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## A new feedback on climate change from the hydrological cycle

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[1] An intensification of the hydrological cycle is a likely consequence of global warming. But changes in the hydrological cycle could affect sea-surface temperature by modifying diffusive ocean heat transports. We investigate this mechanism by studying a coupled general circulation model sensitivity experiment in which the hydrological cycle is artificially amplified. We find that the amplified hydrological cycle depresses sea-surface temperature by enhancing ocean heat uptake in low latitudes. We estimate that a 10% increase in the hydrological cycle will contribute a basin-scale sea-surface temperature decrease of around 0.1°C away from high latitudes, with larger decreases locally. We conclude that an intensified hydrological cycle is likely to contribute a weak negative feedback to anthropogenic climate change. **Citation:** Williams, P. D., E. Guilyardi, R. Sutton, J. Gregory, and G. Madec (2007), A new feedback on climate change from the hydrological cycle, *Geophys. Res. Lett.*, 34, L08706, doi:10.1029/2007GL029275.

### 1. Introduction

[2] The hydrological cycle, i.e. the global circulation of water (in liquid, vapour and ice phases) throughout the earth system, appears to be intensifying. Over land, *Dai et al.* [1997] have reported observations of a linear increasing trend in precipitation between 1900 and 1988. Over ocean, *Wong et al.* [1999] have inferred an increase in high-latitude precipitation between 1930–80 and 1985–94, from measurements of the salinity of intermediate waters. Globally-integrated precipitation increases are predicted by almost all coupled general circulation models (GCMs) in response to increased greenhouse gas forcing [*Cubasch et al.*, 2001]. Averaged over the globe, and over time scales longer than the mean residence time of water vapour in the atmosphere (calculated to be around ten days by *Chahine* [1992]), any increase in precipitation must be balanced by a similar increase in evaporation in a steady state.

[3] To illustrate the modelled intensification of the hydrological cycle under increased greenhouse gas forcing, we show results from a multi-model ensemble of twenty different coupled GCMs, each running the same 1%-per-year CO<sub>2</sub> increase experiment. The integrations, from the IPCC's 4th Assessment Report, are initiated from a pre-

industrial control run in eighteen of the models and a present-day control run in the remaining two. Figure 1 (top) shows the ensemble-mean precipitation minus evaporation (PmE) near the start of the integrations. Figure 1 (bottom) shows the ensemble-mean change in PmE at around the time of CO<sub>2</sub> doubling. The pattern of change is broadly similar to the original pattern: regions of net precipitation (evaporation) become regions of even stronger net precipitation (evaporation). The pattern correlation between Figures 1 (top) and 1 (bottom) is 0.58. Therefore, to a reasonable first approximation, the projected change in the net surface fresh water flux under increased greenhouse gas forcing is a simple amplification, without a change in the spatial pattern. *Held and Soden* [2006] have argued on theoretical grounds that the modification to the hydrological cycle should take this simple form.

[4] Accepting that an intensification of the hydrological cycle is a likely consequence of global warming, the intensification might itself affect ocean heat uptake and sea-surface temperature (SST), giving a feedback on the resulting rate of climate change. This possibility exists because the hydrological cycle maintains salinity structures in the ocean, which influence diffusive ocean heat transports. The influence between SST and PmE is therefore two-way, and precipitation has been characterized as a parameter not only modified by climate change but also itself changing climate [*Cubasch et al.*, 2000].

[5] *Williams et al.* [2006] proposed a specific mechanism by which PmE can affect SST. Changes in the intensities or spatial patterns of PmE over ocean give changes in sea-surface salinity gradients. Ventilation processes transmit these surface changes to the ocean interior, resulting in modified salinity gradients and density-compensating temperature gradients on isopycnal (i.e. constant-density) surfaces. Since isopycnal surfaces are surfaces of greatly enhanced mixing, the modified isopycnal temperature gradients lead to modified diffusive heat transports, which may give SST changes.

[6] In this modelling study, we investigate the role of the conjectured mechanism of *Williams et al.* [2006] under an accelerated hydrological cycle. Section 2 describes the coupled ocean-atmosphere GCM and the experiments performed. The impacts of the accelerated hydrological cycle on ocean temperature are studied in section 3, and are linked to the mechanism of *Williams et al.* [2006] in section 4. We extrapolate our findings to speculate about the response of the climate system to a future intensified hydrological cycle in section 5. Finally, we summarize our findings in section 6.

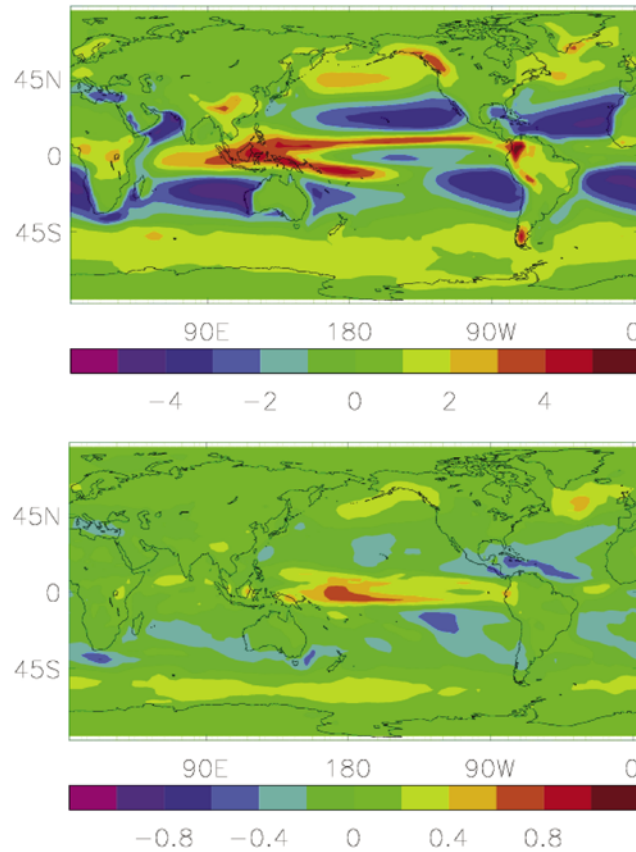
### 2. Experimental Set-Up

[7] We use the SINTEX coupled ocean-atmosphere GCM [*Guilyardi et al.*, 2003; *Gualdi et al.*, 2003]. The ocean

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**Figure 1.** (top) Ensemble-mean precipitation-minus-evaporation averaged over years 1–20 of the multi-model greenhouse gas scenario experiment discussed in the text. (bottom) Ensemble-mean change in precipitation-minus-evaporation between years 1–20 and 61–80 of the same experiment. The value of the latent heat of vaporization at  $10^{\circ}\text{C}$  is used to estimate the evaporation from the latent heat flux for this analysis. Therefore, global mean precipitation is not exactly balanced by global mean evaporation, but the imbalance is small (less than 1% of the global mean precipitation). The units are mm/day.

component is ORCA2, the global configuration of the OPA model [Madec *et al.*, 1998]. The longitudinal resolution is  $2^{\circ}$  and the latitudinal resolution varies from  $0.5^{\circ}$  at the equator to  $2.0^{\circ}$  at the poles. 31 vertical levels are used, with 10 levels in the top 100 m. The atmosphere component is ECHAM4 with a resolution of T106L19. The ocean and atmosphere are coupled using OASIS2.4 [Valcke *et al.*, 2000]. The coupling strategy is described by Guilyardi *et al.* [2003]. Fluxes between the atmosphere and ocean are exchanged once every two hours. There is no interactive sea ice model, since we wish to study the direct impacts of the modified hydrological cycle rather than any indirect impacts caused by modified high-latitude processes. Therefore, sea ice is relaxed towards an observed monthly climatology using a heat flux damping term. Apart from this, no flux correction is applied. The climatological PmE in SINTEX (not shown) agrees very well with the multi-model ensemble-mean PmE of Figure 1 (top), demonstrating the model's basic ability to simulate the main features of the hydrological cycle.

[8] We wish to design an idealised experiment that will isolate the impacts on SST of a global warming-induced

hydrological cycle intensification. In the constant- $\text{CO}_2$  simulation hereafter referred to as EPR2, PmE from the atmosphere (including continental run-off) is doubled before being passed to the ocean. This approach is justified by recalling that the unperturbed PmE (Figure 1, top) and the global warming-induced perturbation (Figure 1, bottom) share broadly similar patterns. Since the hydrological cycle is not actually projected to double in intensity under global warming, this is to be regarded as a sensitivity experiment designed to allow a clear exposition of the relevant physical mechanisms. Note that this experimental set-up deliberately leaves the latent heat flux unmodified and allows us to study the fresh water impacts in isolation. EPR2 is compared to a control integration (hereafter CTRL), and both are initiated from an ocean at rest with the thermohaline structure of Levitus [1982]. The integration length is 50 years, which Williams *et al.* [2006] showed is long enough for equilibrium to be established in the SST (expect perhaps at very high latitudes).

### 3. Impacts of the Amplified Hydrological Cycle on Temperature

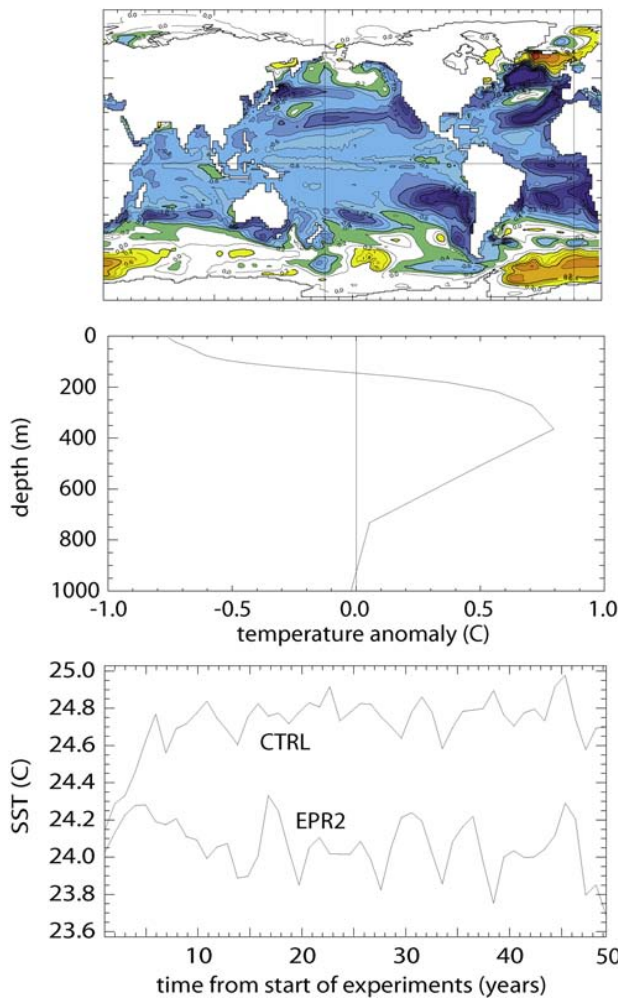
[9] Figure 2 (top) shows the equilibrated SST anomaly in EPR2. The anomaly is equilibrated in the sense that there is no drift in either the spatial pattern or its amplitude between the penultimate and final decades of the integrations. There is a cooling of most of the surface of the ocean. The basin-averaged anomalies in the final decade are  $-1.0^{\circ}\text{C}$  in the Atlantic Ocean ( $0-60^{\circ}\text{W}$ ,  $40^{\circ}\text{N}-40^{\circ}\text{S}$ ),  $-0.8^{\circ}\text{C}$  in the Pacific Ocean ( $110^{\circ}\text{E}-70^{\circ}\text{W}$ ,  $40^{\circ}\text{N}-40^{\circ}\text{S}$ ), and  $-0.6^{\circ}\text{C}$  in the Indian Ocean ( $60-100^{\circ}\text{E}$ ,  $0-40^{\circ}\text{S}$ ).

[10] The cooling generally extends down to around 150 m and is accompanied by an anomalous warming of the approximate depth range 150–900 m. The sub-surface warming reaches  $0.8^{\circ}\text{C}$  near 400 m in the Pacific Ocean (Figure 2, middle), and is of similar magnitude in the other basins. The magnitude of the vertical temperature gradient is therefore reduced in EPR2 compared with CTRL, suggesting that an increased hydrological cycle acts to reduce the thermal stratification of the upper ocean and the tightness of the thermocline. In order to determine the time scale on which the anomalous surface cooling develops, Figure 2 (bottom) shows time series of annual-mean SST in the Pacific Ocean in both integrations. There is a small drift in CTRL from the observed initial state. The SST anomaly in EPR2 is established within 10 years. Inter-annual variability in SST is noticeably greater in EPR2 than CTRL.

### 4. Mechanism Responsible for the Temperature Changes

[11] The impacts seen in section 3 are consistent with the mechanism proposed by Williams *et al.* [2006]. Over ocean, the amplified PmE produces increased sea-surface salinity gradients, which are transmitted to the ocean interior by ventilation processes. The equilibrated isopycnal salinity gradients (Figures 3, top and middle) and density-compensating temperature gradients (not shown) are therefore strengthened in EPR2 compared to CTRL.

[12] Osborn [1998] estimated the vertical component of the isopycnal diffusive heat flux from observations of the



**Figure 2.** (top) Map of sea-surface temperature anomaly in EPR2 relative to CTRL, averaged over years 41–50. The contour interval is 0.2°C. (middle) Vertical profile of temperature anomaly in EPR2 relative to CTRL, averaged in the Pacific Ocean (110°E–70°W and 40°N–40°S) over years 41–50. (bottom) Time series of annual-mean sea-surface temperature, averaged in the Pacific Ocean (110°E–70°W and 40°N–40°S) in CTRL and EPR2.

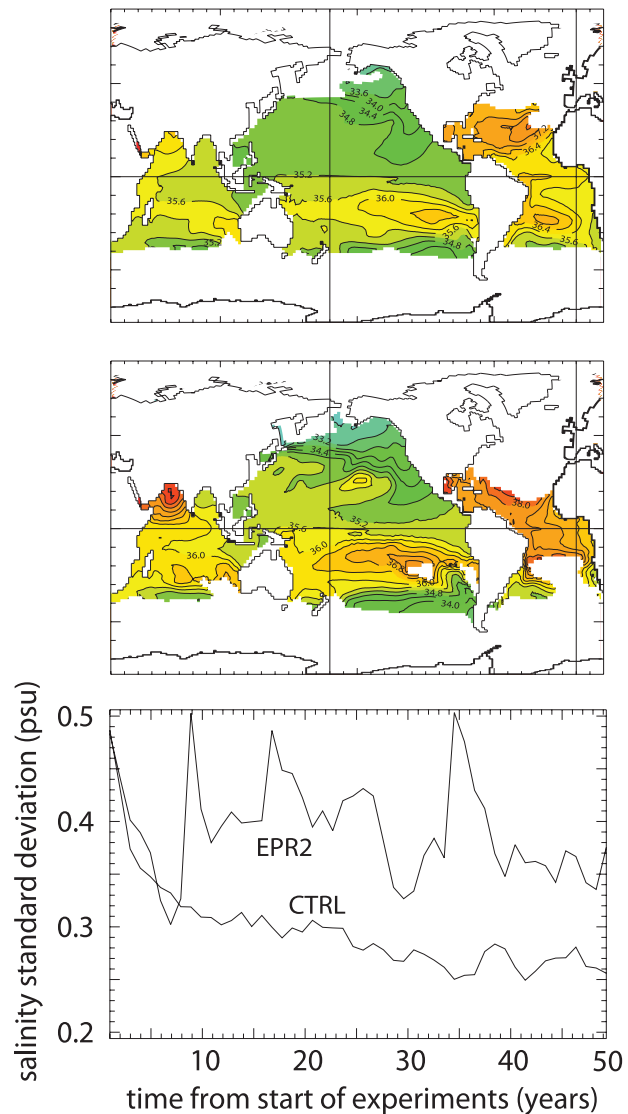
time-mean thermohaline state of the ocean. In the global zonal mean, between 40°N and 40°S and from the surface down to 2000 m, he reports it to be directed downwards and to be of magnitude around  $1 \text{ W m}^{-2}$ . This sign and magnitude were reproduced in the present CTRL integration by Williams et al. [2006]. The strengthened isopycnal temperature gradients in EPR2 therefore increase the downward isopycnal diffusive heat flux compared to CTRL. An anomalous net downward heat transfer results, producing the surface cooling and sub-surface warming reported in section 3. A different mechanism appears to dominate the SST response at very high latitudes, especially in the Southern Ocean (Figure 2, top).

[13] The time evolution of a measure of the strength of the isopycnal salinity gradients in CTRL and EPR2 is shown in Figure 3 (bottom). Following a common drift away from the observed initial value, strengthened gradients are established in EPR2 within 10 years. This is the time required for ventilation processes to establish new isopycnal

salinity and temperature gradients (and hence diffusive heat fluxes) in response to the imposed PmE anomaly [Williams et al., 2006], and it matches the time scale on which SST anomalies are established in EPR2 (section 3). Finally, the temporal variability of the isopycnal salinity gradients is much greater in EPR2 than in CTRL (Figure 3, bottom) because the variability of PmE is doubled. This gives an increased isopycnal diffusive heat flux variability, and is consistent with the higher SST variability in EPR2 seen in Figure 2 (bottom).

### 5. Application to Future Global Warming

[14] We may interpolate our results to estimate the SST response to projected future hydrological cycle changes.



**Figure 3.** Annual-mean at year 50 of salinity on the constant potential density surface defined by  $\sigma_0 = 26.0$ , in CTRL (top) and EPR2 (middle). The units are psu. White shading in the ocean indicates outcropping. (bottom) Time series of the spatial standard deviation of annual-mean salinity in CTRL and EPR2, on the global constant potential density surface defined by  $\sigma_0 = 24.0$ .



Simple one-dimensional radiative-convective models predict hydrological cycle sensitivities to global warming of up to 4%/°C [e.g., *Pierrehumbert, 2002*]. GCMs suggest smaller sensitivities of around 2%/°C [*Held and Soden, 2006*]. At the time of CO<sub>2</sub> doubling, increases in globally-integrated precipitation in the 1%-per-year CO<sub>2</sub>-increase CMIP experiments [*Cubasch et al., 2001*] are in the range 0–7%.

[15] To proceed with the interpolation we must assume that the SST anomaly due to our mechanism scales linearly with the hydrological cycle amplification. As justification, we note that *Huang et al. [2005]* have studied the response of the Pacific Ocean to idealized fresh water anomalies, and found that the temperature response was almost linear with respect to the magnitude, duration and location of the applied fresh water forcing.

[16] EPR2 uses a hydrological cycle amplification factor of 2 (i.e. a 100% increase) and attains basin-scale SST decreases of the order of 1.0°C away from high latitudes. Linear interpolation of this result predicts basin-scale SST decreases of the order of 0.1°C for a hydrological cycle amplification factor of 1.1 (i.e. a 10% increase). This SST decrease is a small component of the greenhouse-gas induced warming signal. We conclude that an intensification of the hydrological cycle is likely to contribute a weak negative feedback to anthropogenic climate change.

## 6. Summary and Discussion

[17] We have studied the impacts of an accelerated hydrological cycle on ocean temperature. We used the SINTEX coupled general circulation model to run a constant-greenhouse gas sensitivity experiment in which the fresh water flux from the atmosphere is doubled before being passed to the ocean. This idealised approach is motivated by noting that the projected change in precipitation minus evaporation is a simple amplification, without much change in the spatial pattern. We analysed 50-year integrations initiated from observations.

[18] The ocean temperature response is consistent with the modified diffusive heat transport mechanism proposed by *Williams et al. [2006]*. The equilibrated basin-scale sea-surface temperature decreases by up to 1.0°C away from high latitudes, accompanied by a sub-surface warming. The increased hydrological cycle variability is associated with increased sea-surface temperature variability. Our results reveal a new aspect of the inter-relation between precipitation minus evaporation and sea-surface temperature.

[19] Interpolation of our results predicts that a 10% increase in the strength of the hydrological cycle will cause basin-scale sea-surface temperature decreases of around 0.1°C (except at high latitudes). We conclude that an intensified hydrological cycle is likely to contribute a weak negative feedback to anthropogenic climate change. Our proposed feedback might play a role not only in the response to future global warming but also on palaeoclimatic time scales in the past, when hydrological cycle variations have exceeded those projected for this century.

[20] We call our effect a feedback because it is a process whereby surface temperature change could feed back on itself via the hydrological cycle. The feedback process involves ocean heat uptake, whose effect on climate change

is usually regarded as transient, unlike other processes called climate feedbacks, such as those involving water vapour, clouds and surface albedo [*Soden and Held, 2006*]. Ocean heat uptake retards global climate change, but if the forcing is stabilised and the system is allowed to approach a steady state, ocean heat uptake (but not additional ocean heat content) decays to zero. On the global scale and in the long term, ocean heat uptake will then give no feedback on climate change. However, on the decadal time scale most relevant for climate prediction, ocean heat uptake in coupled models is roughly proportional to the magnitude of global climate change [*Gregory, 2000; Raper et al., 2002*]. Our process may therefore be treated as a feedback at least on these time scales, and probably would scale with the magnitude of change in the hydrological cycle in time-dependent climate change under increasing forcing.

[21] Our inferred surface temperature feedback of around –0.1°C is quite weak, at around 2–10% of the projected warming signal under CO<sub>2</sub> doubling [*Cubasch et al., 2001*]. However, the inferred –0.1°C is a basin-scale average, and the surface temperature anomaly map of Figure 2 (top) suggests that the local magnitude of the feedback is larger in places. Furthermore, the hydrological cycle increase at CO<sub>2</sub> doubling might be more than 10% locally, for example under the inter-tropical convergence zone. Finally, we note that the atmosphere is very sensitive to small surface temperature changes in the tropical Pacific Ocean, especially in the Indo-Pacific warm pool. Therefore, although our proposed feedback appears to be small on the basin scale, its local effects are not necessarily negligible.

[22] **Acknowledgments.** We are grateful to Buwen Dong for producing Figure 1. We thank Silvio Gualdi for assistance with setting up the model. Discussions with Didier Swingedouw, Rémi Tailleux, and Chris Brierley were helpful. Calculations were performed on the NEC SX-6 at the Deutsches Klimarechenzentrum (DKRZ). We acknowledge funding from the Rapid Climate Change thematic programme of the UK Natural Environment Research Council (award reference: NER/T/S/2002/00442). With regard to Figure 1, we acknowledge the modelling groups for providing their data for analysis for the AR4 and CFMIP, the Program for Climate Model Diagnosis and Intercomparison (PCMDI) for collecting and archiving the model output, and the JSC/CLIVAR Working Group on Coupled Modelling (WGCM) for organising the model data analysis activity. The multi-model data archive is supported by the Office of Science, US Department of Energy.

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