Atmospheric Moisture Transport Moderates Climatic Response to the Opening of Drake Passage

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(Manuscript received 13 February 2008, in final form 28 September 2008)

ABSTRACT

The absence of the Drake Passage (DP) gateway in coupled models generally leads to vigorous Antarctic bottom water (AABW) formation, Antarctic warming, and the absence of North Atlantic deep-water (NADW) formation. Here the authors show that this result depends critically on atmospheric moisture transport by midlatitude storms. The authors use coupled model simulations employing geometries different only at the location of DP to show that oceanic circulation similar to that of the present day is possible when DP is closed and atmospheric moisture transport values enhanced by Southern Ocean storm activity are used. In this case, no Antarctic warming occurs in conjunction with DP closure. The authors also find that the changes in poleward heat transport in response to the establishment of the Antarctic Circumpolar Current (ACC) are small. This result arises from enhanced atmospheric moisture transport at the midlatitudes of the Southern Hemisphere (SH), although the values used remain within a range appropriate to the present day. In contrast, homogeneous or (near) symmetric moisture diffusivity leads to strong SH sinking and the absence of a stable Northern Hemisphere (NH) overturning state, a feature familiar from previous studies. The authors’ results show that the formation of NADW, or its precursor, may have been possible before the opening of the DP at the Eocene/Oligocene boundary, and that its presence depends on an interplay between the existence of the DP gap and the hydrological cycle across the midlatitude storm tracks.

1. Introduction

The global ocean and atmosphere are subject to different terrestrial boundary conditions in each hemisphere as the continental landmasses are distributed asymmetrically about the equator. A higher land concentration in the Northern Hemisphere (NH) and the absence of an NH equivalent to the circumpolar Southern Ocean (SO) lead to significant differences in the general circulation of the ocean and atmosphere between the hemispheres. Unlike the NH, the absence of zonal boundaries at the latitudes of Drake Passage (DP) allows the development of the deep Antarctic Circumpolar Current (ACC) in the ocean, and precludes geostrophic flow across the DP gap and above the DP sill. In the atmosphere, the Southern Hemisphere (SH) westerly winds are stronger than their NH counterparts (e.g., see Kalnay et al. 1996), as is SH poleward moisture transport as a result of the mean flow and atmospheric eddies [e.g., see the reanalysis data of Wenzel et al. (2001)].

Process studies using ocean general circulation models have sought to examine the consequences of removing the latitudinal asymmetry posed by the DP by running numerical models to equilibrium while applying a small fictitious land bridge between the Antarctic Peninsula and the tip of South America (Mikolajewicz et al. 1993; Toggweiler and Bjornsson 2000; Nong et al. 2000; Sijp and England 2004, 2005), thus “closing” the DP. Despite the variety of models used [see Sijp and England (2005) for an overview], these studies each exhibit the development of a large SH overturning cell in response to the closing of the DP, and the absence of NH overturning when no restoring boundary conditions are used. Sijp and England (2005) examine multiple equilibria in a geometry where DP is closed, and are unable

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DOI: 10.1175/2008JCLI2476.1

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to shut down this vigorous Antarctic bottom water (AABW) using freshwater (FW) perturbations, and are equally unable to induce a steady NH overturning state.

The above model studies were also intended to examine the climatic consequences of the opening of the circumpolar ocean in the geological past. The ACC has not been present throughout the entire Cenozoic, and the timing of its onset is uncertain (e.g., Barker and Thomas 2004). The presence of obstructing bathymetry in the Tasman Seaway south of Australia and the Drake Passage constitute the final barriers to a mature circumpolar flow. The opening of the Tasman Seaway to deep flow is estimated to be $32 \times 10^6$ yr before present (32 Ma) at the latest and is relatively well constrained (Lyle et al. 2008; Lawver and Gahagan 2003; Stickley et al. 2004), whereas the timing of the opening of Drake Passage remains uncertain. An early Oligocene DP opening older than 30.5 Ma (Lawver and Gahagan 2003; Livermore et al. 2005) and an early Miocene opening between 22 and 17 Ma (Barker and Burrell 1977) have been suggested. Kennett (1977) and Berggren and Hollister (1977) hypothesized that the apparently close temporal proximity between the opening of the Tasman Seaway and the glaciation of Antarctica at the Eocene/Oligocene boundary (35–33 Ma) may indicate a causal relationship. Here, the thermal isolation of Antarctica resulting from the development of the ACC leads to favorably cool conditions for the emergence of a terrestrial ice sheet. The 1°–3°C cooling in response to opening the DP inferred by modeling studies lends some support to this contention. However, recent improvements in age estimates by Stickley et al. (2004) suggest that the opening of the Tasman Seaway occurred about 2 Ma before the onset of Antarctic glaciation, making a causal connection less likely. Nonetheless, Livermore et al. (2005) argue that the opening of DP constituted a trigger for Antarctic glaciation at the Eocene/Oligocene boundary based on their timing estimates. However, Huber et al. (2004) find no enhanced poleward heat transport (HT) in their coupled model of the Eocene where the Tasmanian Seaway is closed, and conclude that the opening of any of the SO gateways is unlikely to have caused Antarctic glaciation. In addition, recent modeling work by DeConto and Pollard (2004) indicates that changes in atmospheric carbon dioxide may have played a more significant role.

The present bipolar mode of deep-water formation in the Northern and Southern Hemispheres has not been a persistent feature of the Cenozoic ocean circulation. There is increasing evidence to support the Southern Ocean as the dominant region of deep-water formation during the early Paleogene (e.g., Mountain and Miller 1992; Via and Thomas 2006). North Atlantic (NA) sinking may have been inhibited because the northern North Atlantic basins had not yet fully opened (Saunders et al. 1997). Nonetheless, Corfield and Norris (1996) find evidence of deep-water production in both the Southern Hemisphere and the North Atlantic also during the early to late Paleocene, followed by a sole southern source of deep water in the latest Paleocene. It should be noted that the existence of the Central American Seaway until its final closure 3 Ma may have modified but not precluded deep-water formation in the North Atlantic (Huber and Sloan 2001; Nisancioglu et al. 2003). Thomas et al. (2008) find that the deep Pacific may have been ventilated from the Southern Ocean and the North Pacific (NP) for much of the early to middle Paleogene. Neodymium isotope measurements by Scher and Martin (2008) suggest an increase in export to the SO of deep water formed in the NA concurrent with the maturation of the ACC over the Oligocene–Miocene period. The apparent synchronicity with increases in the flux of Pacific water through DP in the late Oligocene supports hypotheses that link North Atlantic deepwater (NADW) formation to the maturation of the ACC. This is in agreement with Mikolajewicz et al. (1993), Toggweiler and Bjornsson (2000), and Sijp and England (2004, 2005), whose numerical studies show that the presence of the ACC favors the formation of NADW. Furthermore, some of these modeling studies (see Sijp and England 2005) show an absence of stable NH overturning states in the DP “closed” case, suggesting the absence of the precursor of NADW, northern component water, during the Eocene.

The more idealized model studies investigating the DP effect do not allow the wind field and associated moisture transport to change. In contrast, model studies using a fully coupled model (e.g., Huber et al. 2004) and reconstructions of Eocene topography allow a sophisticated atmospheric response, but are more difficult to analyze and do not belong to the paradigm of process studies. Here, we aim to bridge this gap by examining the response of a model of intermediate complexity to the opening of the DP as a function of the hydrological cycle. We will demonstrate that the response of the climate system to the opening of the DP is critically dependent on the hydrological cycle.

2. The model and experimental design

We use the University of Victoria intermediate complexity coupled model (UVic) described in detail in Weaver et al. (2001). This comprises an ocean general circulation model [Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model (MOM) version 2.2; Pacanowski 1995] coupled to a simplified one-layer
energy–moisture balance model for the atmosphere and a dynamic–thermodynamic sea ice model of global domain and horizontal resolution 1.8° latitude × 3.6° longitude. Air–sea heat and freshwater fluxes evolve freely in the model, yet a noninteractive wind field is employed. The wind forcing is taken from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis fields (Kalnay et al. 1996), averaged over the period 1958–97 to form a seasonal cycle from the monthly fields. Vertical mixing is modeled using a diffusivity that increases with depth, taking a value of 0.3 cm² s⁻¹ at the surface and increasing to 1.3 cm² s⁻¹ at the bottom. We use version 2.7 of the UVic model. The effect of subgrid-scale eddies on tracer transport is modeled by the parameterizations of Gent and McWilliams (1990).

The model calculates humidity as a prognostic variable associated with the one-layer atmosphere. Precipitation occurs as snow or rain once relative humidity is greater than 85%, whereby instantaneous runoff from land occurs via prescribed river basins. Moisture transports are accomplished through advection (by wind) and Fickian diffusion. The atmosphere employs a spatially varying moisture diffusion coefficient, as shown in Fig. 1, except for the control (CNTRL) case, which adopts a constant value of 1 × 10⁶ m² s⁻¹. This CNTRL case is equivalent to the profile used in our previous investigations of the DP effect (Sijp and England 2004, 2005). Moisture diffusion in the model represents transport by subgrid-scale atmospheric eddies (the low pressure systems), and thus allows a direct physical interpretation. The higher values of the moisture diffusion coefficient in the SH in experiment ASYMM reflect the stronger transports by atmospheric eddies at midlatitudes in the SH. The standard configuration of the UVic model employs the moisture diffusion coefficient “ASYMM” shown in Fig. 1a. Saenko et al. (2003) show that this higher moisture diffusion in the SH leads to more realistic zonally averaged oceanic freshwater transports (Fig. 1c) in this model for the present day.

We have run an experiment with the DP at its present-day depth, DPopen, and an experiment with the DP closed by a small land bridge, DPclsd, using the standard moisture diffusion coefficient ASYMM. We denote these experiments by DPopen ASYMM and DPclsd ASYMM respectively. To examine the effect of the magnitude of the difference in NH and SH moisture diffusion, we have also run a set of experiments employing a diffusion coefficient field ASYM2/3 whereby the maximum SH moisture diffusion coefficient is reduced from 3 × 10⁶ m² s⁻¹ to 2 × 10⁶ m² s⁻¹ (Fig. 1a). To examine the effect of symmetry in the moisture diffusion coefficient field, we have also run our DPopen and DPclsd experiments using a constant moisture diffusion coefficient, denoted
CNTRL, and an experiment SYMM where the SH moisture diffusion enhancement of ASYMM is mirrored about the equator (Fig. 1b). All experiments have been integrated from idealized initial conditions for 7000 yr and yield stable equilibria.

3. Overturning circulation response

The global atmospheric moisture transport in the model must be compensated by an opposite and equal global oceanic FW transport. The asymmetric moisture diffusion fields ASYMM and ASYMM2/3 (Fig. 1a) therefore lead to an asymmetric oceanic northward FW transport field (Fig. 1c) in DPclsd, whereby the larger SH moisture diffusion leads to larger SH oceanic FW transport from the subpolar regions to the subtropics. The FW transports for DPopen are very similar to their DPclsd counterparts and are therefore not shown. The symmetric moisture diffusion fields CNTRL and SYMM lead to near-symmetric oceanic FW transport patterns (Fig. 1d).

We will now discuss the CNTRL and the ASYMM experiments in more detail, and refer to Table 1 for the equivalent results obtained from the ASYMM2/3 and SYMM experiments.

Figure 2 shows the meridional overturning circulation (MOC) for DPopen and DPclsd for the CNTRL case. Not surprisingly the results are similar to the study of Sijp and England (2004). The DPopen CNTRL experiment exhibits 13.0 Sv (1 Sv = 10^6 m^3 s^-1) of NADW formation (Fig. 2b), an abyssal overturning cell of about 15.4 Sv, and a Deacon cell of 31.1 Sv (Fig. 2a and Table 1). In contrast, NADW is absent when the DP is closed (Fig. 2d) and a strong SH overturning cell of 34.9 Sv develops, along with 7.9 Sv of AABW inflow into the Atlantic. This result is in agreement with previous studies examining the closing of the DP (e.g., Mikolajewicz et al. 1993; Sijp and England 2004, 2005). The strength of this cell results from the vigorous oceanic heat loss around Antarctica. A similar result is obtained for the DPclsd SYMM experiment, where a vigorous AABW cell of 29.3 Sv develops (see Table 1), despite the southward diffusion of moisture now present over the Southern Ocean. In contrast, when employing the asymmetric moisture diffusion field ASYMM, we do not obtain the characteristic strong SH overturning when the DP is closed. Instead, there is only 7.6 Sv of sinking around Antarctica (Fig. 3c), and unlike experiment DPclsd CNTRL, there is 15.1 Sv of sinking in the North Atlantic, with 9.5 Sv of outflow at 33^oS (Fig. 3d). This is because of the enhanced moisture transport from the subtropics to the subpolar regions in the SH compared to DPclsd CNTRL, leading to freshening (as we will see in Fig. 4) inhibiting Antarctic sinking, and boosting NADW formation. Similarly, NADW is enhanced to 21.4 Sv in DPopen ASYMM (Fig. 3b) when DP is open, because of the enhanced salinity gradient between the Antarctic Intermediate Water (AAIW) and NADW formation regions. The results for DPclsd, along with the results of the ASYMM2/3 experiments, are listed in Table 1. Also, in experiment DPclsd ASYMM2/3, where the enhanced SH FW transport is reduced, there are similar results to DPclsd CNTRL, with vigorous SH overturning (31.8 Sv). This shows that this more weakly asymmetric moisture diffusion field does not yield a preference for NH overturning in DPclsd. These results demonstrate that the NH overturning state found in the asymmetric case in DPclsd results from the magnitude of the asymmetry in the moisture diffusion field, and that symmetric moisture diffusion fields (CNTRL and SYMM) and weakly asymmetric states (ASYMM2/3) lead to the previously found vigorous SH overturning state.

Figure 4 shows the temperature–salinity (T–S) properties at the sea surface in the sinking regions for DPopen and DPclsd in the asymmetrical case (ASYMM) and DPclsd in the CNTRL case. The higher temperature and salinity at the NADW formation regions in both the DPopen ASYMM and DPclsd ASYMM cases reflect the presence of NADW formation, whereby the thermohaline circulation (THC) engenders enhanced transport of warm saline subtropical water into the subpolar gyre. Despite the significantly higher temperature, the NADW formation site is denser than the AABW formation site in DPclsd ASYMM resulting from the more saline conditions over the North Atlantic. Sijp and England (2005) show that in a geometry where the DP is closed, the strength and polarity of the MOC is determined by the density contrast between the AABW and NADW formation regions. This explains why DPclsd ASYMM exhibits stable NADW formation. When comparing the

<table>
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<th>NADW (Sv)</th>
<th>Brazil Current (Sv)</th>
<th>δ SAT SH (°C)</th>
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properties of the AABW formation site for the three experiments shown, we see that the freshest conditions prevail in DP_{clsd} ASYMM, whereas temperatures remain similarly cold and near the minimum possible (above freezing) temperature for all three experiments. The freshening in DP_{clsd} ASYMM compared to the other two experiments arises from the increased atmospheric moisture transport to the subpolar regions in the SH associated with the ASYMM moisture diffusion field. The DP_{open} ASYMM case is also subjected to this enhanced moisture transport, but in that experiment the isolation of Antarctica beyond the ACC and south of the wind-driven upwelling regions of saline Circumpolar Deep Water (CDW) results in a reduced impact of this freshening at high latitudes. Furthermore, unlike in DP_{clsd} ASYMM, AABW is now denser than NADW when the DP is open. This is in agreement with the present-day real ocean, where AABW constitutes the densest and therefore deepest-penetrating water mass. However, this does not preclude the formation of NADW in this experiment (DP_{open}), as here the strength and polarity of the MOC depends on the density difference between the AAIW and NADW formation regions because of the existence of the DP gap (Sijp and England 2005). In contrast to the ASYMM experiments, the NADW formation region is significantly fresher and cooler in DP_{clsd} CNTRL. This is because of the absence of NADW formation, leading to a cooling and freshening of the North Atlantic catchment area. Again, the AABW formation region is more dense than the NADW formation region in DP_{clsd} CNTRL. This is in agreement with the SH sinking state found in this experiment and previous studies that ignored the possible effects of asymmetric atmospheric moisture transport.

4. Climatic consequences

The opening of the DP is thought to have cooled Antarctic climate (Kennett 1977), and modeling studies (e.g., Sijp and England 2004) lend some support to this thesis. Figure 5 shows the annual average SST cooling arising from opening the DP (i.e., the SST difference for DP_{open} − DP_{clsd}) for the CNTRL and ASYMM diffusion fields. The CNTRL experiment shows good agreement with Sijp and England (2004), with a maximum SH cooling of up to 8°C in the Atlantic sector of the SO north of 60°S (Fig. 5a). Opening the DP leads to a warming in the North Atlantic that results from the establishment of NADW formation in DP_{open} CNTRL, a feature absent when the DP is closed in that experiment. In contrast, the opening of the DP under the ASYMM moisture diffusion field leads to significantly less cooling (Fig. 5b) compared to the CNTRL experiment, with only up to around 2°C cooling in the South.
Atlantic and up to around 3°C cooling in the Indian Ocean sectors of the SO. This is because of the absence of a strong SH overturning cell in DPclsd when moisture diffusion is increased to more realistic values over the Southern Ocean. As such, there is no significant response in the SH THC to the opening of the DP in ASYMM, and therefore no strong cooling occurs. Warming in the North Atlantic is also significantly less pronounced than in the CNTRL experiments resulting from the existence of NADW formation in both ASYMM experiments, regardless of whether the DP is open or closed.

Figure 6 shows the annual average barotropic streamfunction for DPclsd ASYMM, DPclsd CNTRL, and difference in streamfunction for DPclsd ASYMM – CNTRL. Under DP closure under the ASYMM moisture diffusion case, there is a vigorous 48-Sv subpolar gyre spanning the Atlantic and Indian Ocean sector of the SO, a 34-Sv gyre in the South Pacific, and a 34-Sv Brazil Current (Fig. 6b and Table 1). This value for the Brazil Current is close to that in the DPopen ASYMM case (29.5 Sv; Table 1). In contrast, the subpolar gyres are significantly reduced in strength and spatial extent in DPclsd CNTRL. Also, the southward-flowing Brazil Current is now increased to 66.9 Sv. This shows that the hydrological cycle can have a significant effect on the SH subpolar gyres and the strength of the Brazil Current when the DP is closed. A more modest impact is seen when the DP is open (Table 1). This is because in DPclsd CNTRL the strong SH deep-water formation, and the associated increased northward deep western boundary current, introduces an enhancement of the upper-western boundary current as a result of vorticity dynamics in a similar fashion to the enhancement of the Atlantic subtropical gyre resulting from NADW formation. This results in an enhancement of the Brazil Current and a reduction in strength of the subpolar gyres. This thermohaline strengthening of the South Atlantic subtropical gyre is absent when SH moisture diffusion is enhanced, because of the absence of vigorous AABW formation in this experiment. Comparison of Figs. 6b and 6c reveals a significant control of the SH depth integrated streamfunction by the rate of moisture diffusion in the atmosphere when DP is closed.

Figure 7 shows the northward heat transport and the zonally averaged surface air temperature (SAT) change in response to the opening of the DP for the CNTRL and ASYMM experiments. Opening the DP in the CNTRL experiments leads to a significant increase in northward heat transport in the NH associated with the onset of NADW formation (Fig. 7a). Decreased southward heat transport in the SH reflects the reduction in SH sinking and the associated reduction in the southward flow anomaly. This result is in agreement with previous modeling studies (e.g., Sijp and England 2004). In contrast, when moisture diffusion varies as in ASYMM, the northward heat transport is less sensitive to the opening of the DP.
This is because there are only modest changes in NADW formation (around 3-Sv increase) when DP is opened under the ASYMM moisture diffusion scenario. Poleward heat transport in the SH is only slightly decreased in this situation, and is mainly caused by the increase in NADW formation. The reduced poleward heat transport in the SH upon opening the DP in the CNTRL experiments leads to a maximum of 2.5°C SAT cooling in the SH zonal mean (Fig. 7b and Table 1). This is because of the reduced poleward HT upon opening the DP, whereas the NH warms by a maximum of 2.7°C as a result of the onset of NADW formation (Fig. 7b). A similar result is obtained when moisture diffusion is enhanced across both hemispheres’ subpolar regions (the SYMM experiments), where an SH cooling of 2.7°C is obtained (figure not shown; see Table 1). In contrast, only up to 0.5°C zonal-mean cooling occurs upon opening the DP in the ASYMM experiments, with a 1.1°C warming resulting from the enhancement of NADW formation. The ASYMM2/3 experiments yield 2.6°C cooling (figure not shown; see Table 1) because of the vigorous SH overturning in the DPclsd ASYMM2/3 experiment. In summary, the DP results from previous studies are fundamentally altered when moisture diffusion is set to values appropriate to the present day (i.e., varying significantly between the Southern and Northern Hemispheres). Symmetric or weakly asymmetric atmospheric moisture diffusion fluxes yield unchanged results.

Our experiments have been carried out in a cold climate, whereas considerably warmer ocean temperatures prevailed during the Eocene. Warmer ocean temperatures may affect ocean circulation sensitivity to surface salinity differences (de Boer et al. 2008), as temperature depends less on salinity as temperature increases. Therefore, we now examine the possible influence of this effect in our experiments. Following de Boer et al. (2008), we have replicated our experiments while increasing the apparent temperature of the ocean by 12°C at each instance where density is calculated in the model. We follow this idealized procedure rather than changing the radiative balance in the model because of difficulties in raising polar SST by 10°–12°C without using extremely high CO₂ concentrations.

Figure 8 shows the steady-state global annual-mean meridional overturning streamfunction for the +12°C
experiments where ocean density is modified. The overturning in the DPclsd and DPopen experiments in the CNTRL case remains largely unaltered. One notable exception is the appearance of an 11.6-Sv North Pacific cell in the DPopen experiment (figure not shown). This is because of the reduced influence of salinity on density at higher temperatures (see also de Boer et al. 2008), whereby the halocline in the North Pacific no longer inhibits wintertime sinking of cold surface water. The large SH cell remains the sole overturning agent in the DPclsd experiment, and this cell increases in strength to 47.0 Sv. The overturning remains similar also for DPopen ASYMM, again with 12.6-Sv sinking emerging in the NP. In contrast, overturning now occurs in both hemispheres in DPclsd ASYMM, whereby the NH sinking cell is accompanied by 34.5 Sv of AABW formation. As per the North Pacific, the Antarctic halocline no longer ensures stable vertical stratification there as a result of the reduced sensitivity of density to salinity at higher temperatures. The 26.7-Sv NH sinking (18.1 Sv in the NA and 8.6 Sv in the NP) in the +12°C DPclsd ASYMM experiment stands in stark contrast to the absence of NH sinking in the +12°C DPclsd CNTRL experiment (cf. Figs. 8d,b). The southward atmospheric moisture transport in DPclsd ASYMM allows the NH ocean to remain dense enough for sinking to occur, whereas the symmetrical moisture transport in DPclsd CNTRL renders the NH ocean too fresh to allow deep sinking.

Similar to the original experiments, the zonally averaged SST change in response to opening DP is significantly greater in the +12°C CNTRL case than in the +12°C ASYMM case, as shown in Fig. 9. This is because unlike DPclsd CNTRL, SH sinking occurs in conjunction with NH sinking in DPclsd ASYMM and is weaker than in

Fig. 6. Yearly averaged barotropic streamfunction (Sv) for (a) DPclsd ASYMM, (b) DPclsd CNTRL, and (c) difference in streamfunction for DPclsd ASYMM – DPclsd CNTRL.
Nonetheless, a weak temperature response occurs in the modified DPcld ASYMM experiment. In conclusion, the symmetry of atmospheric moisture transport clearly remains influential on the ocean’s MOC even in a very warm climate. The notable difference with a cooler climate here is that the exclusive NH sinking we found originally in DPcld ASYMM is now accompanied by strong SH sinking in a “warm climate.” It should be noted that raising the apparent dynamic ocean temperature by 12°C leaves the meridional density gradients unchanged. In contrast, reconstructions of Eocene ocean temperature gradients suggest significantly reduced temperature gradients. This means that our warm climate experiments may overestimate the relative importance of temperature gradients over salinity gradients relative to the actual Eocene, and in particular that our warm climate experiment for DPcld ASYMM may overestimate the role of SH sinking.

5. Conclusions

The absence of the ACC resulting from the closure of the DP ocean gateway in ocean-only and simple coupled models generally leads to vigorous AABW formation (e.g., Mikolajewicz et al. 1993; Toggweiler and Bjornsson 2000; Nong et al. 2000; Sijp and England 2004, 2005), Antarctic air temperature warming, and the absence of NADW formation. Similar studies by Nong et al. (2000) and Najjar et al. (2002) find NH overturning in conjunction with strong AABW. However, bipolar sinking is only maintained as a result of their use of unrealistic restoring conditions for sea surface salinity (SSS), and this bipolar solution would likely be eliminated in their model had they used air–sea freshwater fluxes or otherwise more realistic nonrestoring salinity conditions [see Sijp and England (2004) for a discussion]. Here we have shown that this result is critically dependent on atmospheric moisture transport by midlatitude storms. A sufficiently symmetric atmospheric moisture transport field across the two hemispheres leads to the establishment of the large SH overturning cell found in previous studies when the DP is closed. In contrast, when SH atmospheric moisture transport is set to values appropriate to the present day—namely, with sufficiently stronger SH moisture diffusion compared to the NH—the model admits an NH overturning state even when the DP is closed. In this case, no steady vigorous SH overturning state exists, and the Antarctic SST and SAT response to the opening of the DP is much reduced. Atmospheric moisture diffusion in the asymmetric ASYMM case leads to improved simulations for the present-day climate in the model we employ (Saenko et al. 2003), and may also apply to climates where the DP is closed. We therefore conclude that the formation of NADW, or its precursor, may have occurred before the opening of the DP. Its existence depends on an interplay between the DP effect and the global hydrological cycle. In climates where atmospheric moisture transport caused by midlatitude storms is more or less symmetric about the equator (the CNTRL and SYMM experiments), or only weakly asymmetric (the ASYMM2/3 case), no Northern Hemisphere overturning state develops, and a strong SH cell dominates the global THC when DP is closed. This leads to a stronger Antarctic SST and SAT response. Our warm climate experiments wherein temperature is increased by 12°C in the calculation of density yield similar results to our original findings, with the notable exception that SH sinking now occurs in conjunction with NH sinking in DPcld ASYMM. However, as in the original experiments, the climatic response to opening the DP remains significantly larger in the CNTRL case than in the ASYMM case for these +12°C experiments. This demonstrates the robust result that meridional moisture fluxes in the atmosphere can affect the bipolar density seesaw, and in turn the climatic response to opening the DP, even in a warmer world where density depends more on temperature than salinity.

In contrast to the studies yielding strong SH sinking, Huber and Sloan (2001), Huber et al. (2004), and
Huber and Nof (2006) find a weak but present Northern Hemisphere overturning state in their fully coupled model. The atmospheric component of their model allows a relatively free evolution of the wind field and atmospheric moisture transport. They attribute the NH sinking cell to hemispheric asymmetry in surface salinity, where saline conditions in the Tethys and North Atlantic Oceans facilitate deep sinking there, while fresh conditions resulting from strong net precipitation over the SO and the NP inhibit sinking there. Our results suggest that the results of Huber and Sloan (2001) could be realized when atmospheric moisture transport by midlatitude storms favors fresh SO conditions (as in our ASYMM case), and that the results of previous studies yielding vigorous SH sinking depend on sufficient hemispheric symmetry in atmospheric moisture transport. The NH overturning found in the model of Huber and Sloan (2001) suggests that our ASYMM case may represent a simplified version of more sophisticated models of the Eocene climate.

It should be noted that the onset of NH overturning in response to the opening of SO gateways at the Eocene–Oligocene boundary may not have been restricted to the NA alone. Von der Heydt and Dijkstra (2006) find NH sinking in both the Pacific and the Atlantic in their model of the Oligocene, whereby North Pacific sinking is facilitated by westward salt transport through the Panama Seaway. They proceed to show that a later flow reversal in this equatorial gateway during the Miocene leaves the Pacific fresher, rendering NADW formation the dominant NH sinking process. This flow reversal is thought to arise from a widening of the SO gateways and narrowing of the Tethys, as also examined by Omta and Dijkstra (2003). Furthermore, these results suggest that if northern sinking occurred during the Eocene (as suggested by our ASYMM case), deep sinking was present also in the NP as a result of the presence of the Panama Seaway and the absence of a circumpolar Southern Ocean. However, Huber and Sloan (2001) find

![Fig. 8. Steady-state global annual-mean meridional overturning streamfunction (Sv) for the +12°C experiments where ocean density is modified (we add 12°C to the ocean temperature when calculating density): (a),(c) DP_{open} and (b),(d) DP_{clus} for (a),(b) the CNTRL experiments and (c),(d) the ASYMM experiments.](http://journals.ametsoc.org/jcli/article-pdf/22/9/2483/3955098/2008jcli2476_1.pdf)

![Fig. 9. The zonally averaged SST change (°C) in response to the opening of the DP for the +12°C experiments where ocean density is modified. That is, the SST difference DP_{open} - DP_{clus} is shown for the CNTRL (solid) and the ASYMM (dashed) experiments.](http://journals.ametsoc.org/jcli/article-pdf/22/9/2483/3955098/2008jcli2476_1.pdf)
no sinking in the North Pacific in their Eocene simulation because of strong precipitation there.

In conclusion, the effects of the hydrological cycle need to be considered when examining the global climate response to changes in continental geometry, particularly where a circumpolar Southern Ocean is absent. In this study, we found a greatly reduced oceanic response to the closure of the Drake Passage when moisture diffusion rates are set to values appropriate to the present-day hydrological cycle.

Acknowledgments. We thank the University of Victoria staff for support in usage of their coupled climate model. This research was supported by the Australian Research Council and Australia’s Antarctic Science Program.

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