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Contribution of continental water to sea level variations during the 1997–1998 El Niño–Southern Oscillation event: Comparison between Atmospheric Model Intercomparison Project simulations and TOPEX/Poseidon satellite data

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[1] Satellite altimetry from TOPEX/Poseidon (T/P) is used to estimate the variation of the global sea level. This signal, once corrected for steric effects, reflects water mass exchange with the atmosphere and land reservoirs (mainly ice caps, soils and snowpack). It can thus be used to test the capacity of general circulations models (GCMs) to estimate change in land water storage. In this study, we compare the land hydrology contribution to global mean sea level variations during the major 1997–1998 El Niño–Southern Oscillation event from two data sets: (1) the results of the Organizing Carbon and Hydrology In Dynamic Ecosystems (ORCHIDEE) land surface scheme, developed at the Institute Pierre Simon Laplace, coupled to the Laboratoire de Météorologie Dynamique Atmospheric General Circulation Model (LMD AGCM) and (2) the T/P-based estimates. We show that the seasonal variation of the continental water storage is well represented in the model. The drastic amplitude change between the two contrasted years, 1997 and 1998, observed from satellite altimetry, is also simulated. We analyze the role of each component of simulated water fluxes (precipitation, evaporation, and runoff) in determining the range of annual continental water mass variation and its interannual variability. The difference between the two years, 1997 and 1998, is, for an essential part, due to land precipitation in the 20°N–20°S domain. This analysis emphasizes the important role of tropical regions in interannual variability of climate.

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1. Introduction

[2] A number of previous studies have shown that, at the annual frequency, the global mean sea level, as measured by TOPEX/Poseidon (T/P) altimetry, results from two main contributions: thermal expansion of the oceans and water mass exchanged with other surface reservoirs (atmosphere, land water reservoirs, and ice caps) [*Chen et al.*, 1998; *Minster et al.*, 1999; *Cazenave et al.*, 2000; *Chen et al.*, 2002]. These studies showed in particular that while the observed annual mean sea level has an amplitude of about 4 mm, correcting for thermal expansion gives a residual signal of about 8–9 mm, i.e., twice the observed signal. This is so because thermal expansion has an amplitude of also about 4 mm, in terms of global mean, but is almost out of phase compared to the observed annual sea level. The residual signal represents a volume of about 3000 km^3 added/removed seasonally to the oceans. A small contribution comes from atmospheric water vapor (2 mm equivalent sea level or 670 km^3).

[3] Most of the remaining results from seasonal change in soil water, underground water and snowpack. Comparison with outputs of global hydrological models showed that snow is by far the largest contribution [*Cazenave et al.*, 2000; *Chen et al.*, 2002].

[4] So far, similar approaches applied to the interannual/decadal signal have been hampered by several problems: until recently lack of long-term ocean temperature and land water balance time series, uncertainty of contributions from mountain glaciers and ice cap melting. Recently, *Milly et al.* [2003] used the Land Dynamics (LaD) global hydrological model to estimate for the period 1981–1998, the sea level changes associated with climate-driven changes in storage of water as snowpack, soil water and ground water. They found a small positive sea level trend, of 0.12 mm yr^{-1} over

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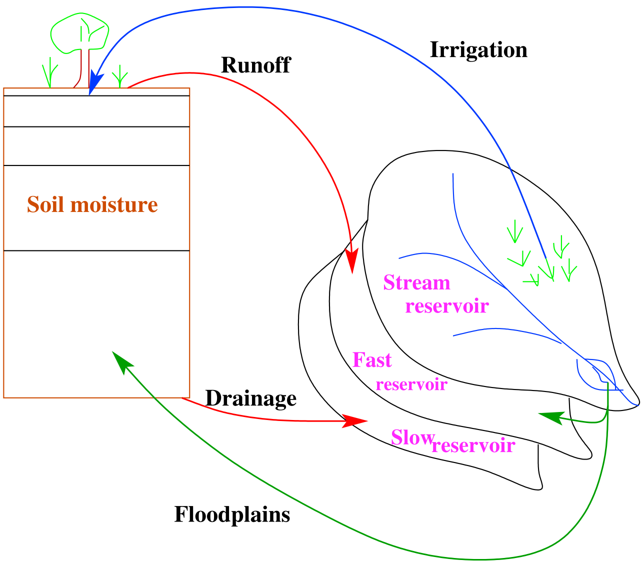


Figure 1. Principle of the river routing scheme.

this 18-year period, corresponding to a downtrend in continental water storage. However, substantial interannual anomalies in land water storage, hence sea level, were reported, mostly driven by interannual variations in precipitations in humid equatorial and mid to high-latitude regions.

[5] Over the period of T/P observations (from 1992 to present), the two years 1997 and 1998 represent a very strong El Niño–Southern Oscillation (ENSO) event. In this study, we focus on the interannual variability of continental water between these two contrasted years.

[6] We infer the continental water budget, using the T/P altimetry data and the recently released historical data of ocean subsurface temperature from *Ishii et al.* [2003]. The drastic change between the two years, 1997 and 1998, is compared to the result obtained by a Laboratoire de Météorologie Dynamique Atmospheric General Circulation Model (LMD AGCM) numerical experiment which simulates the interannual variation of climate, using prescribed sea surface temperature (SST).

2. Description of the Numerical Experiment

[7] Atmospheric Model Intercomparison Project (AMIP) is a standard experimental protocol for global AGCMs. It provides a community-based infrastructure in support of climate model diagnosis, validation, intercomparison, documentation and data access. This framework enables a diverse community of scientists to analyze AGCMs in a systematic fashion, a process which serves to facilitate model improvement.

[8] The AMIP experiment itself is simple by design; an AGCM is constrained by realistic SST and sea ice cover from 1979 to near present, with a comprehensive set of fields saved for diagnostic research.

[9] In this study the AMIP simulation performed with the LMD GCM is used for estimating the time-varying storage of continental water. The version of the LMD GCM used here has a regular grid with a resolution of 96 points in longitude, 72 points in latitude, and 19 levels. A general description of this version of the LMD GCM is available from *Hourdin et al.* [2002] and *Li* [1999]; nevertheless it is useful to recall some of the model aspects which are relevant for the present study. The lower boundary conditions used for this simulation are the SST and sea ice cover provided for the AMIP experiments over the 1979–1999 period.

[10] The grid of the LMD GCM has for each mesh a fractional cover for land, glaciers, sea ice and ocean. This ensures that the area of the four surfaces is very close to reality. The water balance is exact for the atmosphere and all surfaces except the ocean which is an infinite source of moisture. This limit on the simulated water cycle is not relevant for this study as the water mass changes over the ocean will be diagnosed as the residual from the 4 other systems.

[11] As the main component which acts on the water mass changes in the ocean is the continental surface, a few comments on the LSM are in order here. Land surface processes are simulated by the Organizing Carbon and Hydrology In Dynamic Ecosystems (ORCHIDEE) LSM. This model represents the water and energy cycle, the carbon cycle as well as the ecological processes. However, for this study we have only used the water and energy cycle component which is derived from Schématisation des Echanges Hydriques l'Interface entre la Biosphère et l'Atmosphère (SECHIBA) [*de Rosnay and Polcher*, 1998; *Ducoudré et al.*, 1993]. This model has a two meter deep soil moisture reservoir which is split into two levels. The soil moisture can evaporate through the vegetation or directly from the bare soil. The water which leaves the soil moisture reservoir either through runoff or drainage is stored in the three reservoirs of the routing scheme and cascades progressively toward the oceans (see Figure 1). The routing scheme is based on a simple linear cascade of reservoirs as used, for instance, by *Hagemann and Dümenil* [1998]. Each reservoir is characterized by a different time constant with the fastest being considered as the stream flow. A more detailed description of this configuration of ORCHIDEE is available from *Verant et al.* [2004]. Generally, runoff routing schemes compute the river flows in an off-line mode, using output of runoff simulated by GCMs or land surface schemes [*Ducharne et al.*, 2003]. In this experiment, the river routing scheme was included in the LMD GCM.

[12] In this study, AMIP simulation is used to analyze the terrestrial water variations. Before doing such analysis, it is thus important to show that the model simulates continental precipitation rather realistically. Figure 2 shows the averaged 1979–1999 land precipitation for the simulation and the Climate Research Unit (CRU) data [*New et al.*, 2000], which is a high-resolution (0.5°) gauge-only product for the 1901–2000 period. Figure 2 shows that the intertropical convergence zone is well simulated. Over South America, the precipitation rate exceeds 4 mm d^{-1} over a large domain, covering Colombia, Peru, Brazil. However, these high values do not extend as much as in the CRU data, toward Venezuela and Paraguay. Over Africa, the rain belt appears over tropical forest but, also, does not extend enough toward the north. The strong maximum over Indonesia and Malaysia are well simulated. Over southern

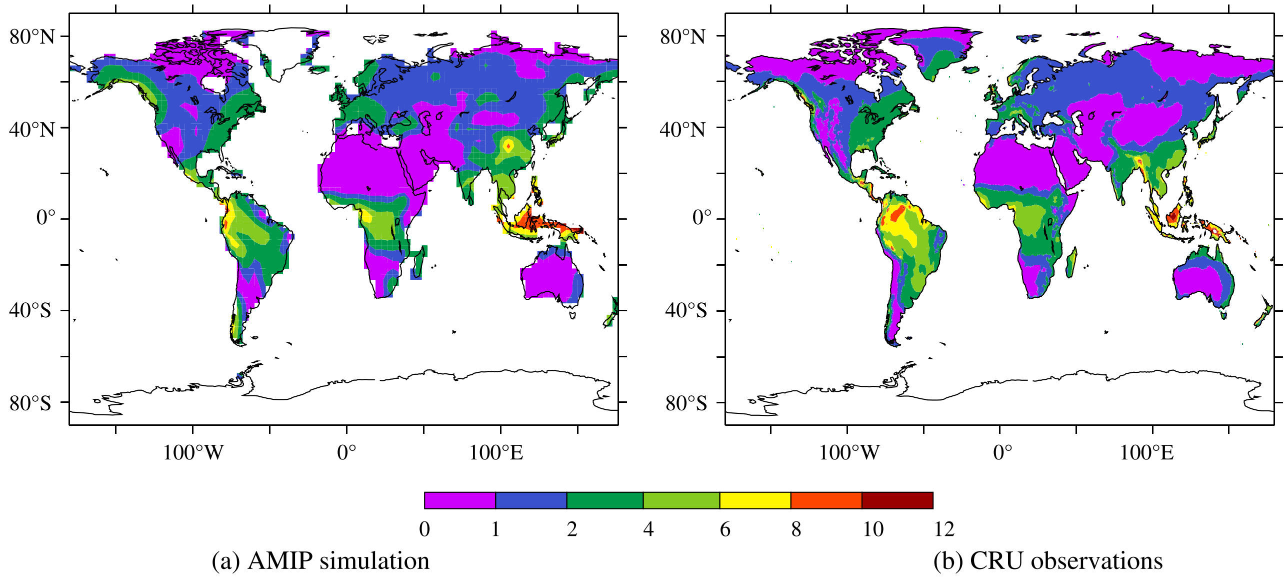


Figure 2. Global mean value of land precipitation from 1979 to 1999. (a) AMIP simulation. (b) CRU observation. Units in mm d^{-1}.

middle latitudes, one may note a non realistic maximum over Patagonia and too dry signal over northeast Australia. The high precipitation rate over the two coasts of North America appears in the simulation but does not extend sufficiently, for example over the Mississippi basin.

[13] Although this comparison shows some biases of the simulated precipitation, the overall pattern seems sufficiently realistic for the following study.

3. Water Mass Change Inside the Oceans

[14] T/P altimetry data for 1993–2002 are analyzed to estimate the seasonal variation of the global mean sea level. Data processing accounts for the most recently updated Geophysical Data Records (Aviso, 2003). The inverted barometer correction associated with the instantaneous local response of sea level to atmospheric pressure variations has been applied as explained by *Minster et al.* [1999]. To estimate the seasonal signal over 1997 and 1998, the sea level time series have been detrended. The positive trend, amounting to 3 mm yr^{-1} over the 1993–2003, mostly results from warming of the world ocean plus mountain glacier and ice sheets melting [*Cazenave and Nerem*, 2004].

[15] To correct for thermal expansion (also called steric effect), we used historical data of ocean subsurface temperature made recently available by *Ishii et al.* [2003]. This data set consists of monthly $1^\circ \times 1^\circ$ gridded ocean temperatures and associated uncertainties, down to 500 m for 1950–1998. It has been derived from objective analysis methods applied to the raw temperature data (see *Ishii et al.* [2003] for a detailed description of the data).

[16] To estimate the steric sea level for a given month, we computed density change with respect to a reference density

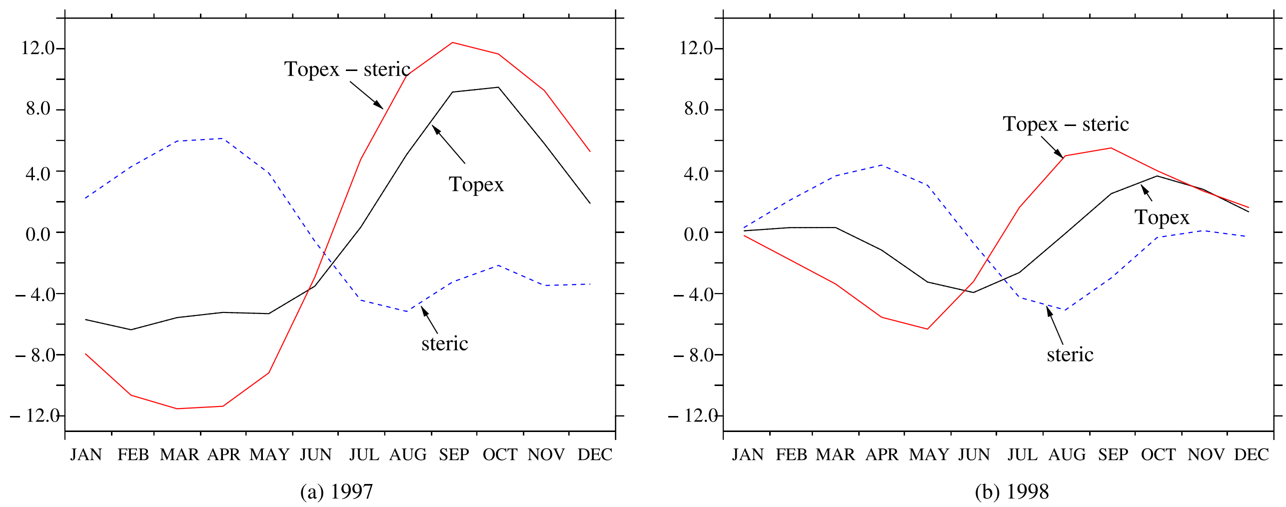


Figure 3. T/P-derived sea level (black curve). Steric sea level estimated from *Ishii et al.* [2003] data (blue curve). Residual signal (T/P sea level minus steric effect) (red curve). Units in mm.

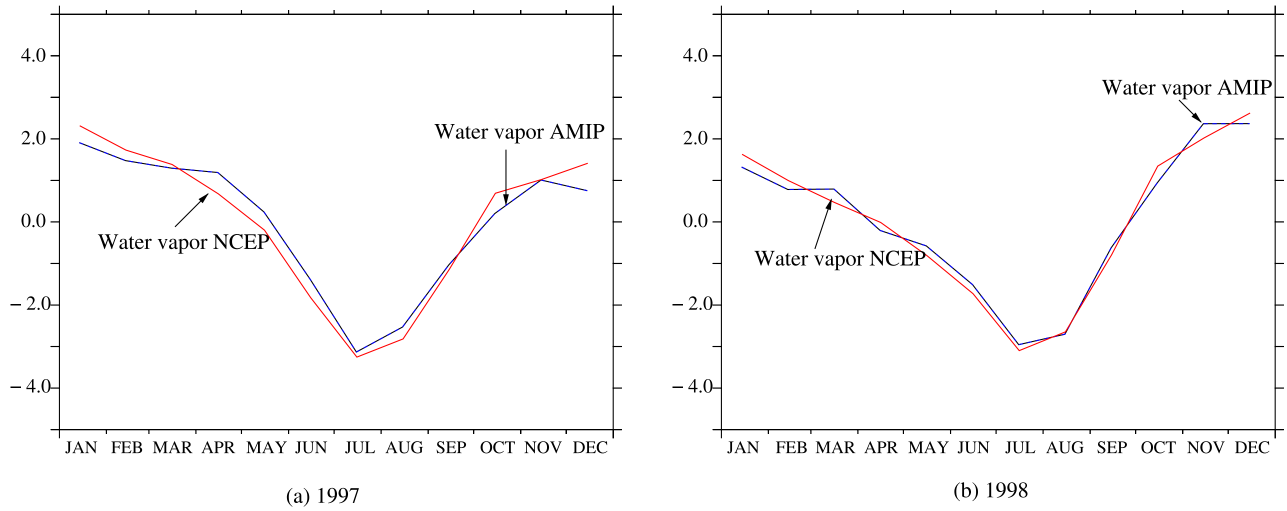


Figure 4. Water vapor contribution expressed in terms of equivalent sea level from AMIP (blue curves) and NCEP/NCAR (red curves). Units in mm.

at any level and grid point according to the classical expression in which the density is obtained in a sequence of steps [*Gill*, 1982].

[17] Figures 3a and 3b show detrended T/P sea level, steric correction and residual signal (T/P sea level minus steric effect) for 1997 and 1998, respectively.

[18] Comparing Figures 3a and 3b indicates that the annual mean sea level is significantly different over the two years, with smaller amplitude in 1998 compared to 1997. On the other hand, the steric component, although not being exactly similar, shows less amplitude variation. As a consequence the residual signal exhibits strong difference between the two years.

[19] For 1997, the residual signal has an amplitude of 12 mm, with minima occurring in April and maxima occurring in September. For 1998, it has an amplitude of 6 mm, with the minima in May and the maxima in September. The variability between the two years seems sufficiently large and significant to attempt to reproduce it with a climate model.

[20] It is worth mentioning that a recent study by *Willis et al.* [2004] estimates the steric sea level change over 1993–2003 using in situ hydrographic data of various sources. Subtracting the steric sea level curve from the T/P-derived curve, they note that the residual displays a large peak during the year 1997, while in 1998, the residual curve shows rather a minimum. As discussed by *Willis et al.*, such a behavior, which is in good agreement with our own results based on a different temperature data set, comes from water mass exchange with the continents.

4. Contribution of Water Vapor in the Atmosphere to Sea Level Variation

[21] As discussed in previous studies, the residual signals shown in Figures 3a and 3b represent water mass changes

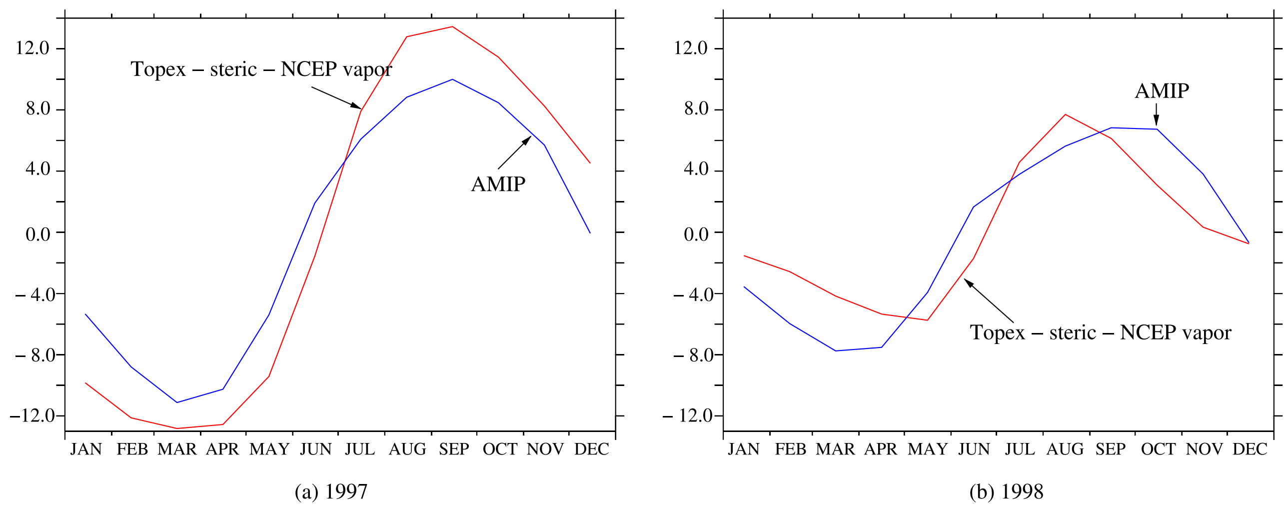


Figure 5. Contribution of continental water (snow and soil moisture) to sea level variations simulated by the AMIP simulations (blue curves) and given from the observations (red curves). Units in mm.

Table 1. Precipitation, Evaporation and Runoff in the AMIP Simulations, Global Mean Value From April to September of 1997 and 1998[a]

	1997	1998
−P	−1.991	−2.321
E	1.561	1.638
R	0.737	0.928
−P + E + R	0.307	0.245

[a]In mm d^{-1}.

inside the oceans which are related to water mass changes in the atmosphere and the terrestrial reservoirs according to the water mass conservation equation [e.g., *Minster et al.*, 1999]:

$$\Delta M_{\mathrm{oceans}} + \Delta M_{\mathrm{watervapor}} + \Delta M_{\mathrm{continentalwater}} = 0 \qquad (1)$$

where ΔM is the water mass change inside each of the three main reservoirs: oceans, atmosphere, and continents.

[22] To estimate the water vapor contribution, we used water vapor distribution simulated by the AMIP experiment. Corresponding contribution to global mean sea level is shown in Figures 4a and 4b for 1997 and 1998, respectively. In Figure 4, estimates based on the 50-year National Center for Environmental Prediction/National Centers for Atmospheric Research (NCEP/NCAR) reanalysis [*Kistler et al.*, 2001] are superimposed.

[23] There is a clear agreement between the AMIP simulation and the NCEP/NCAR reanalysis, which gives us some confidence in the model results. For 1997, the water vapor contribution has its maximum (about 2 mm) in mid-January and its minimum (−3.1 mm) in mid-July. For 1998, the maximum, occurring in mid-December is about 2.5 mm. The minimum, occurring in mid-July is the same for NCEP/NCAR and AMIP results, with an amplitude of −3 mm. The differences between 1997 and 1998 in precipitable water are small compared to the sea level changes.

[24] The water vapor contribution based on the NCEP/NCAR reanalysis has been subtracted from the residual signal shown in Figure 3. We call this new residual the "continental water contribution." Note that variations of continental water storage can also be estimated from moisture flux convergence over land (estimated with reanalysis data) and observed runoff after some corrections have been applied [e.g., *Seneviratne et al.*, 2004], though this approach is difficult to apply in the present case due to the lack of runoff observations.

5. Contribution of Continental Water to Sea Level Variation

5.1. Comparing the GCM Simulation With Observations

[25] The soil moisture and snow contents, which define the water storage over continents are assessed through the simulation of our climate model. In this analysis, we suppose that their changes induce instantaneously the level changes in the oceans. The continental water contribution corresponds to soil moisture and snow changes; it is calculated in equivalent sea level variations via a multiplicative factor *f*:

$$f = -\frac{S_{\mathrm{continents}}}{S_{\mathrm{oceans}}} \qquad (2)$$

where $S_{\mathrm{continents}} = 1.362 \times 10^8$ km^2 and $S_{\mathrm{oceans}} = 3.387 \times 10^8$ km^2 are the surface of the continents and the oceans, respectively, simulated in the model. The continental water contribution to sea level variations is compared to T/P-derived values in Figure 5.

[26] A sharp contrast between the two years is obtained with the LMD GCM. Although the simulated signals may have large differences with the observed ones, the drastic changes between 1997 and 1998 are clearly captured by the model.

[27] For 1997, the continental water contribution rises up to 13.5 mm of amplitude for the observations and 10 mm for AMIP, with a maximum in mid-September and a minimum in March. For 1998, this contribution has a smaller amplitude in both observations and simulations. The maximum reaches only about 8 mm in August for the observations and 7 mm in September for AMIP. The minimum, occurring in March–April for AMIP and in May for the observations is −8 mm and −6 mm, respectively.

[28] The differences between AMIP and T/P-derived signals could be due to the following reasons: (1) data uncertainties (error associated with the measures of T/P, the steric effect, the water vapor contribution); (2) missing signal in the high-latitude oceans because the T/P satellite flies between 66°N and 66°S and the contribution of Greenland and Antarctica were neglected in the model; and (3) uncertainties in soil moisture, snow and horizontal water simulations, due for a large part to errors in precipitation rates simulated by the GCM. In section 5.2,

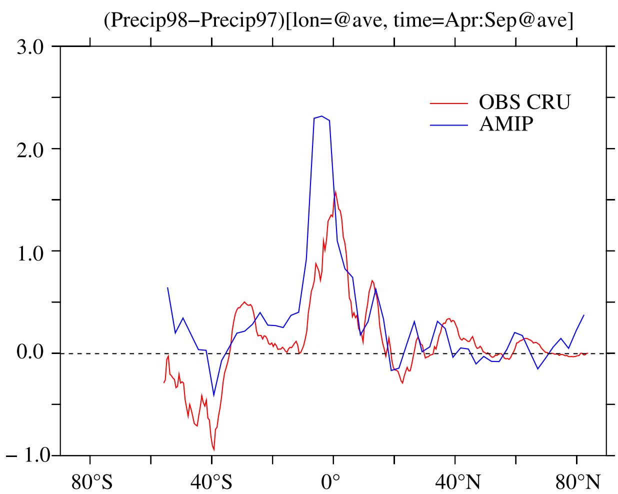
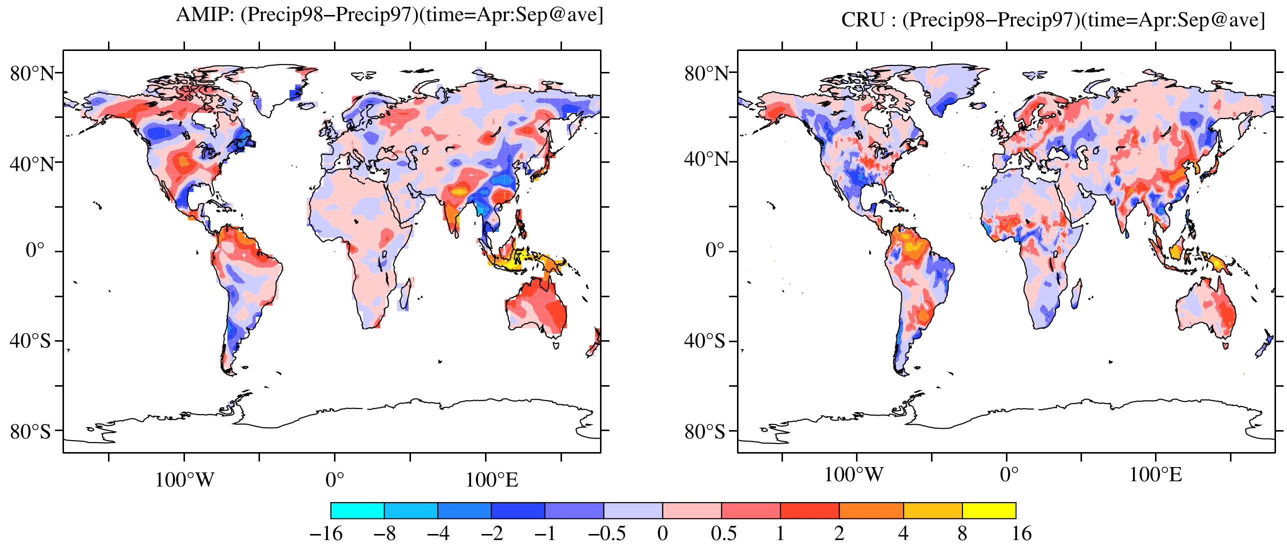


Figure 6. Difference of zonal average value from April to September for land precipitation between 1997 and 1998, simulated by AMIP (blue curve) and observed by CRU (red curve). Unit in mm d^{-1}.



(a) AMIP simulation (b) CRU observations

Figure 7. Difference of global mean value from April to September for precipitation between 1997 and 1998. (a) AMIP simulation. (b) CRU observations. Units in mm d^{-1}.

we analyze the causes of the contrast that occurs between 1997 and 1998.

5.2. Processes That Explain the Interannual Variability of Sea Level

[29] In the LMD GCM, continental water variations are controlled by

$$\frac{\partial W}{\partial t} = P - E - R \qquad (3)$$

where W is continental water (sum of snow and soil moisture), P is precipitation, E is evaporation, and R is runoff over the continents.

[30] From the AMIP simulations, the averaged slopes of the curves of Figure 5 between April and September are computed by adding the contribution of the three components on the right-hand side of equation (3) over these 6 months:

$$s = \frac{f}{6} \sum_{\mathrm{April}}^{\mathrm{September}} (P - E - R) \qquad (4)$$

where s is the averaged slope and f is the ratio defined in equation (2).

[31] From April to September, the continental water contribution to sea level increases more rapidly for 1997 than for 1998. The 1997 slope is thus greater than the 1998 slope during this time interval (Figure 5).

[32] Table 1 shows the contribution of each term of the continental water storage, averaged over these 6 months. During this period, the sum of the evaporation and runoff rates are larger than the precipitation rate and the soil is drying. The drying is slower in 1998 than in 1997. This is due to a rather large precipitation increase in 1998 (0.33 mm d^{-1}). Although a sizable runoff increase (60% of the precipitation increase) and a smaller one in evaporation occur, the last two contributions do not totally compensate the precipitation increase and the net effect is that the soil is wetter in 1998 compared to 1997. Thus the sea level increases less in 1998 than in 1997.

[33] Two questions arise: first to evaluate if this precipitation variation between the two years is realistically simulated; then whether the hydrological balance provides any insight into the response of the other components to this variation. To assess the precipitation difference between the two years, precipitation observations are needed. It is known that there are uncertainties in precipitation estimates. For example, *Fekete et al.* [2004] have compared five monthly precipitation data sets and found some variations

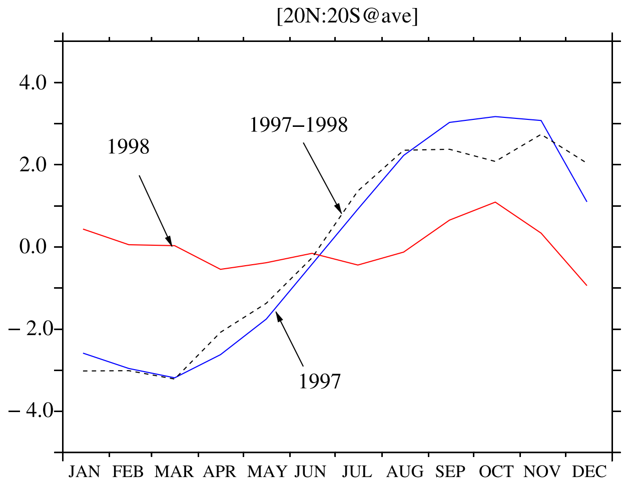


Figure 8. Seasonal variations of tropical continental water mass (average for the 20°S, 20°N latitudinal band) simulated by AMIP for 1997 (blue curve), for 1998 (red curve) and the difference between 1997 and 1998 (black dashed curve), expressed in terms of equivalent sea level. Units in mm.

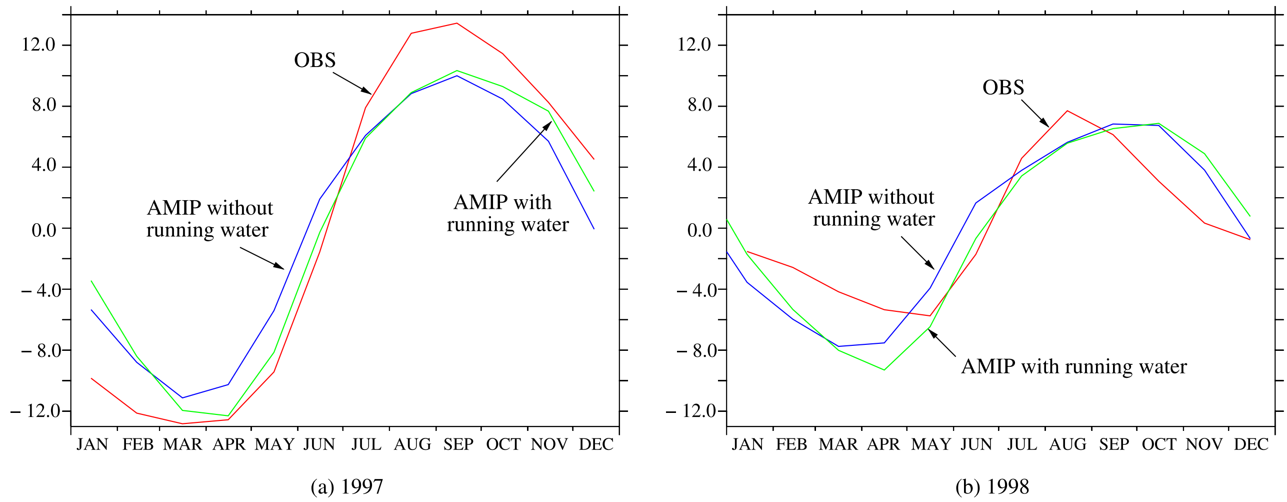


Figure 9. Contribution of continental water to sea level variations. Blue and green curves are AMIP without and with running water, and red curve is observation. Units in mm.

between the individual data sets. However, they showed that the overall precipitation pattern was fairly similar in each data set and the mean annual values were within 10% range. For this study, we choose CRU data set which, with the resolution of 0.5° × 0.5°, can be used for an average over land more accurately than the others.

[34] The zonal mean of the precipitation difference between 1998 and 1997, averaged from April to September for the AMIP simulation and CRU data, is represented in Figure 6. The observed results show a large maximum over equatorial latitudes that extends toward 20° in the northern hemisphere and two other local maximum at 30°S and 40°N. The negative values at 40°S and southward do not give contribution to the global signal because the land surface over these latitudes is really small, covers only 1.34% of global continental surface (according to the GCM simulation). Compared to the observations, the AMIP simulations roughly capture the observed differences in precipitation between 1998 and 1997 and correctly locate the largest difference in the 20°S:20°N latitudinal band, but with an overestimation in the southern part. The two other maximum, at 30°S and 40°N are not well simulated and can be found equatorward to the observations. At high latitudes, the model is not realistic. Figure 7 shows that the overestimation in the 10°S:0°N latitudinal band comes principally from the overestimation of precipitation differences between 1998 and 1997 over the east of Brazil, the south of Africa and the north of Australia. When looking at geographical details, Figure 7 also suggests that the interannual variability might not be very well captured by the AMIP simulations, except the north of South America, some parts of Central Africa, India, Malaysia and Indonesia. The estimate of global mean (from April to September) land precipitation variation between the two years from CRU data set is 0.2 mm d^{-1}, which is less than 0.3 mm d^{-1} of model estimate. This difference is mostly due to the overestimation in the southern part.

[35] Using the AMIP simulation, we determine if the precipitation changes, taken over the tropical continents only, are responsible of the sea level changes. The evolution of soil moisture, averaged from 20°S to 20°N, is expressed in equivalent sea level change, taking into account the surface of this latitudinal band of continents. The results (Figure 8) show that during 1997, the equivalent sea level height increases from April to September by 6 mm; on the contrary, 1998 is a year when the contribution of tropical continental water is almost zero or slightly negative. Thus the GCM simulation shows that the continental water storage, over tropical areas, is strongly reduced during 1997, when the precipitation anomaly is rather large. We have checked that the contribution of the other latitudes to this contrast was much smaller (less than 1 mm) in the GCM simulation, even over Antarctica and Greenland.

[36] We conclude that a major part of the interannual variability of the continental water storage between 1997 and 1998 comes from the strong variability of precipitation over the tropical continents (between 20°N and 20°S). Our results do not confirm the analysis of *Chen et al.* [2002], who attributed the contrast between the two years, 1997 and 1998, to a change in snow cover at high latitudes; their analysis was conducted with European Centre for Medium-Range Weather Forecasting data. In LMD GCM, the precipitation variation over high latitudes has a relatively small effect.

5.3. Influence of the River Routing Scheme

[37] In the previous discussions, we considered that the variations of water in the continental reservoirs (the drainage and the runoff) would be instantaneously injected to the sea reservoir, i.e., the time for water to run from the continents to the oceans was zero, which is not realistic.

Table 2. Characteristic of the Experiments

Experiments	Start Atmosphere
CT97_1	1 January 1991
CT97_2	1 January 1992
CT97_3	1 January 1993
CT97_4	1 January 1994
CT98_1	1 January 1992
CT98_2	1 January 1993
CT98_3	1 January 1994
CT98_4	1 January 1995

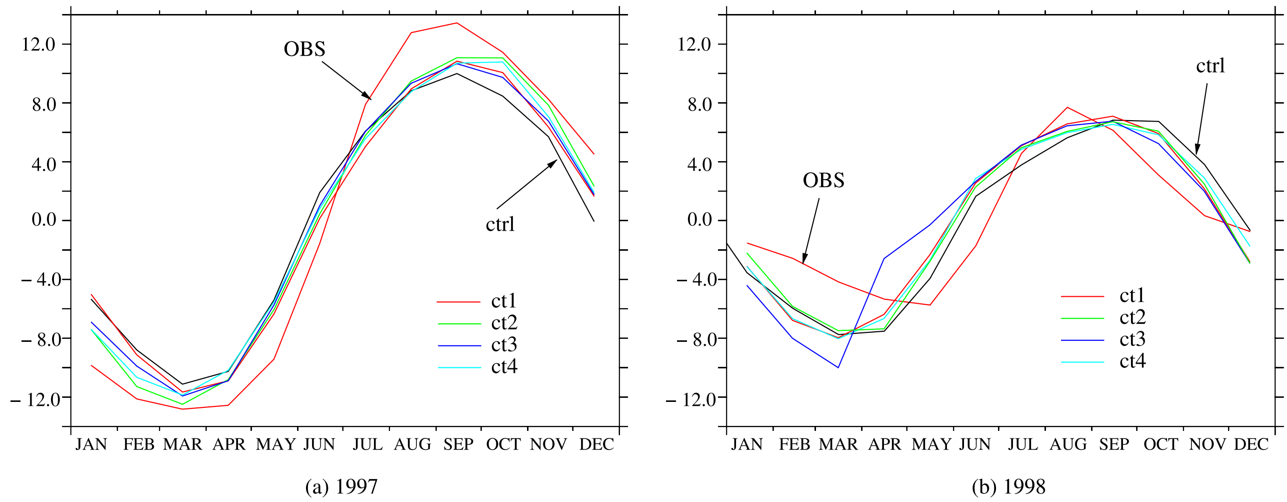


Figure 10. Contribution of continental water to sea level variations, simulations compared to observations. Units in mm for 1997 and 1998.

The ORCHIDEE scheme allows us to explore the time shift which is introduced by the river systems of the world between the soil moisture and snow balance of the continents and the sea level changes.

[38] In Figure 9, we show how the sea level contribution of simulated continental water (Figure 5) is shifted when we take into account the river systems. This contribution (green curves) has a phase shift of about 15 days compared to the curves without running water (blue curves). The amplitudes are slightly increased and the new phase is in good agreement with the observations, especially from April to July in 1997 and May to July in 1998. The simulated curves are very similar to the observations (red curves) during these months when the largest transfer from the continents to the oceans occurs through the river systems. The shift of 15 days is related to the delay due to the flow of water from the area where it exits the soil moisture reservoir to the ocean.

[39] Figure 9 shows that the new development of our land surface scheme has a positive effect in our attempt to reproduce the link between sea level and the continental water balance. It also shows that if one aims to understand the annual cycle of the contribution of the water cycle to the sea level changes, it is essential to take into account the role of the large river systems otherwise a phase shift of about 15 days may be neglected.

5.4. Internal Variability of the Water Cycle Component

[40] The atmosphere has a strong internal variability, thus small differences in the initial conditions can lead to quite different evolutions on timescales of a few weeks to years [*Harzallah and Sadourny*, 1995]. This is especially true for precipitation and the water cycle in general. It is thus legitimate to ask whether the above diagnostics of the continental water cycle are affected by the internal variability of the atmosphere and how representative the results of the GCM are.

[41] In order to study this effect, two ensembles of four simulations were performed for the years 1997 and 1998. The initial conditions used to generate the ensemble were taken from the full AMIP simulation (see Table 2). It is assumed here that these initial conditions are close enough to the state of the model on 1 January 1997 and 1998 not to lead to any significant spin-up periods. Should this hypothesis not be fulfilled it will lead to an overestimation of the intra-ensemble spread.

[42] Figure 10 shows that the internal variability of the water cycle is strongly damped by the large-scale averaging which is performed in this diagnostic. The spread of the water budget contributes to the sea level within the ensemble is smaller than 4 mm. The minimum occurs during the increasing phase of the sea level contribution (April to September). To understand this behavior one must consider the two fluxes which determine the sea level changes induced by the water cycle: the moisture divergence over the oceans (evaporation-precipitation) and the discharge of rivers (Figure 11). They are both affected by very different internal variabilities.

[43] To explain the variations of the internal variability of the sea level changes it is useful to distinguish two periods:

[44] 1. In January to April and September to December, the ocean is loosing water to the continents as the moisture divergence is larger than discharge of rivers. Thus the internal variability is dominated by the oceanic evaporation and precipitation. Both of these fluxes have high internal variability and explain the spread in Figure 10.

[45] 2. In April to September, the river discharge is the dominant hydrological contribution to the sea level changes. As it is larger than the moisture divergence, the water cycle raises the sea level. The internal variability of the river discharge is small as the inertia of the land surface processes contributing water to the rivers dampens the high variability of precipitation and evaporation.

[46] For 1998 the internal variability of the water cycle contribution to the sea level is larger than in 1997. This is explained by the fact that during that year the moisture divergence over the ocean has a weaker annual cycle and the minimum reached in June–July is not as low as in 1997. The contribution of the internal variability of the moisture divergence to the total signal is thus larger.

[47] These analysis of the internal variability strengthen the results presented above. The difference between the

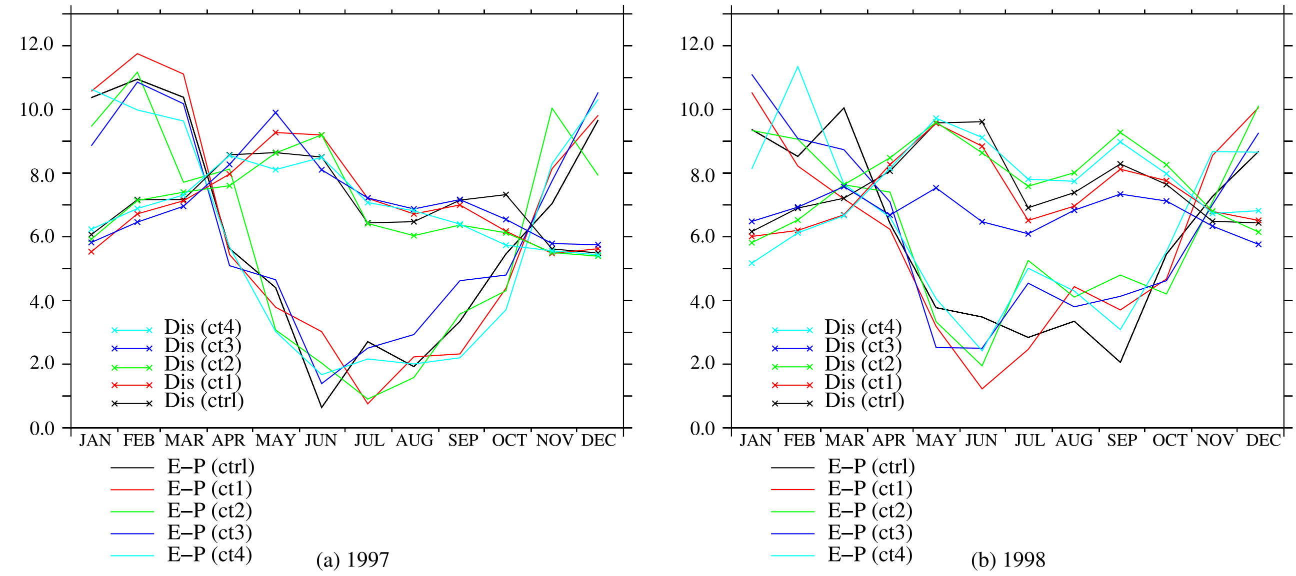


Figure 11. Moisture divergence over the oceans: Evaporation minus Precipitation (E-P, lines without symbols) and river discharge (Dis, line with crosses). Units in mm month^{-1} for 1997 and 1998.

observations and the model diagnostics are significant even at the extremes of the continental water cycle contribution to the sea level changes. The improvements in phasing obtained with the inclusion of the river routing is also significant as it occurs at a time when the internal variability of the system is at a minimum.

6. Conclusions

[48] The measurements of global sea level, by T/P has been used to estimate the interannual variability of continental water storage. The contrast between the seasonal cycles of 1997 and 1998 is especially sharp. Our study aims to understand the mechanisms that explain this contrast.

[49] We have used the integrations of an AGCM to simulate this variation. We have shown that the AMIP simulation, performed with the LMD GCM reproduces accurately the seasonal variation of sea level, and also, the contrast between the two years. The analysis of the model results shows that the interannual variation is caused by an important variation of land tropical precipitation. We have also shown that this precipitation difference reproduces some features of the observed one, reported by CRU, specially when we average the results zonally. The study of the hydrological budget in the AMIP simulation shows that, over tropical latitudes, a substantial drying of soil is occurring during 1997 but soil moisture is hardly changing at these latitudes during 1998. However, one should also realize that the main limitation of this study concerns the uncertainty of the precipitation fields used in the AMIP simulations. Such an uncertainty mostly affects the interannual component of the precipitation, in particular when looking at geographical details. Further simulations using observation-based forcing [e.g., *Ngo-duc et al.*, 2005] could help investigating this issue.

[50] We have also shown that, by introducing a runoff routing scheme, the contribution of the continental water cycle to sea level changes is delayed by 15 days. This result improves the comparison of the model simulation with the T/P-derived signal.

[51] With the GRACE space mission launched in 2002 to measure tiny gravity variations due to water mass redistribution in the surface fluid envelopes of the Earth, it is now possible to directly estimate the month-to-month variations in land water storage, with a resolution of about 400 km at present but likely better in the near future [*Tapley et al.*, 2004; *Wahr et al.*, 2004]. With these observations, we will able to identify more precisely which river basins contribute to the sea level changes and thus better understand the interannual variations of the continental water cycle.

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