

Did changing ocean circulation destabilize methane hydrate at the Paleocene/Eocene boundary?

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Abstract. During the Paleocene-Eocene Thermal Maximum (PETM, ~55 Ma), marine and terrestrial carbon isotope values exhibit a negative shift of at least 2.5‰, indicative of massive destabilization of marine methane hydrates, releasing ~1100 gigatonnes of methane carbon. The cause of the hydrate destabilization is unknown but has been speculated to be warming due to a change from high-latitude to low-latitude deepwater formation. Here, we present results from a numerical ocean model indicating that a sudden switch of deepwater formation from southern to northern high latitudes caused mid-depth and deep-ocean warming of 3-5°C. The switch is caused by a slow increase in the intensity of the atmospheric hydrologic cycle, as expected under increasing temperatures and consistent with PETM sedimentary evidence. Deepened subduction prior to the thermohaline circulation switch causes warming of 1-4°C in limited areas at thermocline through upper intermediate depths, which could destabilize methane hydrates gradually and at progressively greater depths. The switch itself occurs abruptly, with up to 5°C warming resulting everywhere in the deep ocean.

1. Introduction

Fifty-five million years ago, the Earth underwent an abrupt climate change known as the Paleocene-Eocene Thermal Maximum (PETM). Against the backdrop of an already warm climate with reduced pole-equator temperature contrasts, intermediate to deep ocean water temperatures and high-latitude surface temperatures increased by 6-8°C over less than 20 kyr [Röhl *et al.*, 2000]. The PETM is characterized by a negative shift of 2.5‰ (PDB) or more in marine carbonate $\delta^{13}\text{C}$ records [Kennett and Stott, 1991; Thomas and Shackleton, 1996] and of >5‰ in terrestrial $\delta^{13}\text{C}$ records [Koch *et al.*, 1992]. A feasible explanation for the negative carbon isotope excursion at the PETM is the introduction of 1120 gigatonnes of isotopically depleted (-60‰) carbon to the ocean-atmosphere system [Dickens *et al.*, 1995, 2000; Matsumoto, 1995]. The only known source for this quantity of depleted carbon today are the vast reserves of natural gas hydrate in oceanic sediments and methane gas trapped beneath these deposits [Kvenvolden, 1998].

The suite of dramatic global changes inferred for the PETM also includes increased aridity in subtropical latitudes and increased high latitude precipitation [Robert and Kennett, 1994; Schmitz *et al.*, 2001]. This pattern is consistent with an increase in the intensity of the global hydrologic cycle, which would be expected with increasing atmospheric CO₂ concentration, a change that might have accompanied late Paleocene volcanism [Eldholm and Thomas, 1993; Bralower *et al.*, 1997].

The thermal forcing of the late Paleocene and early Eocene thermohaline circulation (THC) was reduced compared to today, because the pole-equator temperature contrast was smaller.

Therefore, it has often been speculated that the oceans might have been characterized by deepwater formation at low latitudes as first suggested by *Chamberlin* [1906]. Abrupt bottom warming caused by the onset of, or an abrupt increase in this putative low-latitude deepwater source has been invoked as a feature of the PETM [*Kennett and Stott*, 1991; *Kaiho et al.*, 1996; *Bains et al.*, 1999], but there has been no evidence from data or from model studies that deep water originated from low latitudes, rather than from warm high latitudes [*Crowley*, 1999; *Bice and Marotzke*, 2001].

Nonetheless, changing thermohaline circulation is perhaps the most likely mechanism for abrupt, global bottom water temperature change. *Zachos et al.* [1993] noted that the abrupt and brief PETM event may be integrally related to the gradual late Paleocene-early Eocene global warming trend itself. In other words, the PETM may have occurred because the global climate was gradually warming (with the attendant hydrologic cycle increase believed to accompany global warming) and that some climate system threshold was passed, causing the system to “jump” to a new equilibrium state. Such jumps between different stable equilibria have been noted in coupled models with Quaternary, modern and hypothesized future boundary conditions, both with and without a forcing perturbation [e.g. *Manabe and Stouffer*, 1988, 1993, 1995; *Rahmstorf and Ganopolski*, 1999; *Hall and Stouffer*, 2001]. The precise nature of such a threshold response in the latest Paleocene thermohaline circulation has not previously been identified.

Here, we show new ocean model results suggesting that an abrupt switch in deep convection from high southern to high northern latitudes is caused by a gradual strengthening of the idealized atmospheric hydrologic cycle, a change consistent with a gradually warming climate and PETM sedimentary evidence. Gradual intermediate water warming would have occurred prior to the THC switch due to deepened subduction (downward flow of subtropical surface water into the main thermocline), a process that could have destabilized hydrates locally and progressively deeper in the water column above ~ 1500 m. The switch in THC causes abrupt deep water warming that could be sufficient to destabilize methane hydrate over much of the latest Paleocene ocean.

Bice and Marotzke [2001] describe the subduction response observed in uncoupled ocean model experiments and examine the sensitivity of the solution to the parameterizations of continental runoff and ocean diapycnal mixing for one early Eocene paleogeographic reconstruction. In this study, we first repeat the experiment of *Bice and Marotzke* but with atmospheric surface forcings representative of pre-PETM conditions and with a much smaller maximum increase in the intensity of the hydrologic cycle. We then examine the sensitivity of the model thermohaline circulation response to plausible changes in paleogeography around the time of the Paleocene-Eocene boundary.

2. Water Cycle Perturbation Experiments

The reader is referred to an excellent non-technical description of intensification of the water cycle by *Manabe* [1996], but the fundamental features of the hydrologic cycle response to an increase in the atmospheric concentration of carbon dioxide are these: (1) the global mean evaporative flux (or rate of water transfer between the ocean and atmosphere) increases with increased net downward radiation and surface warming, (2) increased mean evaporation rate is balanced by increased precipitation rate, (3) evaporation rate increases more in the subtropics than at other latitudes, and (4) increased precipitation exceeds increased evaporation at high latitudes, an effect enhanced by increased poleward transport of moisture due to tropospheric

warming [Manabe, 1996]. In general, then, the hydrologic cycle response to increased atmospheric carbon dioxide concentration is represented by increased subtropical evaporation and increased high latitude precipitation [Manabe and Bryan, 1985; Manabe, 1996]. In order to examine the possible ocean response to a gradual increase in the strength of the water cycle, we impose this type of change in the moisture flux forcing to an ocean general circulation model.

The model and forcing technique are the same as those used by *Bice and Marotzke* [2001]. The ocean model is the Geophysical Fluid Dynamics Laboratory's Modular Ocean Model (MOM) version 2.2 [Pacanowski, 1996]. The ocean surface is forced initially with zonally invariant temperature, wind stress and moisture fluxes (evaporation rate minus precipitation rate, E-P) predicted by an atmospheric general circulation model (AGCM), GENESIS v. 2.0 [Thompson and Pollard, 1997] (Figure 1). The model atmospheric composition ($\text{CO}_2 = 645$ ppm; $\text{CH}_4 = 16.5$ ppm) and poleward heat parameterization (4 times the modern control value) were set in order to produce a qualitatively good match to estimates of latest Paleocene high

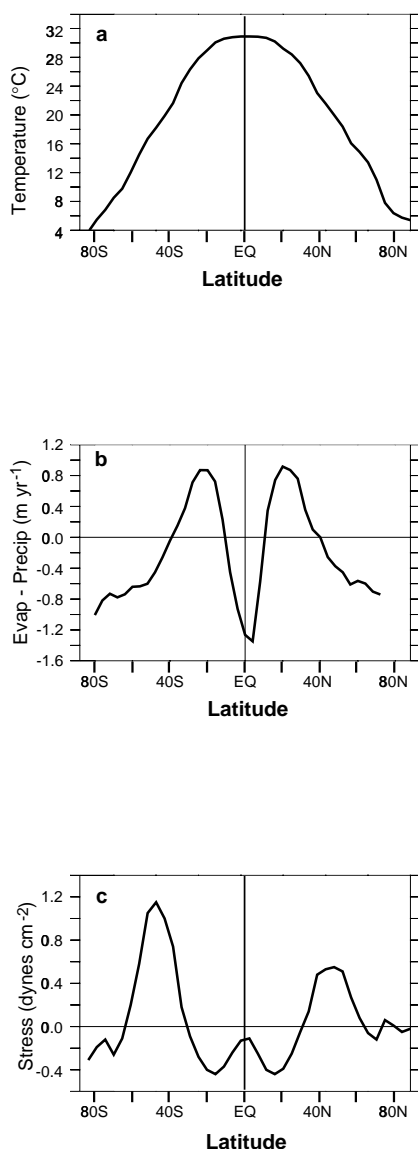


Figure 1. (a) Temperature, (b) evaporation minus precipitation rate for the control case, and (c) zonal wind stress forcings from the atmospheric model.

latitude sea surface temperatures, just prior to the PETM. The AGCM yields an ocean surface temperature of 11°C at paleolatitude 63°S, a value within the uncertainty in the 13-14°C estimate from the planktonic foraminifera *Acarinina* at ODP Site 690 below the PETM [Stott *et al.*, 1990; Thomas and Shackleton, 1996].

Initially, the basin configuration used is the same as that used by Bice and Marotzke [2001], based on an early Eocene reconstruction by Chris Scotese [Bice *et al.*, 2000a] (Figure 2). Paleobathymetry is based on seafloor age-depth relationship as described by Bice *et al.* [1998]. Important paleogeography sensitivity tests are presented below. A detailed discussion of the modeling approach, its limitations, and further sensitivity tests are given in Bice and Marotzke [2001].

The ocean model is run for 2700 years to near steady-state with the E-P forcing predicted by the AGCM and corrected to account for continental runoff [Bice and Marotzke, 2001]. The moisture flux is then multiplied, uniformly, by factors increasing from 1.0 to 2.0 in steps of 0.1 every 500 years, representing a simple linear increase in the hydrologic cycle that produces higher evaporative fluxes in the subtropics and higher net precipitation at high latitudes, while maintaining a net zero surface flux. This imposed perturbation of E-P represents the increase in water cycle simulated in coupled models with increased atmospheric CO₂, as described in Section 1. Thus, the uncoupled ocean model used here parameterizes one of the most important changes of the coupled system, albeit in a relatively crude form.

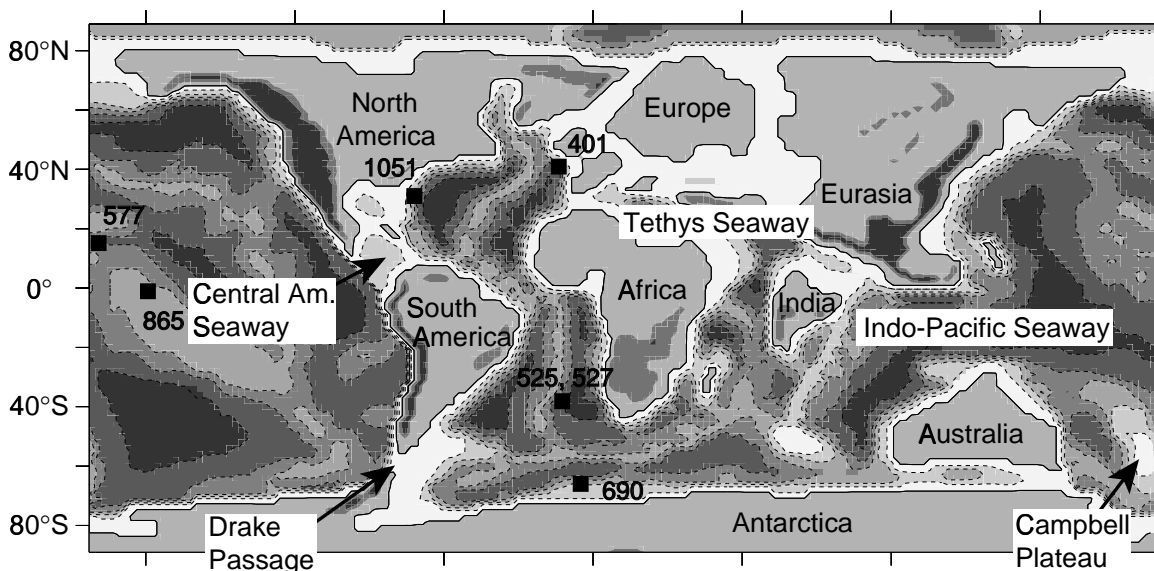


Figure 2. Baseline paleogeography at 2° resolution. Solid curve indicates the shoreline, Dashed contours indicate ocean depths of 1, 2, 3, 4, and 5 km. DSDP and ODP sites mentioned in the text are shown.

2.1 Baseline Paleogeography Results

Using the baseline paleogeography (Figure 2), the global circulation is very similar to the control run of *Bice and Marotzke* [2001], despite cooler polar temperatures, a greater pole-equator temperature gradient, and stronger zonal wind stresses. Deep water is formed in a broad region of the southern hemisphere along the Antarctic margin between Australia and the Antarctic Peninsula. This bottom water is carried west and northward along bottom topography east of Australia. It flows westward along the New Guinea margin into the eastern Indian Ocean, and is turned southward around Ninetyeast Ridge, Broken Ridge and Kerguelen Plateau, which had paleodepths as shallow as 500-1500 m in the early Paleogene [*Driscoll et al.*, 1989; *Peirce et al.*, 1989; *Quilty*, 1992]. Some deep water circulates northward into the deep western Indian Ocean. Deep water is prevented by topography from flowing directly into the deep Atlantic basins. Instead, it mixes with intermediate water across topography in the Southern Ocean, at Walvis Ridge and in the Equatorial Fracture Zone to fill these deep, silled basins. In the North Pacific, an intermediate (to 1500 m) watermass forms near the Alaskan margin and flows southwestward along the Asian margin. Much of this water is recirculated in the interior of the North Pacific basin; a smaller amount mixes with Southern Ocean water north of New Guinea to contribute to eastern Indian Ocean water.

Although the THC changes very little as E-P factor increases, the mean ocean temperature increases gradually, globally by about 1°C (black curve in Figure 3), with maximum warming (up to 7°C) occurring at between 400-1200 m in the subtropical North Atlantic. As discussed in detail by *Bice and Marotzke*, thermocline and upper intermediate waters warm from deepened *subduction*. Subduction is the process by which warm, relatively saline, subtropical surface waters are moved downward and spread horizontally along isopycnal (equal density) surfaces into the main thermocline [*Luyten et al.*, 1983; *Price*, 2001]. As E-P factor increases, small increases (less than 1 psu) in subtropical surface salinities result and cause subduction to carry subtropical water to greater depth, warming the thermocline. This response suggests deepened subduction as a plausible mechanism for mid-water warming and consequent methane release, but the magnitude of the subduction-induced warming is too small by itself to account for PETM bottom water warming at sites with intermediate paleodepths (sites 401, 690 and 1051, Table 1).

2.2. Sensitivity to the North Atlantic Basin Configuration

We now examine the sensitivity of the ocean response to this hydrologic cycle increase taking into consideration uncertainty in the precise paleogeographic reconstruction. To do this, the moisture flux perturbation experiment is repeated with three additional global ocean configurations, specifying either or both of the following changes: (a) North Atlantic basin modified to represent uplift and island construction associated with North Atlantic Volcanic Province magmatism and heat flow (Figure 4a, b), and (b) closed low latitude gateway with the earliest collision of western India and Eurasia (Figure 4 c, d). These changes represent plausible geographic configurations during the late Paleocene-early Eocene interval [*Eldholm and Thomas*, 1993; *Ritchie and Hitchen*, 1996; *Beck et al.*, 1998; *Knox*, 1998]. Of particular relevance to these experiments is evidence that extensive subareal volcanism existed in the late Paleocene in the North Atlantic [*Roberts et al.*, 1984; *Boulter and Manum*, 1989; *Clift et al.*, 1995; *White and Lovell*, 1997]. *Clift and Turner* [1998] describe evidence (including coal deposits) of subareal exposure in the latest Paleocene in the Greenland-Iceland-Faeroe Ridge region. Substantial relative sea level fall in the North Sea at the PETM due to high heat flow also supports the plausibility of a significant barrier to ocean flow in the North Atlantic region [*Knox*,

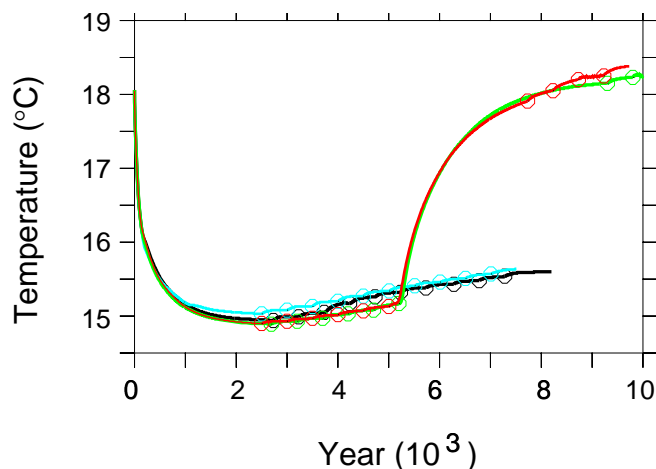


Figure 3. Time series of global mean ocean temperature for the four paleogeography sensitivity tests. The circles indicate the points at which the E-P forcing to the ocean model was increased by 10%. Temperature and wind stress forcings were held constant throughout. Colors indicate the following cases: black = open North Atlantic, open India-Eurasia gateway (Figure 4a, c); blue = open North Atlantic, closed India-Eurasia gateway (Figure 4a, d); green = Brito-Arctic bridge, closed India-Eurasia gateway (Figure 4b, d); red = Brito-Arctic bridge, open India-Eurasia gateway (Figure 4b, c).

1998]. Additionally, there is evidence to support the existence of a North Atlantic land bridge as a high latitude corridor for mammal migration from Europe to North America via Greenland at the Paleocene-Eocene boundary [Knox, 1998; Tiffney, 1998]. In fact, the appearance of many mammal orders in North America by migration across this bridge and a possible Bering land bridge occurs precisely coincident with the terrestrial record of the PETM [Koch *et al.*, 1995; Maas *et al.*, 1995]. We will use the shorthand “Brito-Arctic bridge” to refer to a plausible barrier to flow in the North Sea region associated with magmatism and high heat flow activity of the Iceland Plume in the latest Paleocene.

All three sensitivity experiments exhibit thermocline and upper intermediate water warming due to deepened subduction. However, when a Brito-Arctic bridge is specified in the North Atlantic basin, with India-Eurasia either open or closed, the THC undergoes an abrupt switch from dominant southern hemisphere sinking to northern hemisphere (Pacific) sinking at a factor of 1.6 x (E-P). Northern hemisphere overturning increases abruptly from 20 Sv to 45 Sv (Figure 5a), with what had been intermediate depth convection near the Alaskan margin deepening abruptly to 5200 meters. Bottom water formation abruptly decreases along the Antarctic margin region, but intermediate depth (~1200 m) convection continues in a small area to the west of the Antarctic Peninsula. Southern hemisphere THC strength decreases from 60 Sv to 12 Sv (Figure 5b). This switch in THC produces 3-5°C warming in the deep ocean. Prior to the switch, 11-12°C surface water convected at 74-80°S in the Pacific sector Southern Ocean. Bottom water warming occurs with the switch to convection of 15-16°C surface water at ~66°N on the Alaskan margin. The global mean ocean temperature increases by 3°C over several thousand years.

The abrupt switch in THC occurs because North Pacific upper ocean salinities gradually increase, despite that fact that the stronger atmospheric water cycle causes more net precipitation at high latitudes. The primary mechanism for maintaining surface salinities in the northern North Pacific is increased transport of saline subtropical North Atlantic thermocline water through the Central American Seaway into the North Pacific basin. With the creation of a Brito-Arctic

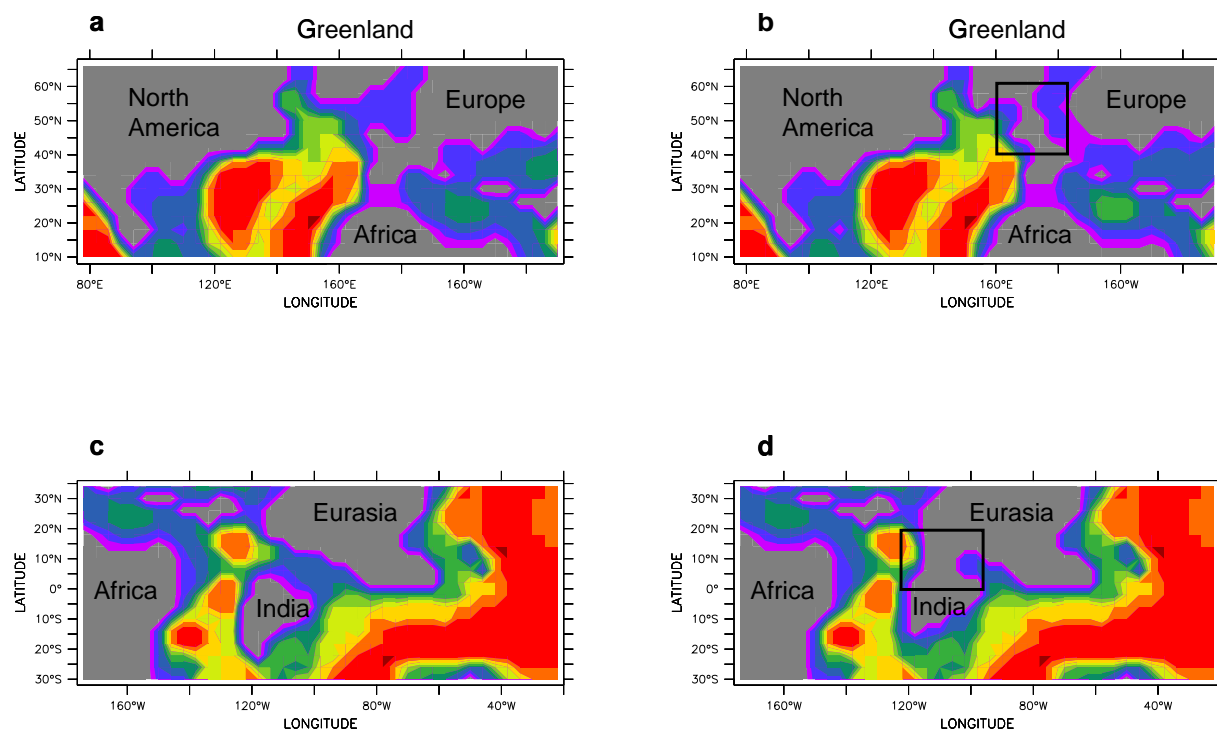


Figure 4. Maps showing regional paleogeographies examined in ocean model sensitivity experiments. (a) open North Atlantic, (b) Brito-Arctic bridge in the North Atlantic basin, (c) open India-Eurasia gateway, (d) closed India-Eurasia gateway. Gray = land; purple and blues = ocean shallower than 1500 m; orange and reds = ocean deeper than 3000 m.

bridge, low salinity North Sea water, which otherwise was introduced directly into the North Atlantic, is instead diverted south and eastward into the northern Tethyan basin. This allows moderately higher-salinity water to develop in the North Atlantic subtropics, which is transported southward and westward into the Pacific. There, much of it is entrained in the subtropical gyre and transported to the high-latitude North Pacific. The intermediate water sinking in the northern hemisphere is therefore not diminished as E-P increases from 1.0 to 1.5 (E-P). It is instead maintained, and North Pacific salinity slowly increases, by less than 0.2 psu (Figure 6), until finally a switch to northern hemisphere sinking occurs at a 60% increase in E-P. Bottom water now formed in the North Pacific fills deep basins north and east of the Emperor-Hawaiian seamount chain and topography associated with the Pacific Superswell. It also flows west and southwestward as a deep Asian boundary current and turns westward through the Indonesian passage, into the eastern Indian Ocean basin.

3. Proposed Feedback for the Paleocene-Eocene Thermal Maximum

The abrupt nature of the PETM bottom warming has long been interpreted as indicative of a THC switch, but the switch has been hypothesized to be from high southern latitude sinking to subtropical sinking [Kennett and Stott, 1991]. Our results suggest that the abrupt warming may instead have resulted from a switch from high southern to high northern latitude sinking at some critical point in the strength of the hydrologic cycle. This south-north switch in thermohaline circulation may therefore represent the “threshold” ocean response invoked for the PETM [Kennett and Stott, 1991; Zachos *et al.* 1993; Dickens *et al.*, 1995].

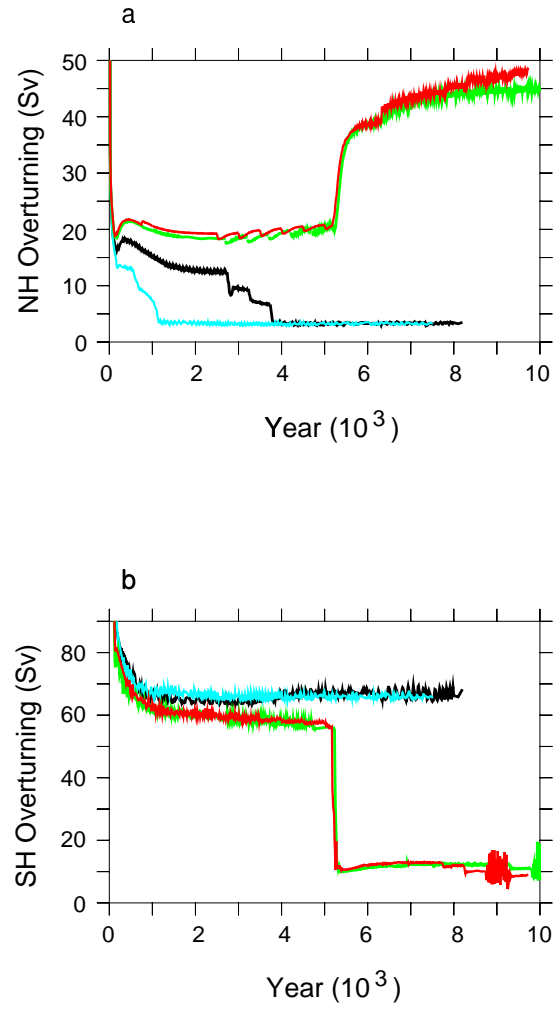


Figure 5. Time series of the maximum meridional overturning strength (Sverdrups) for (a) northern hemisphere and (b) southern hemisphere. Colors are as indicated for Figure 3.

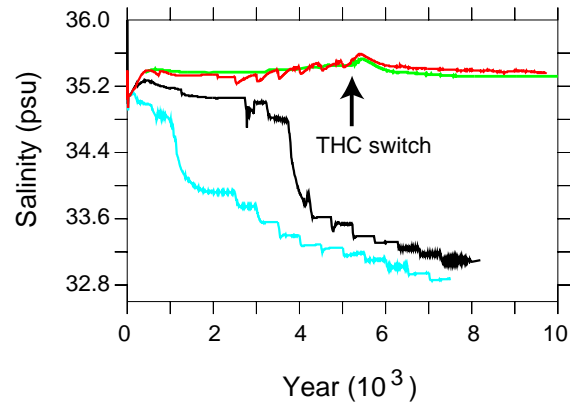


Figure 6. Time series of North Pacific upper ocean salinities in the region of convection, near the Alaskan margin. Colors are as indicated for Figure 3.

We propose a positive feedback (Figure 7) in which an increase in the strength of the hydrologic cycle (perhaps initiated by volcanic outgassing of CO_2) leads first to subduction warming at intermediate depths and spatially limited thermal hydrate destabilization. Some part of the methane thus released is oxidized to CO_2 in either the ocean or atmosphere [Dickens, 2000]. Increased atmospheric concentrations of CO_2 (and plausibly methane) further amplify the warming and hydrologic cycle increase, eventually causing an abrupt switch to sinking in the warmer northern hemisphere. The resulting widespread, abrupt bottom warming would plausibly result in widespread methane hydrate destabilization. This sequence would allow for sequential, or pulsed, injections of methane carbon, consistent with observations in high resolution records [Bains *et al.*, 1999; Röhl *et al.*, 2000], but the chain of events is also self-limiting with regard to methane release: the widespread destabilization caused by abrupt deep-ocean warming at the THC switch could occur only once, until sufficient time had passed for the regeneration of extensive methane hydrate and gas reserves.

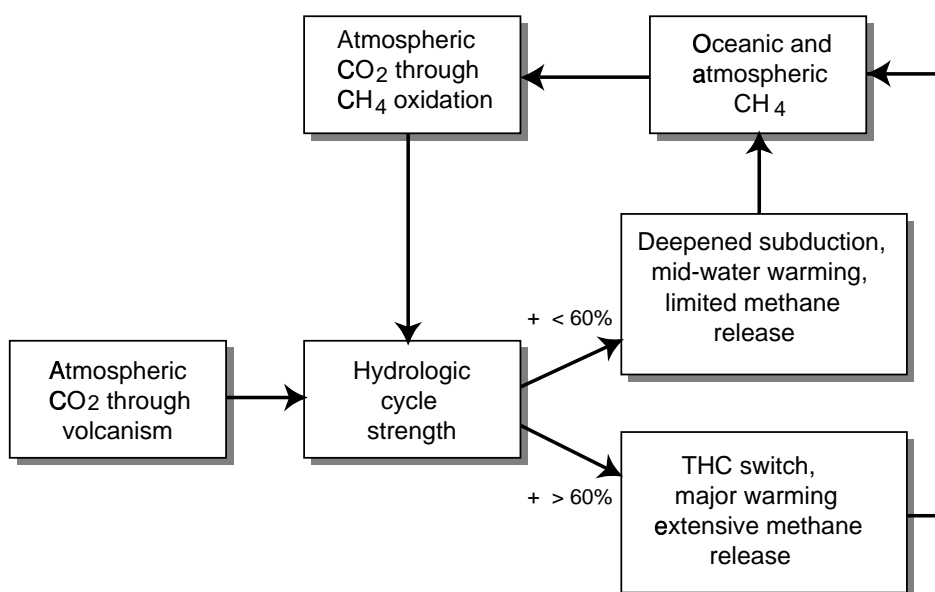


Figure 7. Hypothesized PETM feedback.

4. Model-Data Temperature Comparisons

The unperturbed (pre-PETM) temperatures and predicted temperature change are shown in Figure 8 for two north-south and west-east vertical sections through all ocean basins. Figures 8a and 8d show the model-predicted temperatures for the control E-P case (“pre-PETM”). Nearby deep sea sites from which paleotemperature estimates are available (Table 1) are projected onto the sections. Figures 8b and 8e illustrate the warming between the control E-P case and 1.5 x E-P, just prior to the THC switch. Figures 8c and 8f show the temperature difference between the control E-P case and 1.6 x E-P, after the THC switch has occurred and the ocean model has approached a new steady-state.

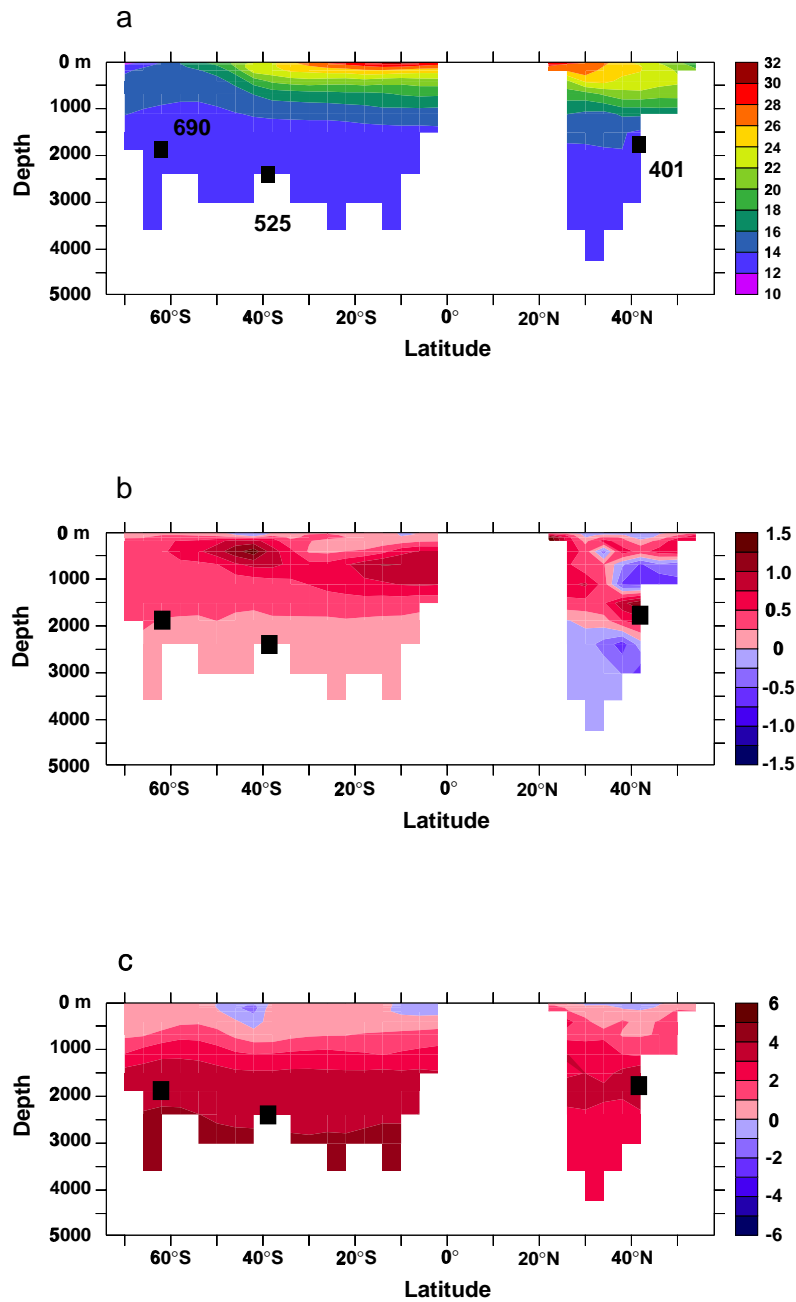


Figure 8. (a, b, c) Meridional cross sections through the Atlantic basins at the longitude of DSDP Site 401. (d, e, f) Zonal cross sections near the paleolatitude of Site 401 (38°N). Temperatures from the control case are shown in panels (a) and (d). Temperature differences due to deepened subduction are shown in panels (b) and (e). Temperature differences due to the THC switch are shown in panels (c) and (f). Positive values (reds) indicate warming caused by deepened subduction or by the switch from southern to northern hemisphere bottom water formation.

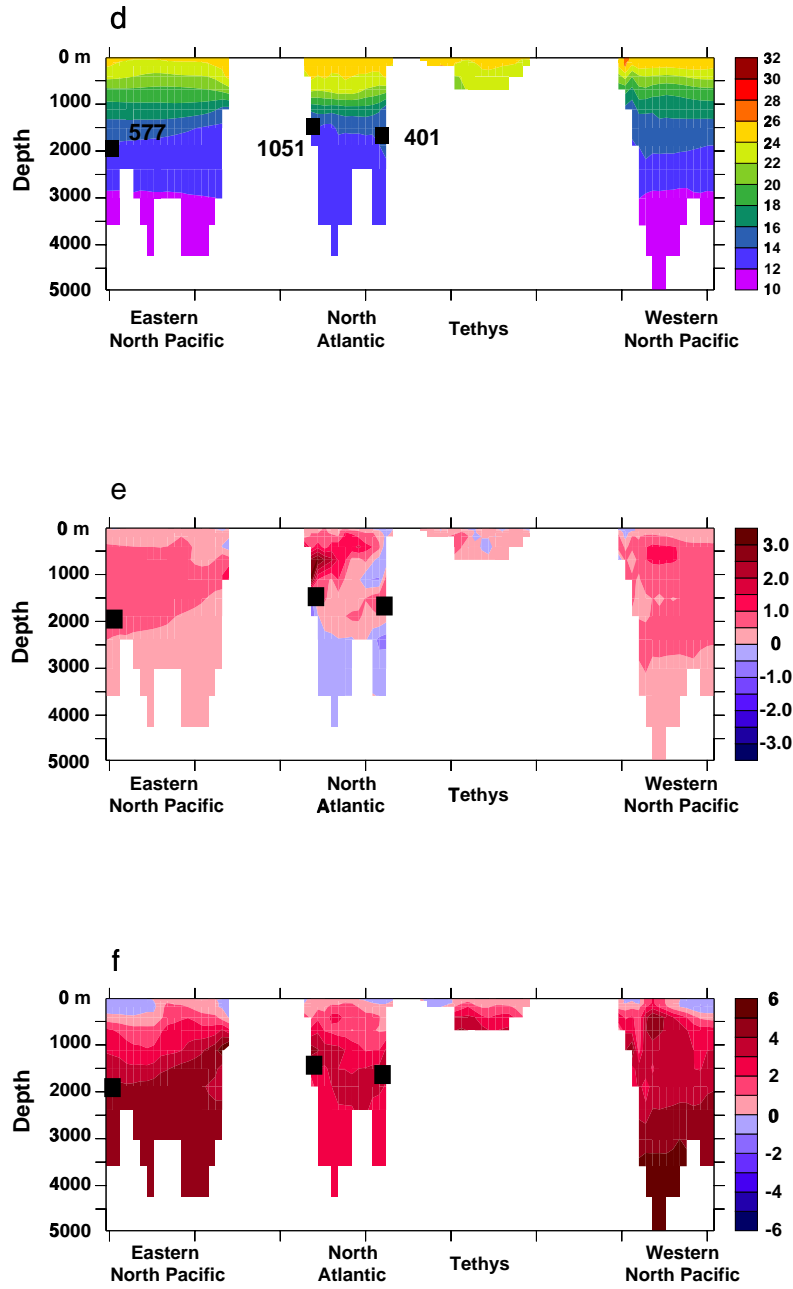


Figure 8. (continued)

One way to assess the validity of the predicted THC switch at the P/E boundary is to compare the model-predicted temperatures and temperature change against pre-PETM and PETM data. Unfortunately, there are very few localities from which both pre-PETM and peak PETM data are available, a result primarily of widespread dissolution of carbonate and hiatuses at the PETM [Thomas, 1998]. At those sites where sufficient carbonate material remains to allow recognition of the carbon isotope event, the size of the PETM excursion as expressed in the geochemistry of foraminifera is often truncated, allowing only a minimum temperature change to be inferred from $\delta^{18}\text{O}$. The sites from which benthic foraminiferal $\delta^{18}\text{O}$ measurements of pre-PETM and PETM (or immediately post-PETM) specimens are known are listed in Table 1. In order to have a somewhat larger dataset for comparison, data from DSDP Sites 401 and 577 are included, however the peak PETM is missing in benthic records from these sites (see Table 1 footnote). Peak PETM changes are well-pronounced at Site 1051, but there is evidence for an unconformity immediately prior to and at the onset of the PETM [Röhl *et al.*, 2000].

In terms of absolute temperatures, the model matches the inferred pre-PETM temperatures to within 2°C at all sites except Blake Nose Site 1051 (Figure 9). The model temperatures are therefore correct to within the estimated uncertainty in the Erez and Luz [1982] isotopic paleotemperature calculation [Bice *et al.*, 2000b]. The reason for the $\sim 4^\circ\text{C}$ model mismatch at 1500 m on Blake Nose is not obvious. Given the good match between the model and North Atlantic data farther north, at Site 401, one possibility is that the model underestimates the volume transport of cooler South Atlantic water into the North Atlantic basin across the Equatorial Fracture Zone region. However, the possibility of burial diagenesis can not be ruled out for Site 1051 PETM samples, given the current burial depth (512 m, Table 1) and post-Eocene erosion on Blake Nose [Norris *et al.*, 1998].

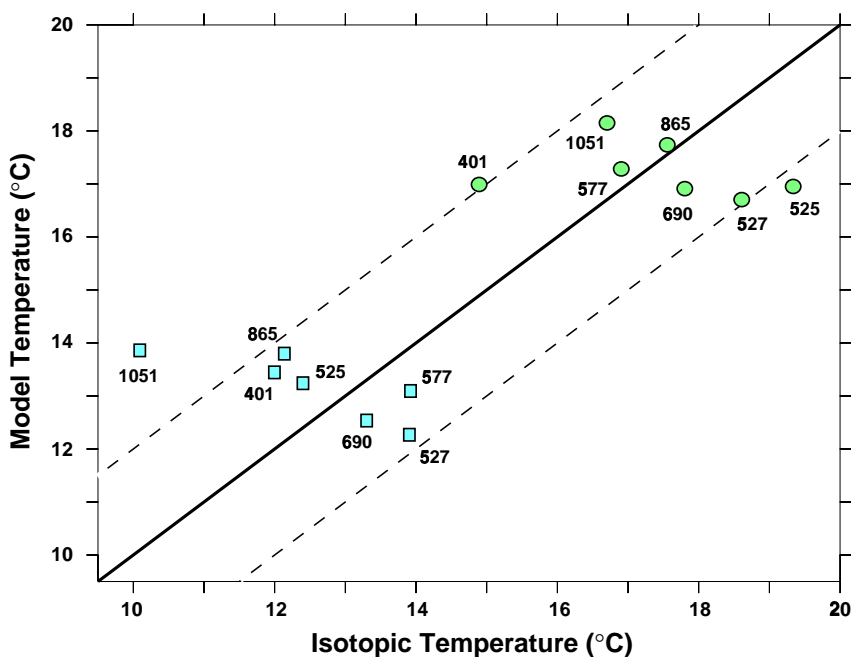


Figure 9. Comparison of model-predicted temperatures and isotopic paleotemperatures for pre-PETM (squares) and PETM (circles) cases. The thick solid line indicates a 1:1 correspondence. Thinner dashed lines indicate errors of $\pm 2^\circ\text{C}$. Numbers correspond to the ODP and DSDP sites listed in Table 1.

The pre-PETM to PETM data show that the magnitude of the change in water temperature inferred from $\delta^{18}\text{O}$ is similar to that simulated by the model at these sites when the THC switch occurs (Table 1). At 5 of the 7 sites, the temperature change predicted by the model is within 1.5°C of the temperature change inferred from $\delta^{18}\text{O}$, assuming no change in water oxygen isotopic composition at the PETM. The model under-predicts the temperature change at Sites 525 and 1051 by 3.3° and 2.4°C , respectively.

5. Predictions of the Hypothesis

In addition to the magnitude of temperature change, the proposed feedback mechanism suggested by the model results has a number of features that should be testable, if high temporal resolution (generally < 5 kyr) and depth-stratified marine isotopic data are obtained. We offer the following general predictions:

(1) Gradual surface warming would accompany (and drive) an increase in the hydrologic cycle. Next, thermocline and upper intermediate depths will warm gradually in some areas due to deepened subduction. Finally, abyssal waters warm abruptly everywhere due to the THC switch. Deepened subduction alone is unlikely to cause deep water warming. The PETM and interval immediately preceding it should therefore exhibit a largely “top-down” warming. *Röhl et al.* [2000] interpret the high resolution record from Site 1051 as indicating both gradual and catastrophic inputs of isotopically depleted carbon to the ocean system. This is consistent with our prediction of two mechanisms for hydrate destabilization, operating on different time scales.

(2) Though local warming (at any depth) would precede the thermal destabilization of hydrate *at that site*, subduction warming is not uniform and the distribution of hydrates is not likely to have been global. At sites with no methane destabilization, carbonate $\delta^{18}\text{O}$ values may decrease (indicating warming) prior to a decrease in $\delta^{13}\text{C}$. Similarly, because of watermass transport and mixing, methane release in one locality could produce a $\delta^{13}\text{C}$ decrease at a site that has not yet experienced warming. Therefore, a decrease in $\delta^{13}\text{C}$ that precedes evidence of warming at the same site is not necessarily evidence against the methane hydrate hypothesis [*Dickens et al.*, 1995]. The temporal resolution necessary to observe such leads and lags is likely to be on the order of thousands of years, requiring high resolution sampling in areas of high sedimentation rate.

(3) Prior to the massive input of light carbon at the PETM, destabilization of hydrates due to deepened subduction would have occurred at progressively greater depths. This progressive response could have been smooth or pulsed, depending on the true rate of change in atmospheric moisture transports and the location of significant clathrate reservoirs relative to regions of maximum warming. *Bains et al.* [1999] and *Röhl et al.* [2000] describe records from Blake Nose Site 1051 that are consistent with a pulsed release of methane carbon prior to the massive global carbon isotope excursion. Regarding a progressively deepening response, it may be unclear even from high resolution, depth-stratified geochemical records if destabilization occurred at progressively deeper water depths, because a $\delta^{13}\text{C}$ signal input at a given depth could presumably have been propagated up or down the water column, depending on local mixing conditions.

(4) Subduction-induced warming would have been greatest in the Atlantic basin and in the regions of the upper ocean western boundary currents, as described by *Bice and Marotzke* [2001] and shown in Figure 8b, e. This suggests that it is most likely that early widespread methane hydrate destabilization occurred in the Atlantic. In an ocean carbon cycle model, *Dickens* [2000] showed that methane release in the Atlantic basin produces a better match to several features of

the PETM record than does release in other oceans. Evidence of PETM calcium carbonate dissolution (assumed to be an indicator of decreased local carbonate ion concentration due to CH₄ oxidation) is apparently most pronounced in the Atlantic ocean [Thomas, 1998], further supporting the idea that extensive methane release occurred in the Atlantic [Katz *et al.*, 1999].

(5) The THC should return to Southern Hemisphere deep water formation as radiative gases input to the atmosphere before and during the PETM are removed by long- and short-term carbon cycle processes [Bains *et al.*, 2000; Beerling, 2000; Ravizza *et al.*, 2001]. When we decrease E-P from its maximum value of 2.0 to the initial profile, the ocean cools, first gradually as subduction shallows, then abruptly, when the THC returns to its original state with deep sinking in the southern hemisphere. This cooling, driven by a hydrologic cycle “spin down,” is consistent with observations showing that benthic $\delta^{18}\text{O}$ values (indicating temperature) decrease to near pre-PETM values over less than 100 kyr [Röhl *et al.*, 2000]. The model switch back to the original THC state occurs at a lower E-P (1.2 x control) than does the forward switch (1.6 x control), indicating multiple equilibria in a small parameter range for paleogeographies with a Brito-Arctic bridge.

A few of the above predictions are broadly consistent with changes inferred from existing records, but most await new high resolution marine records. We note that the prediction of relative timing (i. e., leads and lags between surface, intermediate and bottom water $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ decreases) is not at all straightforward. The distribution of late Paleocene hydrates is unknown and the model ocean exhibits important longitudinal and latitudinal variations in response to subduction warming, so that it is impossible to make firm predictions about relative timing of records at any particular site.

6. Concluding Remarks

If the PETM included a switch in thermohaline circulation and massive methane release, why did the event occur at this time? Our results indicate a strong sensitivity to North Atlantic tectonism: the THC switch and subsequent widespread bottom warming occur always and only when a barrier representing possible Greenland-Iceland-Faeroe Ridge uplift is specified in the model configuration. This suggests that the Paleocene-Eocene paleogeography played a critical role in the response of the ocean system to a gradual perturbation of the atmospheric hydrologic cycle. It may be interpreted as a measure for how many influences must conspire, even in a warm climate, to produce an event as remarkable as the PETM. Still, we caution that there exists substantial uncertainty in the exact details of the Paleocene/Eocene ocean configuration. With any ocean model, is it necessary to explore the sensitivity of the model solution to fundamental uncertainty in this boundary condition, but it is impossible to test all possible configurations. We therefore offer these results as representative of one part of the spectrum of possible THC solutions and sensitivity for the Paleocene-Eocene ocean.

The model runs shown here have employed some significant idealizations, such as the use of zonal mean atmospheric boundary conditions or the absence of an explicit atmospheric model. While either simplification is expected to change the results quantitatively, we argue that qualitatively the results are robust. The existence of, and transitions between, multiple equilibria of the thermohaline circulation is a robust phenomenon, across the range of simple and complex, coupled and uncoupled models [e.g., Manabe and Stouffer, 1988; Marotzke, 1990; Marotzke and Willebrand, 1991; Weaver and Hughes, 1996]. In mathematical terms, the bifurcation structures of coupled and uncoupled models of the THC are very similar [Marotzke, 1996]. Using an uncoupled model might introduce quantitative differences, but it is unlikely to “get things

wrong” qualitatively. The justification of a simple model approach is discussed further by *Bice and Marotzke* [2001].

Although we impose a zonal mean forcing to the ocean model, in reality, there is certainly west-east asymmetry in the atmospheric hydrologic cycle, with generally drier conditions over the eastern subtropical oceans [*Peixoto and Oort*, 1992]. This asymmetry is apparent in PETM records from subtropical latitudes of the North Atlantic. *Gibson et al.* [2000] describe clay mineral changes in the Salisbury Embayment (VA-NJ coastal plain) that suggest an increase in drainage basin-integrated moisture flux over some part of eastern North America. At the same time, PETM drying is supported by mineralogical and isotopic studies in the subtropical eastern North Atlantic and western Europe [*Schmitz et al.*, 2001; *Schmitz and Andreasson*, 2001]. The Salisbury embayment record suggests that western North Atlantic embayment surface waters could have become less saline due to increased runoff during the PETM, but it is unlikely that such a change would prevent deepened subduction in the North Atlantic; The subduction process that transports warmer, saline water into the thermocline tends to remove surface water from the eastern side of the subtropical gyre and deposit it in the western deep thermocline [*Price*, 2001]. The decreased salinity waters inferred by *Gibson et al.* may therefore have existed as a lower density mass in the subtropical marginal western North Atlantic and may have had little influence on thermocline ventilation by subduction.

How literally should we read the fact that, in experiments where a THC switch occurred, it occurred at a 60% increase in the strength of the hydrologic cycle? Given the use of zonal mean forcings and the lack of an ocean-atmosphere feedback in the uncoupled model, we would argue only that, if such a switch occurred, it occurred at “some critical value” of moisture transport. Climate models, both coupled and uncoupled, differ in their prediction of the magnitude of increase in the water cycle given some magnitude increase in atmospheric CO₂ [*Manabe et al.*, 1994], and many fewer studies have examined CH₄-induced changes in the hydrologic cycle. How strong the positive feedback in our hypothesized system is would have to be investigated using reliable coupled physical-chemical atmosphere-ocean models. Strict quantification of forcings and feedbacks in such a system is probably not feasible with existing coupled models, given that such models being used for future climate change research do not agree on the fundamental question of whether a future doubling of atmospheric CO₂ will stabilize or destabilize the modern thermohaline circulation [*Manabe and Stouffer*, 1993; *Rahmstorf and Ganopolski*, 1999; *Latif et al.*, 2000].

The precise location and amount of PETM methane release resulting from gradual subduction warming and abrupt THC switch-induced warming are difficult to predict. The many complexities of late Paleocene marine productivity, hydrate formation and stability, sediment stability, and controls on hydrate phase change under gradual and abrupt warming in already “warm” oceans remain to be sorted out by the diverse research community studying the Paleocene-Eocene Thermal Maximum. The results presented here raise the intriguing possibility that an abrupt change in thermohaline circulation from southern to northern sinking was an integral part of the PETM and that this change could have resulted from a gradual increase in the water cycle under conditions of late Paleocene volcanism and Paleocene-Eocene paleogeography.

Acknowledgements. This material is based upon work supported by the National Science Foundation under Grant No. ATM-9905023 to KLB and under Grant No. ATM-0000545, which supports the Partnership in Modeling Earth System History (PSU/WHOI). The authors benefited

from discussions with J. Pedlosky, R. Norris, D. C. Kelly and J. Scott. We thank J. Dickens and E. Thomas for helpful comments on earlier versions of the manuscript. The MOM2 ocean model was developed under the direction of R. Pacanowski at the Geophysical Fluid Dynamics Laboratory. The GENESIS atmospheric model was developed at the National Center for Atmospheric Research by S. Thompson and D. Pollard. The baseline paleogeography was provided by C. Scotese. The GENESIS model was run on the Cray SV1 system at Penn State University's Environment Computing Facility. Model data visualization was done using Ferret, a freely distributed software package developed by the Thermal Modeling and Analysis Project at NOAA/PMEL. This is WHOI Contribution No. xxxxx.

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Table 1. Comparison of temperatures inferred from $\delta^{18}\text{O}$ of benthic foraminifera across the PETM and those predicted by the ocean model.

Site	Estimated Water Paleodepth (m)	pre-PETM and PETM Sample Depths (mbsf)	pre-PETM and PETM $\delta^{18}\text{O}$ (‰ PDB)	pre-PETM to PETM ΔT ($^{\circ}\text{C}$) from isotopes	Maximum ΔT Response to Subduction	Mean ΔT ($^{\circ}\text{C}$) Response to THC Switch
401	1800	202.60, 202.00	0.11, -0.54 ^a	2.9	1.1	3.6
525	1600	392.81, 392.41	0.03, -1.50 ^b	6.9	0.9	3.6
527	3400	201.07, 200.89	-0.32, -1.35 ^b	4.7	0.4	4.4
577	1950	82.03, 81.94	-0.32, -0.98 ^c	3.0	0.5	4.2
690	1900	170.65, 170.42	-0.18, -1.17 ^d	4.5	0.4	4.4
865	1400	103.00, 102.96	0.0, -1.16 ^e 0.15, -1.09 ^f	5.2 5.6	0.6	4.0
1051	1500	512.90, 512.60	0.54, -0.93 ^g	6.7	2.5	4.3

The isotopic temperature calculation [Erez and Luz, 1982] assumes local bottom water $\delta^{18}\text{O}$ equaled -1‰ (SMOW) and did not change across the PETM.

^a*Cibicidoides* [Pak and Miller, 1992] The peak PETM is missing in benthic records from Site 401, so the $\Delta\delta$ recorded here is an underestimate of the actual change.

^b*Nuttallides* [Thomas and Shackleton, 1996]

^c*Nuttallides* [Pak and Miller, 1992] Samples across the P/E boundary at Site 577 do not contain the PETM. The $\Delta\delta$ recorded here is for pre-PETM to post-PETM specimens.

^d*Bulimina ovula* [Thomas and Shackleton, 1996; Thomas et al., 2000]

^e*Cibicidoides* [Zachos et al., 2001]

^f*Bulimina ovula* [Zachos et al., 2001]

^g*Oridorsalis* [Katz et al., 1999] The onset of the PETM is missing in Site 1051, but peak PETM values appear to have been recorded [Röhl et al., 2000].