LETTERS

The Atlantic–Pacific Seesaw

OLEG A. SAENKO

Canadian Centre for Climate Modelling and Analysis, Meteorological Service of Canada, Victoria, British Columbia, Canada

ANDREAS SCHMITTNER

Institut für Geowissenschaften, Universität Kiel, Kiel, Germany

ANDREW J. WEAVER

School of Earth and Ocean Sciences, University of Victoria, Victoria, British Columbia, Canada

4 December 2003 and 8 January 2004

ABSTRACT

A global, oceanic teleconnection of salinity, meridional overturning circulation (MOC), and climate of the North Atlantic and North Pacific is proposed. Simulations with a global climate model show that an extraction of freshwater from the Pacific results not only in an increase of salinity there, but also in a decrease of salinity in the Atlantic. As a result, a Pacific MOC develops while the Atlantic MOC collapses without freshwater perturbation in the Atlantic. Similarly, an input of freshwater to the Atlantic leads not only to a decrease of salinity there, but also to an increase of salinity in the Pacific. The Atlantic MOC collapses, whereas the Pacific MOC develops without freshwater perturbation in the Pacific. The mechanism behind this antiphase Atlantic–Pacific relationship is the positive feedback between ocean circulation and salinity contrasts, originally proposed by Stommel to operate between low and high latitudes. Here the authors show that the same mechanism operates on the Atlantic–Pacific interbasin scale, with the Southern Ocean acting as a pivot point for the interbasin seesaw. The proposed Atlantic–Pacific seesaw effect helps to explain some major out-of-phase oscillations of the climate states between the North Atlantic and North Pacific during the last deglaciation.

1. Introduction

One of the most distinctive characteristics of the present-day global ocean overturning circulation is its asymmetry between the Atlantic and Pacific. North Atlantic Deep Water (NADW) forms in the Atlantic, whereas no such deep sinking exists today in the Pacific. The resulting meridional overturning circulation (MOC) is responsible for a large fraction of the oceanic heat transport in the Atlantic, but not the Pacific. This is believed to be one of the key reasons for the observed difference between the relatively cold climate of the northwestern parts of North America and the comparatively warmer climate of northwestern Europe. The mechanisms establishing and maintaining this Atlantic–Pacific asymmetry remain controversial, although the salinity contrast between the two ocean basins appears to play an important role (Warren 1983). This is also supported by the ocean modeling studies (e.g., Hughes and Weaver 1994; Seidov and Haupt 2003).

Our understanding of multiple states of ocean circulation is based on the early work of Stommel (1961). He used a box model to show how an anomalously increased salinity contrast between tropical waters and high-latitude waters would reduce the MOC and the associated northward advection of salinity, thereby further increasing the salinity contrast. As a result of this positive feedback, when a certain threshold was exceeded, the location of deep-water formation switched from high to low latitudes. Bryan (1986) applied this feedback to explain the transitions from an equatorially symmetric circulation to an asymmetric circulation in a single-basin ocean model. He showed that deep-water formation in the Northern Hemisphere could be turned on and off by applying freshwater forcing of a corresponding sign to high latitudes of the Southern Hemisphere, that is, without any explicit forcing in the north.

Corresponding author address: Dr. Oleg A. Saenko, Canadian Centre for Climate Modelling and Analysis, Meteorological Service of Canada, University of Victoria, P.O. Box 1700, Victoria, BC V8W 2Y2, Canada.
E-mail: oleg.saenko@ec.gc.ca

Similar north–south MOC transitions have been illustrated to exist in a model of realistic continental configuration (Weaver et al. 2003; Saenko et al. 2003), with Stommel's feedback playing an important role. Here we show that this same feedback operates on an interbasin, Atlantic–Pacific scale, with the Southern Ocean, and in particular the Antarctic Intermediate Water (AAIW), acting as a pivot point for the interbasin seesaw.

2. Climate model and experimental strategy

The coupled model we use is described in detail in Weaver et al. (2001). It comprises an ocean GCM, an energy–moisture balance atmosphere model and a dynamic–thermodynamic sea ice model. All model components have the same horizontal resolution of 1.8° latitude and 3.6° longitude. The ocean model uses isopycnal mixing after Gent and McWilliams (1990); the vertical mixing scheme ensures very small values of diffusivity in the pycnocline (order of 10^{-5} m^2 s^{-1}) away from regions of rough ocean bottom topography (Simmons et al. 2004). The atmospheric model calculates surface heat and freshwater fluxes, as well as the transport of sensible heat and moisture. The annual cycle of winds is prescribed from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis.

At steady state, the model simulates about 15 Sv (1 Sv $\equiv 10^6$ m^3 s^{-1}) of deep-water formation in the North Atlantic and no deep-water formation in the North Pacific. Starting from this reference state, we conduct two sensitivity experiments, both of which illustrate the Atlantic–Pacific seesaw effect. In experiment P, freshwater was extracted from the surface of the North Pacific, whereas in experiment A, freshwater was supplied to the surface of the North Atlantic (Fig. 1a). To keep global average salinity constant, the same total freshwater is extracted (in expt A) and added (in expt P) as a uniform flux over the entire World Ocean area, apart from the perturbation regions. Hereafter we distinguish between the freshwater perturbations, applied in the localized regions of the North Atlantic and North Pacific, and the much smaller (per unit area) compensating fluxes. In both experiments, the rate of change of the freshwater forcing was slow [0.2 Sv (1000 yr)$^{-1}$] to ensure that the climate system was in near equilibrium with the forcing. The experiments were integrated for 3000 yr, decreasing the freshwater forcing in experiment P from 0 to -0.6 Sv and increasing it in experiment A from 0 to $+0.6$ Sv.

3. The Atlantic–Pacific seesaw

In both experiments, qualitatively similar out of phase behavior of the MOC between the Atlantic and the Pacific is found. Extracting freshwater from the North Pacific (expt P) makes the surface waters less buoyant there. As a result, North Pacific Deep Water (NPDW)

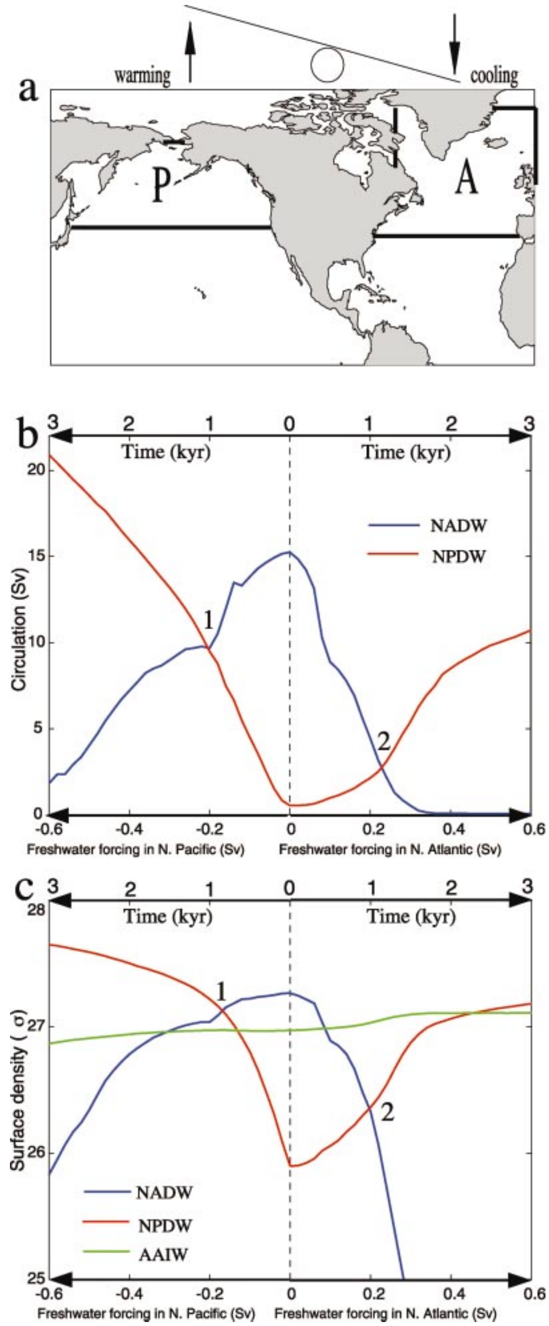

FIG. 1. (a) Schematic illustration of the Atlantic–Pacific seesaw effect and the region of freshwater extraction in expt P (marked P) and the region of freshwater discharge in expt A (marked A). (b) Maximum overturning circulation in the North Atlantic (NADW) and in the North Pacific (NPDW), and (c) surface densities in the North Atlantic (NADW; between 58° and 65°N, 0° and 40°W), in the North Pacific (NPDW between 58° and 65°N) and in the Southern Ocean (AAIW between 55° and 62°S) as functions of the freshwater extraction in the North Pacific (to the left of zero) and freshwater input in the North Atlantic (to the right of zero). The climate states labeled 1 and 2 in (b) and (c) are discussed in the text.

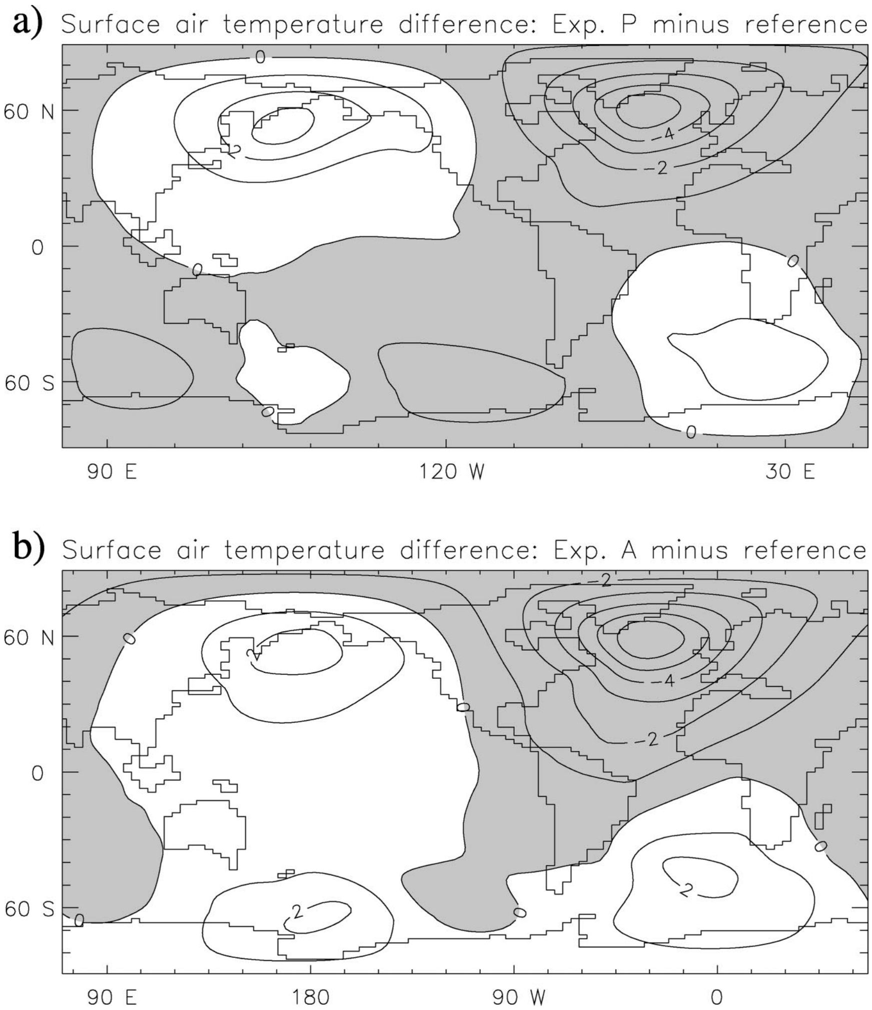

Fig. 2. Surface air temperature differences between the climate obtained (a) after extracting freshwater from the Pacific (final state of expt P) and the reference climate state and (b) between the climate obtained after discharging freshwater to the Atlantic (final state of expt A) and the reference climate state. Negative values are shaded.

begins to develop. With the freshwater forcing becoming more negative, the overturning circulation associated with NPDW gradually intensifies, whereas the circulation of NADW weakens (Fig. 1b, left part). When the rate of freshwater extraction from the North Pacific reaches -0.6 Sv, the maximum overturning in the North Atlantic drops to 2 Sv *without* freshwater perturbation in the Atlantic. Furthermore, the Bering Strait is kept closed in the model so that the only connection between the Atlantic and the Pacific is through the Southern Ocean. Explicitly supplying freshwater to the North Atlantic (expt A) dramatically reduces the Atlantic MOC, as expected. At the same time, the Pacific MOC begins to develop, *without* freshwater perturbation in the Pacific (Fig. 1b, right part).

When the freshwater forcing passes a value of ± 0.2 Sv (Fig. 1b, states ‘‘1’’ and ‘‘2’’), the maximum overturning in both oceans is about equal. However, in experiment P the Atlantic and Pacific MOCs could be said to be equally ‘‘strong’’ (Fig. 1b, state 1), whereas in experiment A they are equally ‘‘weak’’ (Fig. 1b, state 2). Insight as to why this is the case can be gained by considering the changes of water density in the region of NADW formation (ρ_{NADW}) and in the region of NPDW formation (ρ_{NPDW}), and how these compare with the density of Antarctic Intermediate Water (ρ_{AAIW}) in the Southern Ocean. The extraction of buoyancy from the North Pacific results in a rather rapid increase of surface density there, whereas the density decrease in the Atlantic is slow (Fig. 1c, left part). As a result, when the rate of freshwater extraction from the Pacific approaches -0.2 Sv, the relationship between the densities of these water masses is $\rho_{\mathrm{NADW}} \simeq \rho_{\mathrm{NPDW}} > \rho_{\mathrm{AAIW}}$ (Fig. 1c, state 1). This means, according to scaling arguments

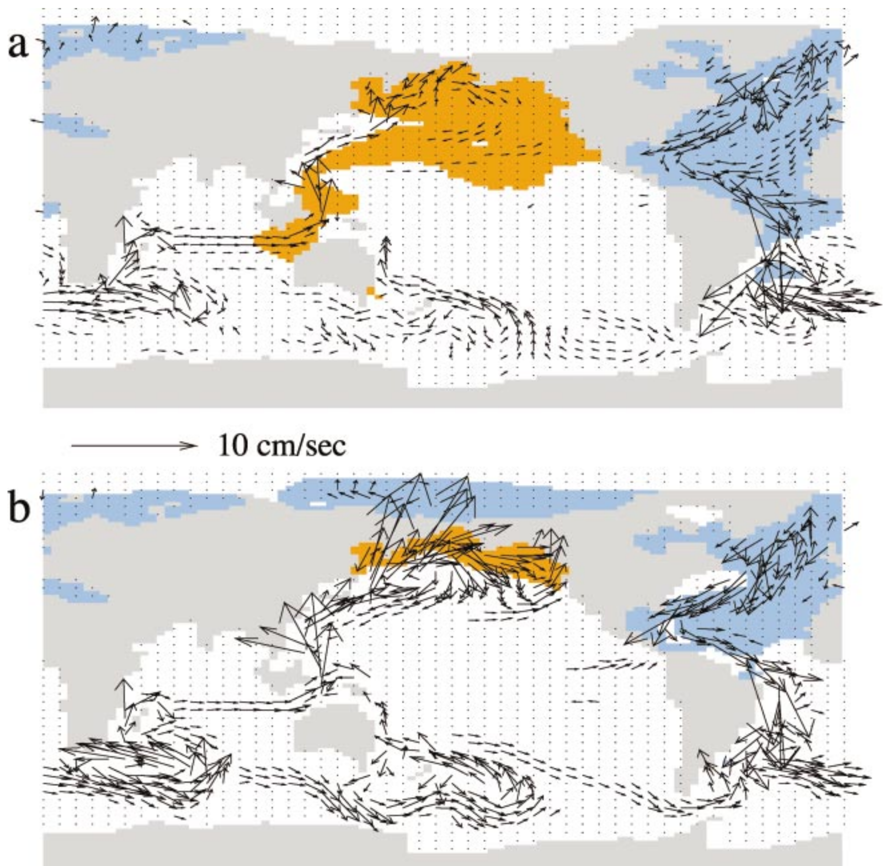

FIG. 3. Large-scale salinity and circulation anomalies in the subsurface ocean (88-m depth) (a) in the final state of expt A and (b) in the final state of expt P relative to the reference state. In (a) salinity anomalies of more than 1 g kg^{-1} are shown in orange and those of less than -3 g kg^{-1} in blue. In (b), salinity anomalies of more than 1 g kg^{-1} are shown in orange and those of less than -1 g kg^{-1} in blue. Vectors of less than 1 cm s^{-1} are not shown.

(Hughes and Weaver 1994; Gnanadesikan 1999), that both NADW and NPDW should be well developed, as indeed they are (Fig. 1b, state 1).

In contrast, supplying buoyancy explicitly to the North Atlantic in experiment A results in a rather rapid decrease of surface density there, whereas the density increase in the Pacific is slow (Fig. 1c, right part). As a result, when the rate of freshwater input to the Atlantic reaches about +0.2 Sv, the density relationship is $\rho_{\mathrm{NADW}} \simeq \rho_{\mathrm{NPDW}} < \rho_{\mathrm{AAIW}}$ (Fig. 1c, state 2). This implies that both NADW and NPDW should be in collapsed or very weak states. In our case, the maximum overturning in the Atlantic and in the Pacific at this forcing level is less than 5 Sv (Fig. 1b, state 2), again in agreement with the scaling arguments.

The northward heat transport in the two ocean basins closely follows their overturning circulations. As a result, the climate between the North Atlantic and the North Pacific flips around in both experiments, making the North Atlantic cooler and the North Pacific warmer (Fig. 2).

The physical mechanism behind this Atlantic–Pacific seesaw effect is the positive feedback between salinity anomalies and ocean circulation (Stommel 1961). A supply of freshwater to the North Atlantic (expt A) weakens deep convection and hence sinking, thereby increasing the residence time of surface waters there. This gives the forcing more time to operate, further freshening the North Atlantic and weakening the deep-water sinking. As shown by Bryan (1986), in a single equatorially symmetric and closed ocean basin this would eventually lead to a redirection of deep-water formation and heat transport to the other hemisphere, given the external constraints (such as radiation, mixing, etc.) on the net amount of deep-water production in an ocean basin.

The real ocean is not equatorially symmetric. The existence of Drake Passage inhibits a net north–south geostrophic flow in the upper ocean around 60°S down to about 2000-m depth. This, as well as the combined effect of strong surface buoyancy input and the northward Ekman flow makes the Southern Ocean a region of deep-water upwelling and transformation into a lighter intermediate water and denser bottom water (e.g.,

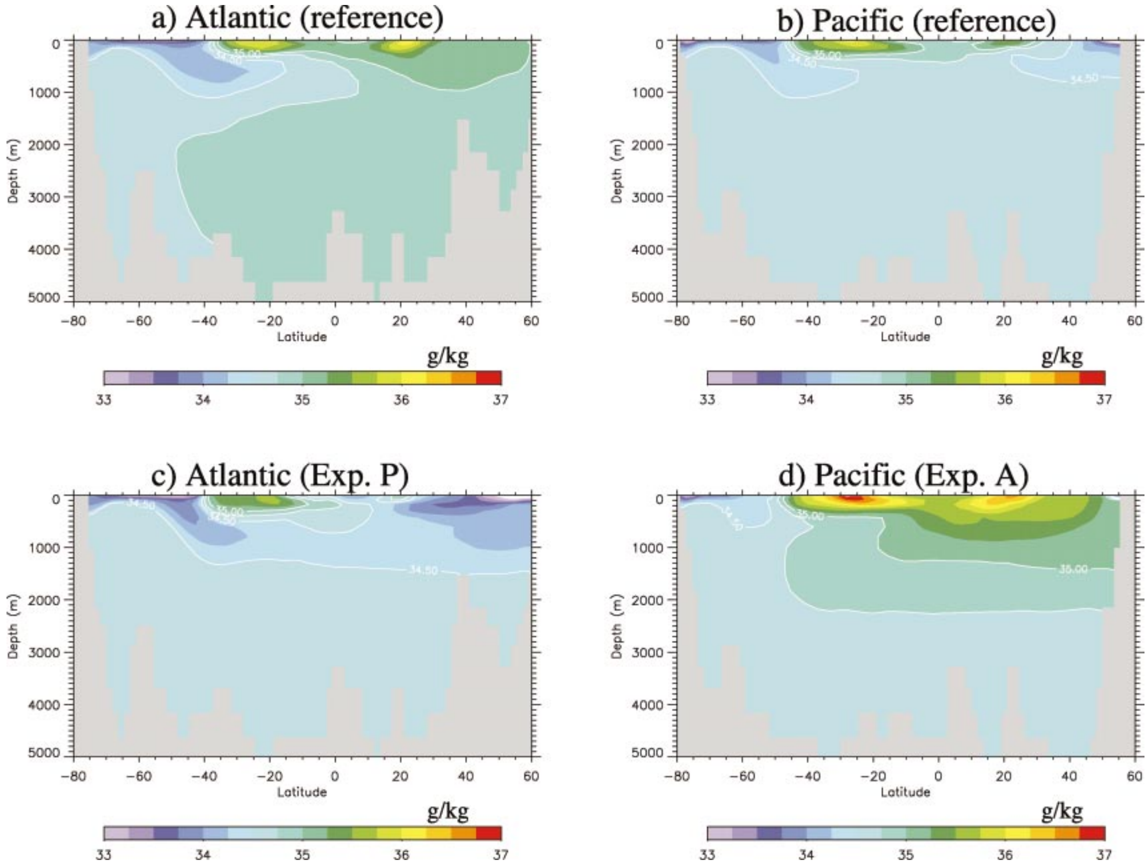

FIG. 4. Salinity sections in the (a), (c) Atlantic at 27°W and (b), (d) in the Pacific at 169°W; (a) and (b) correspond to the reference state while (c) and (d) correspond to the final states of expts P and A, respectively.

Speer et al. 2000), rather than a region of deep-water sinking. As long as the denser deep water is being transformed into the lighter intermediate water in the south, there must be a compensating transformation of light water into a dense water elsewhere, mainly in the north. As a result, when NADW formation weakens in our experiment A, the deep water (as opposed to bottom water) begins to form in the Pacific, rather than in the Southern Ocean. The formation of NPDW is fed by a reduction of the Indonesian Throughflow (by about 8 from 16 Sv in the reference state) and by a redirection of water from the Southern Ocean. The enhanced inflow of saline subtropical waters due to the developing overturning circulation makes the North Pacific much saltier (Fig. 3a). As can be seen from Figs. 1b and 1c, after the Atlantic MOC considerably weakens (state 2), it takes 1–2 kyr for this advective mechanism to spin up the Pacific MOC so that $\rho_{\mathrm{NPDW}} > \rho_{\mathrm{AAIW}}$. Eventually, the vertical salinity structure in the Pacific becomes similar to that of the Atlantic in the reference state, with fresh AAIW overlying saline NPDW (Figs. 4a,d). Note that no freshwater perturbation in the Pacific was needed to reorganize its salinity from that shown in Fig. 4b to that in Fig. 4d.

Similar arguments can be applied in the case of freshwater extraction from the North Pacific in experiment P. The explicit forcing of the NPDW formation leads to an enhanced flow of saline, low-latitude waters to the North Pacific (Fig. 3b). At the same time, the initial surface freshwater cap in the North Pacific is gradually removed by the developing deep circulation and the forcing. In an equatorially symmetric ocean, this would lead to a redirection of deep-water formation and heat transport from another hemisphere (Bryan 1986). However, deep water does not form in the Southern Ocean, so it is being redirected to the Pacific from the Atlantic. The North Pacific becomes saltier, whereas the North Atlantic becomes fresher. The vertical structure of salinity in the Atlantic becomes similar to the Pacific reference state, with fresh intermediate waters in both hemispheres propagating equatorward (Figs. 4b,c). Again, no freshwater perturbation in the Atlantic was needed to reorganize its salinity from that shown in Fig. 4a to that in Fig. 4c.

4. Implications for past climates and conclusions

Proxy data suggest that during the Younger Dryas (YD) period (~11.5–12.9 kyr BP) the Atlantic MOC was weaker than today and that the climate of the North

Atlantic was much colder. Recent reconstructions suggest that the last thousand years of the YD were accompanied by an increase of sea surface temperature (SST) in the North Pacific by as much as 1°–2°C (T. Kiefer 2003, personal communication). This was followed by a gradual decline of North Pacific SSTs by 3°–5°C to their minimum around 8.5–8.0 kyr, whereas the North Atlantic SSTs increased after the termination of the YD. This pattern of opposite Atlantic versus Pacific temperature and ventilation changes on millennial time scale apparently extends back throughout the last 60 000 yr (Kiefer et al. 2001). Previous modeling studies succeeded in producing the ventilation pattern through an atmospheric teleconnection (e.g., Schmittner and Clement 2002), although simulated SST anomalies were of the same sign in both basins. The oceanic teleconnection proposed here helps to explain the findings from the paleorecord.

We showed that deep-water formation in the Atlantic and in the Pacific are fundamentally connected through a seesawlike effect. Specifically, a weakening of the Atlantic MOC as a consequence of a large freshwater perturbation there would result in a strengthening of the Pacific MOC and an associated warming of the North Pacific, without freshwater perturbation in the Pacific. Similarly, a strengthening of the Pacific MOC would lead to a weakening of the Atlantic MOC and an associated cooling of the North Atlantic, without freshwater perturbation in the Atlantic. Taken together, and in the context of our earlier work (Weaver et al. 2003; Saenko et al. 2003) and related work by others (e.g., Seidov and Haupt 2003), it is clear that understanding the behavior of the Atlantic MOC over the last glacial cycle requires an analysis of processes operating in the Southern Ocean and the North Pacific, as well as locally in the North Atlantic itself.

Acknowledgments. We would like to thank J. Fyfe and W. Merryfield, who provided internal reviews. We also thank two anonymous reviewers for their constructive suggestions. AJW is grateful for release time provided by the Killam Foundation and the Canada Research Chair Program and for funding support from NSERC and CFCAS. AS is supported by the Deutsche Forschungs Gemeinschaft within the research unit on ocean gateways.

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