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Global runoff over 1993–2009 estimated from coupled land-ocean-atmosphere water budgets and its relation with climate variability

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Abstract

Whether the global runoff (or freshwater discharge from land to the ocean) is currently increasing and the global water cycle is intensifying is still a controversial issue. Here we compute land-atmosphere and ocean-atmosphere water budgets and derive two independent estimates of the global runoff over the period 1993–2009. Water storage variations in the land, ocean and atmosphere reservoirs are estimated from different types of datasets: atmospheric reanalyses, land surface models, satellite altimetry and in situ ocean temperature data (the difference between altimetry based global mean sea level and ocean thermal expansion providing an estimate of the ocean mass component). Results for the global runoff from the two methods show a very good correlation between both estimates. More importantly, no significant trend is observed over the whole period. Besides, the global runoff appears to be clearly impacted by large-scale climate phenomena such as major ENSO events. To infer this, we compute the zonal runoff over four latitudinal bands and set up for each band a new index (Combined Runoff Index) obtained by optimization of linear combinations of various climate indices. Results show that, in particular, the intertropical and northern mid-latitude runoffs are mainly driven by ENSO and the Atlantic Multidecadal Oscillation (AMO) with opposite behavior. Indeed, the zonal runoff in the intertropical zone decreases during major El Niño events whereas it increases in the northern mid-latitudes, suggesting that water masses over land are shifted northward/southward during El Niño/La Niña. In addition to this study, we propose an innovative method to estimate the global ocean thermal expansion. The method is based on the assumption that the difference between both runoff estimates is mainly due the thermal expansion term not accounted for in the estimation of the ocean mass. Comparison of our reconstructed thermal expansion with two existing datasets shows the relevance of this new method.

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(ground based and remotely sensed), it may suffer from large uncertainties in some sparsely monitored regions. Moreover, very few direct measurements of evapotranspiration exist (flux towers) and modeling E at the global scale is subject to large uncertainties (Vinukollu et al., 2011). The atmospheric water budget has been introduced by some authors (e.g. Oki et al., 1995; Oki, 1999; Syed et al., 2009) to overcome difficulties in estimating P and E . Dai and Trenberth (2002) showed that the use of the atmospheric water budget improved model based estimates of global runoff.

For the past decade, the ocean mass and the terrestrial water storage can be provided by the Gravity Recovery and Climate Experiment (GRACE) (Tapley et al., 2004; Wahr et al., 2004) space gravimetry mission, as done for instance by Syed et al. (2009) for S_l . Before its launch in 2002, no direct measurement of S_l and S_o were available and since it was difficult to validate estimates from models, variations in water storage were usually neglected (e.g. Oki, 1999; Dai and Trenberth, 2002) then leading to an estimation of the global runoff directly from the net precipitation ($P - E$). Nevertheless Syed et al. (2009) showed the importance of taking this term into account by using GRACE estimates of S_l .

GRACE products are available since 2002, which may limit the time span of the study to the last decade. Alternatively, Land Surface Models (LSMs) provide monthly estimations of S_l with a satisfactory accuracy at basin to global scales, as shown in the various studies comparing GRACE and LSMs (see Ramillien et al., 2008 for a review of studies prior to 2008). The covered period of LSMs depends on the model, but simulations generally run at least over the last two decades. Besides, the ocean mass S_o may be derived from satellite radar altimetry observations which provide estimations of the global mean sea level (GMSL) and from in situ hydrographic data. To derive the ocean mass variations, GMSL has to be corrected from the steric component (effect of temperature and salinity), as done by e.g. Syed et al. (2010).

In this study, we use the coupled land-ocean-atmosphere water budgets to estimate the interannual variability of global runoff. Two estimates are computed: one from the land/atmosphere coupling (Eq. 4) and the other from the ocean/atmosphere coupling

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(Eq. 5).

$$R_l = -\frac{\partial W_l}{\partial t} - \text{div}(Q_l) - \frac{dS_l}{dt} \quad (4)$$

$$R_o = \frac{\partial W_o}{\partial t} + \text{div}(Q_o) + \frac{dS_o}{dt} \quad (5)$$

S_l is estimated from three LSMs, S_o from altimetry based GMSL and the net precipitation term (time derivative of W and $\text{div}(Q)$) from atmospheric reanalyses. Since altimetry observations are used, our global runoff estimates cover the altimetry time span (1993–2009). This study expands the previous ones by providing for the first time a comparison of global runoff estimates from land and ocean water budgets, in terms of interannual variability, over the last two decades.

Section 2 presents the data sets used in this study. In Sect. 3, each data set is compared with independent data to ensure its reliability and to give an idea of its uncertainties. The comparison between both global runoff estimates, in terms of interannual variability, is given in Sect. 4. Our global runoff estimate is also compared with global climate indices (ENSO related SOI, AMO). A discussion on the ocean thermal expansion used in the estimation of the ocean mass is provided in Sect. 5.

2 Data and models used in this study

In this section, we present the data and models used to compute global runoff by the two methods (R_l and R_o). Data used for validation purposes are also presented.

2.1 Altimetry-based sea level data

For the altimetry-based sea level data, we use the DT-MSLA “Ref” series provided by Collecte Localisation Satellite (CLS; <http://www.avisioceanobs.com/en/data/products/sea-surface-height-products/global/msla/index.html>). This data set is used over the

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time span from January 1993 to December 2009. It is available as $1/4^\circ \times 1/4^\circ$ Mercator projection grids at weekly interval from a combination of several altimetry missions (Topex/Poseidon, Jason-1 and 2, Envisat and ERS 1 and 2). Most recently improved geophysical corrections are applied to the sea level data (see Ablain et al., 2009, for details). The weekly sea level grids are geographically averaged between 65° S and 65° N to obtain a GMSL time series. The data are further averaged on a monthly basis.

2.2 Steric data

Steric sea level is estimated using the updated in situ ocean temperature and salinity data from Ishii and Kimoto (2009), v6.12 (called hereafter IK09). The IK09 temperature data are corrected for the XBT depth bias. The temperature and salinity data are available at monthly interval over 16 depth levels ranging from the ocean surface down to 700 m depth, on a global $1^\circ \times 1^\circ$ grid from 1955 to 2009. Steric sea level anomalies are computed over the 0–700 m depth range for the period January 1993 to December 2009. The deep ocean contribution cannot be accounted for since hydrographical data below 700 m are too sparse. Recent studies have shown that in terms of trend the deep ocean contributes by $\sim 10\%$ to the total steric effect (Church et al., 2011). Besides, almost all interannual variability in steric sea level is confined in the upper 300–500 m of the ocean (e.g. Llovel et al., 2011). At global scale, salinity does not contribute to the GMSL and is therefore neglected in the following.

2.3 Land surface models

To estimate the terrestrial water storage component (S_l) we use monthly gridded outputs of three different LSMs: (1) the WaterGAP Global Hydrology Model (WGHM) (Döll et al., 2003); (2) the Interactions between Soil, Biosphere and Atmosphere – Total Runoff Integrating Pathways (ISBA-TRIP) model (Alkama et al., 2011; Decharme et al., 2010); (3) the Land Dynamics (LaD) model (Milly and Shmakin, 2002). WGHM outputs are available through December 2009 whereas ISBA ends in December 2008 and LaD

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in July 2007. Discrepancies among S_l derived from the different LSMs may come from differences in the numerical schemes and meteorological forcing. As noted by Syed et al. (2009), discrepancies among models outputs provide an estimation of the model uncertainties.

Independently, the evapotranspiration modeled by the three LSMs is used for the validation of net precipitation over land ($P_l - E_l$) computed from the atmospheric water budget (Eq. 3).

2.4 GRACE data

To validate LSMs derived S_l and altimetry-based S_o , a comparison with GRACE data is proposed in Sect. 3. Here we use GRACE products (release 2) for the period 2003–2009 (with missing data for June 2003), computed by the Groupe de Recherche de Géodesie Spatiale (GRGS) (Bruinsma et al., 2010). It consists of monthly $1^\circ \times 1^\circ$ gridded time series of water volume (S_l or S_o), expressed in terms of Equivalent Water Height (EWH). At each grid mesh, the volume anomalies are obtained by removing the temporal mean. The GRGS data are stabilized during the generation process so that no smoothing or filtering is necessary. When GRACE data is used at basin scales, it has to be corrected from leakage effects due to its low resolution (Longuevergne et al., 2010). Nevertheless such effects have minor impact at the global scale and GRACE data is not corrected here.

As said previously, GRACE provides reliable estimations of spatiotemporal water volume variations since 2002 and it would have been possible to complete GRACE data with LSMs outputs and altimetry based ocean mass for the period 1993–2002. In order to keep consistency in our computations over 1993–2009, we prefer not to use GRACE products to estimate terrestrial water storage and ocean mass variations. Nevertheless, in both cases, comparison with GRACE observations is performed to increase confidence in data.

For each zone, the zonal runoff is then compared with different linear combinations of several climate indices. The indices considered here are the following: the Multivariate ENSO Index (MEI), the Atlantic Multidecadal Oscillation (AMO), the Pacific Decadal Oscillation (PDO), the Pacific-North America teleconnection (PNA) and the Arctic Oscillation index (AO). The reader may refer to Rossi et al. (2011) for a detailed presentation of these indices. For each subset of one or two indices, the linear combination is optimized by maximizing the correlation between the zonal runoff and the combined index. Subsets that give the best results are used to compute a new index called Combined Runoff Index (CRI). Optimization results are given, for each zonal band, by Eq. (7) (climate indices have been normalized before the optimization).

$$\begin{aligned}
 \text{CRI}(-60/ -20) &= 0.53 \times \text{PDO} + 0.47 \times \text{AMO} \\
 \text{CRI}(-20/ +20) &= -0.62 \times \text{MEI} - 0.38 \times \text{AMO} \\
 \text{CRI}(+20/ +60) &= 0.54 \times \text{MEI} + 0.46 \times \text{AMO} \\
 \text{CRI}(+60/ +90) &= 0.73 \times \text{PNA} + 0.27 \times \text{AO}
 \end{aligned} \tag{7}$$

Since LSMs and ERA-Interim outputs are available since 1980 or earlier (apart from WGHM which is only available since 1992), we also compared CRI and zonal R_I over the period 1980–1992, period not used in the calibration of CRI. Figure 9 shows (dashed lines) results of calibration (period 1993–2009) and validation (period 1980–1992); CRI has been normalized to match the range of zonal runoffs variability. The two numbers in the brackets in the legend represent the correlation between zonal R_I and CRI for the calibration period and for the overall period, respectively. Figure 9 clearly shows a very good correlation for the inter-tropical zone (namely during the validation period) and more contrasted correlations in mid- and high-latitudes.

Not surprisingly, MEI contributes for the most part in the inter-tropical and northern mid-latitude zonal runoffs. For these two zonal runoffs, AMO also plays an important role. Figure 9 and Eq. (7) show that these two zonal runoffs are highly anti-correlated, namely with higher/lower runoffs than normal in mid-latitude/inter-tropical zones during

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major El-Niño events (e.g. in 1983 or in 1998). The reciprocal is true for La-Niña events (e.g. in 1989 or in 2000). This suggests that during El-Niño events, while water mass is shifted westward from the South-American continent to the tropical Pacific Ocean (Gu et al., 2007), it is also shifted northward to mid-latitude continents. Besides, ENSO seems to play a less important role in northern high latitudes and southern mid latitudes. Northern high latitude zonal runoff is logically governed by northern mid to high latitude climate phenomena (Pacific-North America teleconnection and Arctic Oscillation). For each of the three other zones, CRI is a combination of a climate index related to the Pacific Ocean and another related to the Atlantic Ocean.

Further investigations are suggested to complete these preliminary results, namely about the relationship between the zonal runoff and climate indices characteristics in the frequency domain (Rossi et al., 2011).

5 Reconstruction of the ocean thermal expansion

In this subsidiary section, we come back to the aforementioned problem of the ocean thermal expansion (TE) correction and propose an innovative method to reconstruct this component of the global mean sea level. Indeed, as said previously, the observed difference between R_I and R_o may be mainly attributed to TE. Assuming that $R_I - E_I = -(P_o - E_o)$ (see Sect. 3.5), $R_o = R_I$ leads to a time invariant interannual variability of $S_o + S_I$. Here we propose to use this property to give a new estimation of TE from Eq. (8).

$$\text{TE} = \text{GMSL} + S_I \tag{8}$$

We further compare our TE reconstruction with TE data from two different data bases: the IK09 (presented in Sect. 2.2) and the WOD09 (Levitus et al., 2009). The WOD09 and IK09 data bases account for depth-bias corrections on XBT temperature data (e.g. Wijffels et al., 2008). The TE data from the two data bases are publicly available at: <http://www.noaa.gov/> for WOD09 and <http://atm-phys.nies.go.jp/~ism/pub/ProjD/v6.9/>

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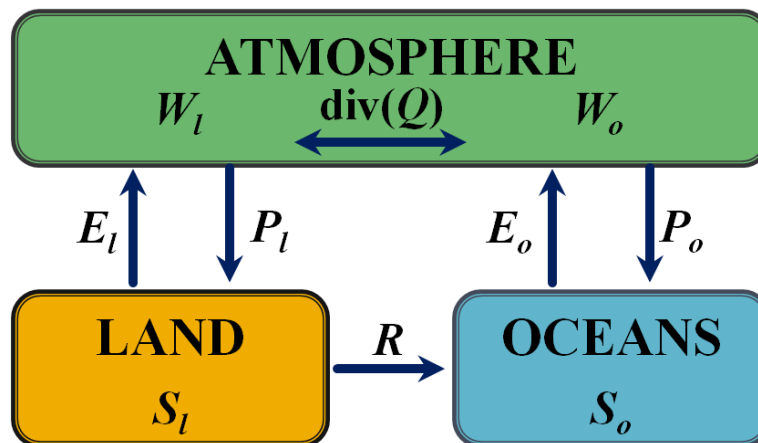


Fig. 1. Stocks and fluxes in the global water cycle. Refer to the text for a detailed description of the variables. S_l represents the terrestrial water storage, S_o the ocean mass, R the global runoff, P the precipitation, E the evapo(transpi)ration, W the total column water vapor and $\text{div}(Q)$ the divergence of the vertically integrated water vapor flux. Subscripts l and o designate the spatial average over land and ocean, respectively.

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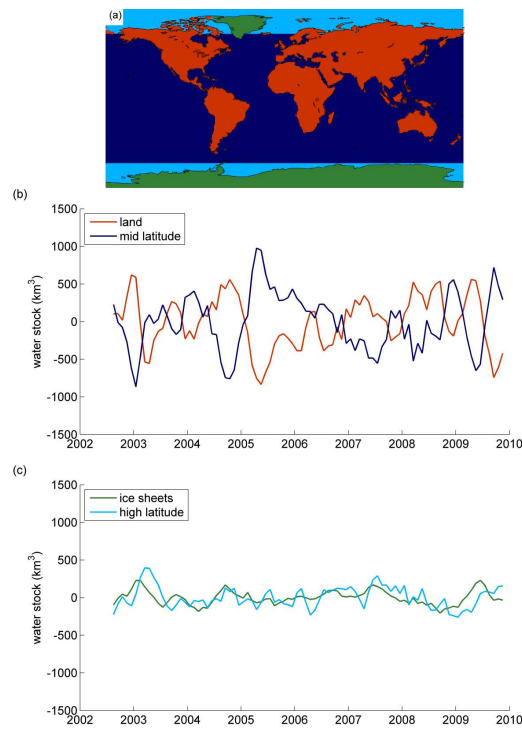


Fig. 2. (a) Land and ocean masks (high latitude Arctic and Antarctic oceans in light blue and ice sheets in green). (b) Detrended interannual variations of GRACE derived water stocks in land (except ice sheets) and oceans (except high latitudes). (c) Detrended interannual variations of GRACE derived water stocks in ice sheets and high latitude oceans.

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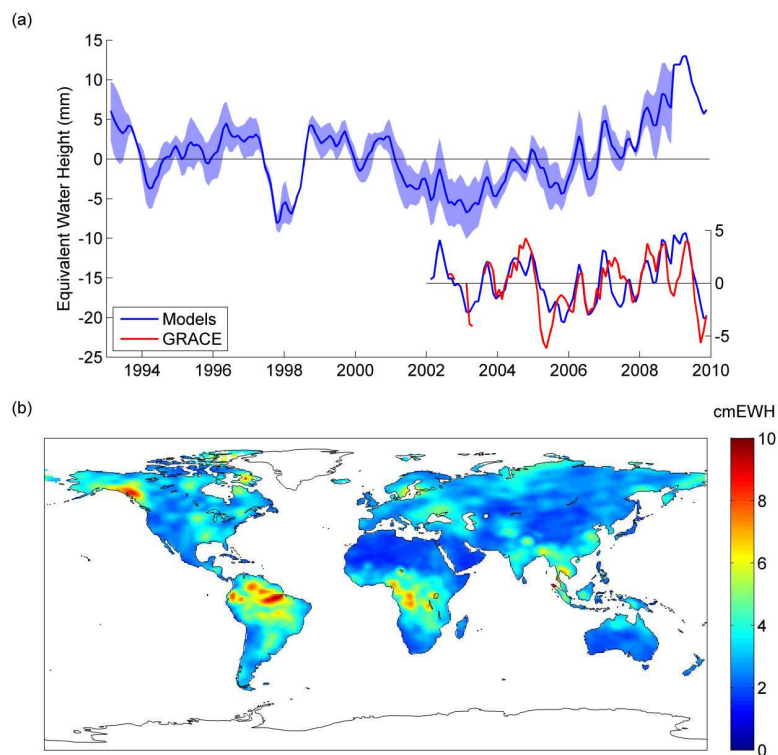


Fig. 3. (a) Terrestrial water storage interannual variations from models and GRACE. Blue shading represents discrepancies between models. (b) Mean RMS between GRACE and LSMs terrestrial water storage.

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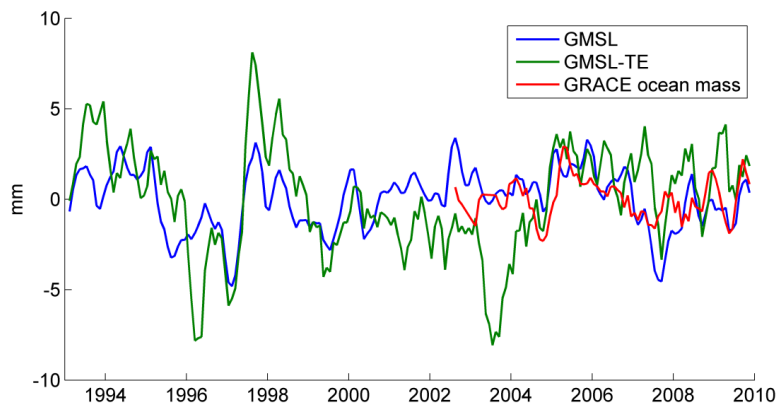


Fig. 4. Comparison of global mean sea level (GMSL) computed from altimetry, GMSL corrected from the thermal expansion TE and GRACE ocean mass.

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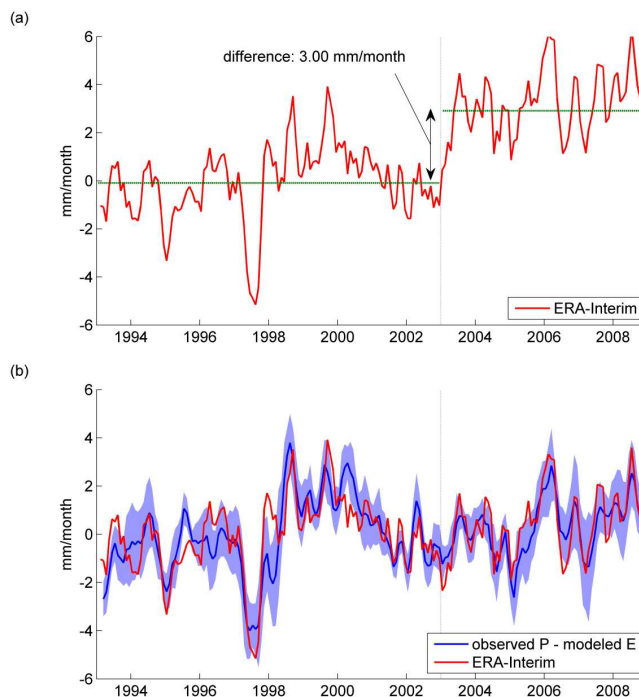


Fig. 5. (a) Interannual variability of precipitation minus evapotranspiration over land ($P_1 - E_1$) computed from the atmospheric water budget and ERA-Interim data. (b) Corrected $P_1 - E_1$ with a uniform offset of $-2.78 \text{ mm month}^{-1}$ over 2003–2009, and observed P_1 minus modeled E_1 used to find the optimum offset (blue shading represents discrepancies between $P_1 - E_1$ obtained from the different the data sets).

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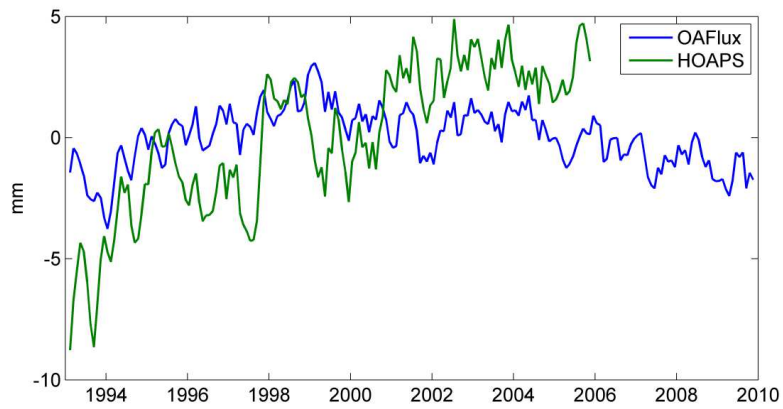


Fig. 6. Interannual variability of evaporation over ocean from OAFIux and HOAPS.

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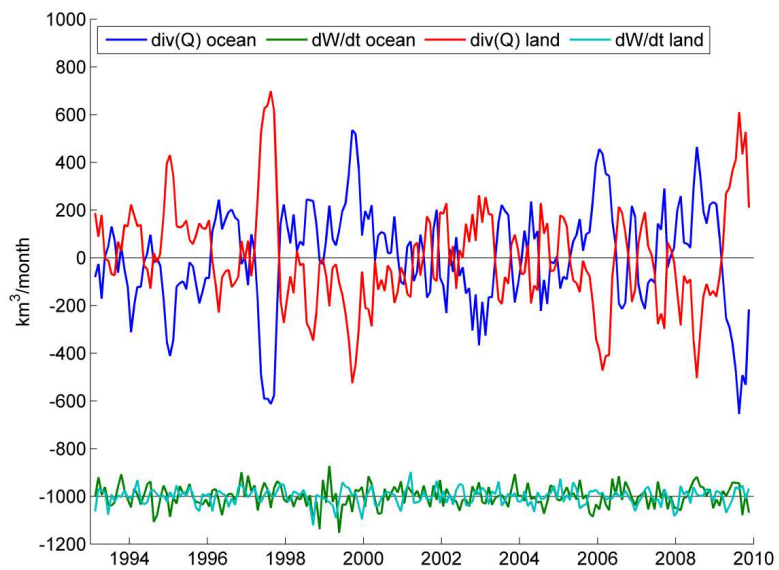


Fig. 7. Components of the atmospheric water budget. Spatial mean over ocean and land. The dW/dt time series are shifted vertically for clarity.

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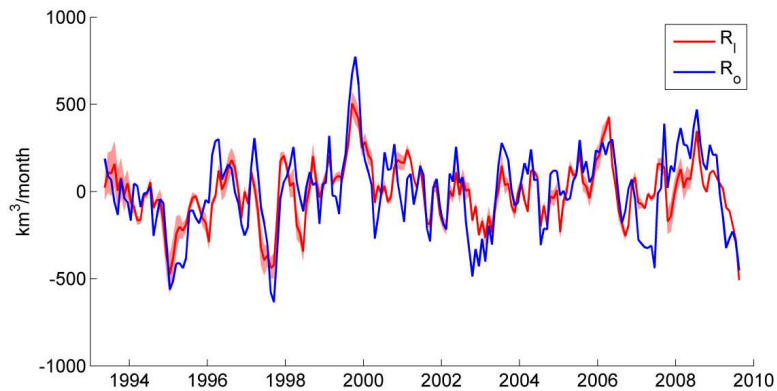


Fig. 8. Comparison of runoff computed from land-atmosphere (R_l) and ocean-atmosphere (R_o) water budgets. The red shading represents discrepancies relative to the different considered LSMs.

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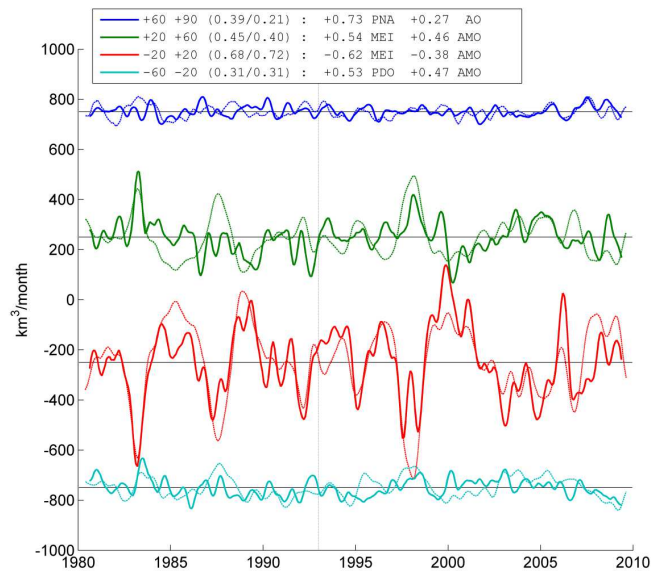


Fig. 9. Comparison of zonal runoff computed from land-atmosphere water budgets (R_l , solid lines) and Combined Runoff Index (CRI, dashed lines) over northern high latitudes (60° N – 90° N), northern mid latitudes (20° N – 60° N), intertropical zone (20° S – 20° N) and southern mid latitudes (60° S – 20° S). For each zone, time series are shifted vertically for clarity. CRI has been normalized to match the range of global runoff variability. The two numbers in the brackets represent the correlation between R_l and CRI for the calibration period (1993–2009) and the whole period, respectively. Results of calibration (climate indices and related coefficients) are given in the legend box.

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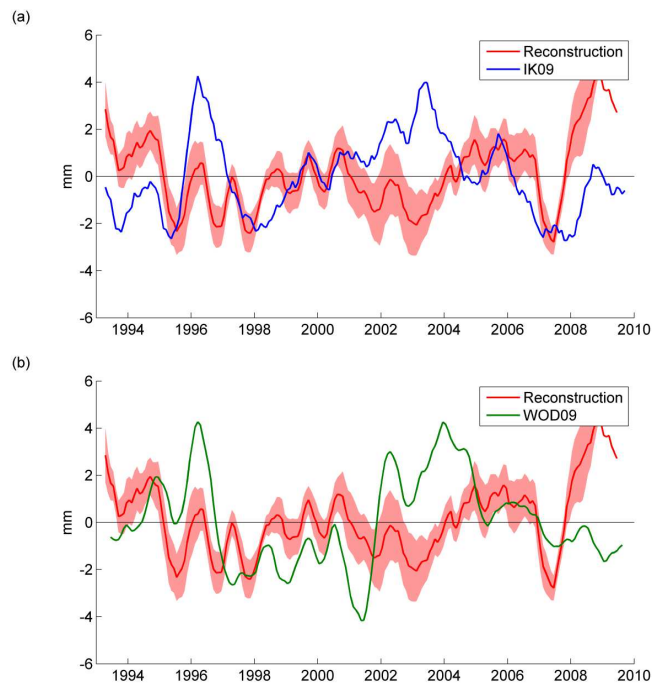


Fig. 10. Comparison of the interannual variations of the thermal expansion obtained from this study and from **(a)** Ishii and Kimoto (2009) v6.12 (IK09) and **(b)** Levitus et al. (2009) (WOD09). Red shading represents discrepancies due to LSMs.