEX experiment described in Hellmer et al. (2005). A more detailed description of the assimilation procedure can be found therein.

3. Model–data comparison

The local differences between the model’s temporal mean SSH and the data are shown in Fig. 1. In most parts of the ocean the deviations are well below 5 cm, giving a global rms value of 11.3 cm. The largest deviations (up to ±30 cm) are found in the regions with strong currents, that is, the western boundary currents as well as the ACC. In particular, the signature in the ACC region implies that these currents are represented too broadly by the model. The temporal rms differences between the modeled SSH and the data are shown in Fig. 2. The global rms value, which is the measure of success in the assimilation, is 2.8 cm; although locally we find higher rms values (up to 7 cm), especially in the tropical Pacific and in the western boundary currents. For the surface temperature (not explicitly shown) the corresponding rms differences between the model and the data are 0.30 K for the temporal mean and 0.51 K for the anomalies.

Figure 3a shows that the optimized model reproduces the global mean sea level data well (rms of difference: 0.15 cm). This is especially true for the interannual variability, while the amplitude of the annual cycle is slightly underestimated by the model. The latter becomes even more apparent on the local scale (not shown) and appears to be a general shortcoming of the OGCM used, which leads to the high rms values apparent in Fig. 2. Figure 3a also shows that the linear trend in the global sea level change (3.25 mm yr⁻¹) originates mainly from the steric contribution (2.50 mm yr⁻¹) while the eustatic contribution (0.74 mm yr⁻¹) adds only a minor part. This is an improvement as compared with the first-guess model results (Fig. 4), that is, using NCEP–NCAR reanalysis forcing fields. In this case the modeled total sea level trend (4.7 mm yr⁻¹) splits into 1.7 mm yr⁻¹ for the steric contribution and 3.0 mm yr⁻¹ for the eustatic contribution. After the optimization, the eustatic trend (0.74 mm yr⁻¹) corresponds well to the 0.87 mm yr⁻¹ value that can be derived by adding
together the estimates reported in Cazenave and Nerem (2004): 0.25 mm yr$^{-1}$ sea level rise from land water and snowmass, 0.1 mm yr$^{-1}$ from mountain glaciers, 0.13 mm yr$^{-1}$ from Greenland, and 0.22 mm yr$^{-1}$ from Antarctica. This improvement is directly related to the changes in the mean surface freshwater flux (Fig. 5a), whose spatial structure does not appear to be much different from the first guess, that is, the NCEP–NCAR-derived flux. To accentuate the changes, the difference from the first guess is shown in Fig. 5b. Most of the global ocean shows negative changes, that is, more evaporation, but there are also pronounced positive changes, especially in the Atlantic.

In contrast to its minor importance in the trend, the global eustatic sea level resamples nearly all the “short term” temporal variability of the global mean sea level (Fig. 3a), while the steric contribution appears more or
less as a straight line. Nevertheless, we also find a small annual cycle in the steric part, which appears to be in antiphase with the eustatic.

On the global scale the steric contribution to the sea level rise is mainly caused by the thermosteric effect (Fig. 3b) with a positive trend stemming from all layers. The halosteric part (Fig. 3c) implies a redistribution of salt from the deeper layers to the top. For the total volume, this reflects the global freshwater balance from precipitation, evaporation, and runoff (see e.g., Wadhams and Munk 2004), but it is of minor importance. However, it cannot be neglected regionally or even locally (Antonov et al. 2002; Wenzel and Schröter 2002).

One possibility for assessing the global thermosteric sea level rise is to compare the ocean’s heat content anomalies with independent data. Figure 6 shows the modeled heat content anomaly for the global ocean, whose total trend corresponds to a 1.5 W m⁻² surface heat flux. Within the top 500 m the trend compares well.
to the trend estimated from the analysis of Willis et al. (2004), although no explicit subsurface temperature data are given in the assimilation. Further confidence in our results is obtained by the good correspondence in the spatial distribution as well as in the size of the local trends in upper ocean heat content between our results (Fig. 7) and those of Willis et al. (2004). Figure 6 also indicates that the deeper layers make an essential contribution to the total thermosteric sea level rise, which is confirmed by the results shown in Levitus et al. (2005). Regardless of this, the trends might be due to a still-existing artificial model drift in our case, although the assimilation includes a constraint to minimize deep-ocean interannual variability. This constraint reduces the thermosteric sea level trend stemming from the deep ocean (below 2250 m) by a factor of 2 and the halosteric trend by a factor of 10 as compared with the first guess. Regardless, the deep layers should not be neglected in general when estimating the ocean’s water mass budget from sea level change and temperature measurements especially on long time scales. This type of shortcoming might be justified when investigating the mean annual cycle only, as in, for example, Chen et al. (1998), Minster et al. (1999), Cazenave et al. (2000), or Chambers et al. (2000). Looking at longer periods, temperature and salinity changes might be small in the deep layers but they are related to a large volume that amplifies their influence on the sea level (see Figs. 3b and 3c).

The modeled mass variations (eustatic part, green curve in Fig. 3a) are well represented by the corresponding variations in the bottom pressure field. These variations should be detectable through variations in the geoid estimated, for example, from the GRACE mission once the measurements have been fully analyzed. The available GRACE data (F. Flechtner 2004, personal communication) are still preliminary and should be treated with caution. Nevertheless, Chambers et al. (2004) show that the global eustatic sea level variations detected by GRACE fit well with the mean

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**Fig. 7.** Local trend of the upper-ocean heat content for ζ-700 m. Cf: 2 W m⁻².

**Fig. 8.** Area mean bottom pressure anomalies as compared with the GRACE geoid variations (given in centimeters water equivalent) for (a) the global ocean, (b) the Atlantic (55°S-65°N), (c) the Pacific (55°S-65°N), and (d) for the western part of the tropical Pacific (20°S-20°N, 140°E-150°W).
annual cycle deduced from TOPEX measurements (corrected for the steric effect using WOA01 data). Here, we do a comparison on different scales using the last 2 yr of the modeled bottom pressure. Examples are given in Fig. 8 for the global ocean, the total Atlantic Ocean, the total Pacific Ocean, and for the western part of the tropical Pacific. We find good correspondence in amplitude and phase between the modeled bottom pressure variations and the GRACE data (given in centimeters water equivalent) for the global ocean (Fig. 8a). The correspondence diminishes when looking at smaller areas and even becomes unacceptable on scales like, or example, the western tropical Pacific (Fig. 8d). The poor correspondence on more local scales is not directly related to the rms errors in the retrieved SSHA (Fig. 2). Nevertheless, it might be due to the above-
4. Ocean mass balance

The good correspondence of the model results to the Willis et al. (2004) analysis, as well as to the GRACE data, on largest scales gives confidence as we look into the model results in more detail. Figure 9 shows the model’s local sea level trends to be split into the total steric contribution (Fig. 9b) and the eustatic part (Fig. 9c). Locally as well as in the global mean, most of the trend (Fig. 9a) is due to the steric contribution while the eustatic part is much smaller but not negligible. Corresponding to the mean freshwater flux into the ocean, the global mean eustatic sea level trend is 0.74 mm yr\(^{-1}\), which is about \(1/3\) of the steric trend (2.50 mm yr\(^{-1}\)). Furthermore, in comparison with the steric part (Fig. 9b) the eustatic trends do not show much spatial variability. Throughout the Atlantic and the Indian Oceans the eustatic trends are positive on a fairly constant level (\(\sim 2\) mm yr\(^{-1}\)) while they are near zero or slightly negative in most parts of the Pacific (see also Table 1). The most conspicuous features in Fig. 9c are the strong negative trends west of Drake Passage leaking into the Scotia Sea (down to \(-20\) mm yr\(^{-1}\)).

The global ocean mass exhibits a strong seasonal cycle (Fig. 10), and an interannual variability is visible with a slight decrease in mass during 1993–95 followed by a stronger rise. This results in an overall positive trend for the total period (1993–2003) that is consistent with the global mean of the eustatic trends (Fig. 9c). This reveals that the magnitude of the trends estimated for this period could be the consequence of looking only at a part of a longtime oscillation, at least in part. Using all 1993–2003 model data, with the longer periods (gray curve in Fig. 10) removed from the time series, we find a mean annual cycle of \(1.66 \times 10^{15}\) kg in amplitude and \(278.4^\circ\) in phase. The corresponding values for the semiannual cycle are \(0.15 \times 10^{15}\) kg and \(119.6^\circ\), respectively. While the phases (defined as the day of maximum value since 1 January) fit well with earlier estimates by Chen et al. (1998), Cazenave et al. (2000), and Chambers et al. (2004), our amplitudes for the corresponding eustatic sea level (4.6 and 0.42 mm, respectively) appear to be smaller than theirs, which obviously is due to the different methods/time periods used for the estimates.

The residuals of the mean mass balance for the single ocean basins are small as compared with the corresponding transport values (Table 1, Fig. 11), but they represent the regional eustatic changes well. The mass balances are positive for all basins (the basins gain mass) except for the South Pacific, that is balanced (\(-0.04\) mSv; 1 mSv = \(10^6\) kg s\(^{-1}\)), and especially for the Pacific sector of the ACC (\(-2.35\) mSv). In the latter area the divergence in the ocean mass transport (\(-0.420\) Sv; Sv = \(10^6\) kg s\(^{-1}\)) is not totally compensated for by the net freshwater supply to the ocean through the surface (0.417 Sv), which leads to the aforementioned anomalous strong eustatic sea level fall. Looking at the other basins in more detail, one finds that the mass surplus results from different balances: in the tropical region of the Indian Ocean and the Pacific the net input by precipitation is not compensated for by the mass divergence, while for the northern and southern ocean basins the mass convergence is not balanced by net evaporation. Within the ACC there is net precipitation in all sectors overcompensating the mass transport divergence (except for the Pacific sector; see above). In summary, most of the mass gain of the ocean goes to the Northern Hemisphere and especially to the Atlantic Ocean. This also becomes apparent through
the changes in the position of the ocean’s center of mass. At the end of the model integration the center of mass is displaced by \(X = 771\) cm, \(Y = 213\) cm, and \(Z = 573\) cm. In contrast, adding the final mass surplus of \(3 \times 10^{15}\) kg (Fig. 10) uniformly to the ocean \((\Delta z = 0.8\) cm\) would shift the oceanic center of mass by about \(X = -42\) cm, \(Y = -17\) cm, and \(Z = -45\) cm, that is, much less and into the opposite direction.

![ocean mass fluxes](image)

**Table 1. Mean oceanic mass balance for 1993–2003.**

<table>
<thead>
<tr>
<th>Ocean basin</th>
<th>Inflow (Sv)</th>
<th>Outflow (Sv)</th>
<th>In + out (Sv)</th>
<th>Surface flux (Sv)</th>
<th>Total balance (mSv)</th>
<th>Eustatic trend (mm yr(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Arctic</td>
<td>0.407</td>
<td>-0.532</td>
<td>-0.125</td>
<td>0.126</td>
<td>1.06</td>
<td>2.19</td>
</tr>
<tr>
<td>Tropical Indian</td>
<td>9.463</td>
<td>-9.610</td>
<td>-0.148</td>
<td>0.149</td>
<td>1.61</td>
<td>1.70</td>
</tr>
<tr>
<td>South Indian</td>
<td>9.610</td>
<td>-9.195</td>
<td>0.415</td>
<td>-0.414</td>
<td>0.91</td>
<td>1.81</td>
</tr>
<tr>
<td>ACC–Indian</td>
<td>136.09</td>
<td>-136.16</td>
<td>-0.066</td>
<td>0.067</td>
<td>0.83</td>
<td>0.84</td>
</tr>
<tr>
<td>Total Indian</td>
<td>136.36</td>
<td>-136.16</td>
<td>0.201</td>
<td>-0.198</td>
<td>3.35</td>
<td>1.37</td>
</tr>
<tr>
<td>North Pacific</td>
<td>0.600</td>
<td>-0.407</td>
<td>0.193</td>
<td>-0.192</td>
<td>0.40</td>
<td>0.24</td>
</tr>
<tr>
<td>Tropical Pacific</td>
<td>9.493</td>
<td>-10.063</td>
<td>-0.569</td>
<td>0.570</td>
<td>0.87</td>
<td>0.46</td>
</tr>
<tr>
<td>South Pacific</td>
<td>10.014</td>
<td>-9.493</td>
<td>0.521</td>
<td>-0.521</td>
<td>-0.04</td>
<td>-0.03</td>
</tr>
<tr>
<td>ACC–Pacific</td>
<td>136.16</td>
<td>-136.57</td>
<td>-0.420</td>
<td>0.417</td>
<td>-2.35</td>
<td>-2.25</td>
</tr>
<tr>
<td>Total Pacific</td>
<td>136.16</td>
<td>-136.43</td>
<td>-0.275</td>
<td>0.274</td>
<td>-1.12</td>
<td>-0.20</td>
</tr>
<tr>
<td>North Atlantic</td>
<td>0.532</td>
<td>-0.387</td>
<td>0.145</td>
<td>-0.143</td>
<td>2.46</td>
<td>2.17</td>
</tr>
<tr>
<td>Tropical Atlantic</td>
<td>0.340</td>
<td>-0.245</td>
<td>0.095</td>
<td>-0.093</td>
<td>1.92</td>
<td>2.71</td>
</tr>
<tr>
<td>South Atlantic</td>
<td>0.245</td>
<td>-0.076</td>
<td>0.169</td>
<td>-0.168</td>
<td>0.91</td>
<td>2.47</td>
</tr>
<tr>
<td>ACC–Atlantic</td>
<td>126.63</td>
<td>-126.90</td>
<td>-0.257</td>
<td>0.258</td>
<td>0.39</td>
<td>0.43</td>
</tr>
<tr>
<td>Total Atlantic</td>
<td>127.09</td>
<td>-126.95</td>
<td>0.152</td>
<td>-0.146</td>
<td>5.68</td>
<td>1.82</td>
</tr>
<tr>
<td>Global</td>
<td>0.000</td>
<td>-0.047*</td>
<td>-0.047</td>
<td>0.056</td>
<td>8.97</td>
<td>0.76</td>
</tr>
</tbody>
</table>

* The global outflow goes to the Mediterranean Sea (not included in the Atlantic balance).

The changes in the position of the ocean’s center of mass. At the end of the model integration the center of mass is displaced by \(X = 771\) cm, \(Y = 213\) cm, and \(Z = 573\) cm. In contrast, adding the final mass surplus of \(3 \times 10^{15}\) kg (Fig. 10) uniformly to the ocean \((\Delta z = 0.8\) cm) would shift the oceanic center of mass by about \(X = -42\) cm, \(Y = -17\) cm, and \(Z = -45\) cm, that is, much less and into the opposite direction.
5. Conclusions

The oceans mass budget for the period 1993–2003 is estimated by constraining the LSG ocean model with hydrographic data, mean ocean transport estimates, and sea level data derived from the TOPEX/Poseidon mission. A comparison with independent data, upper-ocean heat content (Willis et al. 2004), and preliminary bottom pressure estimates from GRACE shows that the constrained model works fairly well.

The global ocean mass exhibits a pronounced annual cycle (amplitude of $1.66 \times 10^{15}$ kg; phase of 278.4°) that is superimposed by interannual variability. The resulting positive trend, valid for the chosen period 1993–2003, corresponds to a 0.74 mm yr$^{-1}$ eustatic global sea level rise. On local or regional scales the mass changes (eustatic sea level changes) are the residual of the horizontal mass transport divergence and the surface fluxes. In comparison with the steric changes, the eustatic sea level varies on very large scales. In summary, there is net mass gain in all basins, except for the Pacific sector of the ACC, but this gain is not evenly distributed. It is stronger in the Northern than in the Southern Hemisphere and stronger in the Atlantic and Indian Oceans than in the Pacific. This unequal distribution of mass gain in the single basins appears to be caused mainly by the internal redistribution of mass.

Although global analyses like that of Willis et al. (2004) have their own deficiencies, the next step will be to include more subsurface information directly into the assimilation scheme to improve the modeled steric sea level change and by this to upgrade the eustatic estimates. Furthermore, once the GRACE data have been improved, an alternate approach will be possible too: to improve the ocean heat content estimates, steric sea level variations by constraining the bottom pressure variations. This method was proposed by, for example, Jayne et al. (2003) and appears to be a necessary path to follow because direct hydrographic measurements are sparse in space and time.

Acknowledgments. The authors thank Carl Wunsch for his inspiring ideas. Inverse modeling, data assimilation, and satellite altimetry started for us more than 20 years ago based on and guided by his pioneering work and lead finally to this paper. Furthermore, we acknowledge Saskia Esselborn (GfZ Potsdam) for providing the reprocessed TOPEX/Poseidon data and Frank Flechtner (GfZ Potsdam) for providing the preliminary GRACE data. The ocean center of mass was calculated using a routine from IERS kindly provided by Richard Gross (JPL).


