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Sea surface salinity as a key to the global ocean conveyor

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1. Introduction

Global ocean thermohaline circulation (THC) is driven by deepwater formation in the high latitudes of both hemispheres. The THC, also known as a 'salinity conveyor belt', is a truly a global system of currents inter-connecting very distant regions of the world ocean (e.g., BROECKER 1991, BROECKER & DENTON 1989, GORDON 1986, and SCHMITZ 1995).

Modern science has solved many problems of present-day climatology and ocean circulation. However, some issues are not yet fully understood. Among them is the role of observed strong asymmetry in sea surface salinity (SSS) between the Atlantic and Pacific Oceans in THC dynamics. This SSS asymmetry has been under discussion for a long time. Since late the sixties (e.g., WARREN 1983 and WEYL 1968) the mechanism that preserves these SSS contrasts continues to be disputed in literature (the most recent review and very detailed discussion including an extensive list of references on the subject is given by WEAVER ET AL. 1999). Fresher surface water in the North Pacific (NP) is also lighter than that in the North Atlantic (NA), since seawater density in the high latitudes depends on salinity variations stronger than on temperature ones (e.g., POND & PICKARD 1986). The key implication of denser NA surface water for the THC is that deepwater forms in the NA and not in the NP. The North Atlantic Deep Water (NADW), saltier and denser than the NP surface water, travels in the deep-ocean branch of the global THC to the NP and underlies the NP upper-ocean layers, thus preventing deep convection there.

Although the SSS asymmetry is considered to be the major cause and the first logical choice for the absence of deepwater formation in the NP, some argue that the difference between the NA and NP geometries is also important (see the details in the review by WEAVER ET AL. 1999). They argue that only when the surface water is carried far to the north into the semi-enclosed GIN sea by the North Atlantic Current, it cools down and becomes dense enough to sink to a substantial depth. SEIDOV & MASLIN (1999) have modeled idealized meltwater events. They have compared the role of deep convection in GINS and in the northern NA (NNA) south of Iceland for THC operation and have concluded that the portion of deepwater produced in GINS is less important for the global conveyor. Furthermore, there is a great deal of deep sinking being observed in the Labrador Sea. Thus, there is no geometrical reason why a similar deepwater formation does not occur at comparable latitudes in the NP. Thus, the lack of deepwater formation in the NP is that the vertical stability is high because of too fresh water in the northern NP.

The transport of atmospheric water vapor by the Trade Winds from the low-latitude NA over the Isthmus of Panama into the Pacific Ocean is the major cause of the build-up of NA-NP SSS contrast (e.g., BAUMGARTNER & REICHEL 1975; WARREN 1983, and WEYL 1968). An opposite balancing atmospheric water transport by the Westerlies from the Pacific into the NA is blocked by the Rocky Mountains. If only evaporation minus precipitation (E-P) were responsible, a net vapor transport of about 0.45 Sv would be necessary to maintain the observed salinity difference between the two oceans (BAUMGARTNER & REICHEL 1975, BROECKER 1989, and WEAVER ET AL. 1999).

In order to investigate the role of freshwater re-distribution between the NA and NP, we have carried out a number of simple experiments. We found that such a re-distribution is necessary, but not sufficient for driving a vigorous and truly global present-day THC. In our experiments, it appears that a freshwater surface impact in the Southern Hemisphere is needed for such a realistic ocean conveyor, if a realistic “air freshwater bridge” between the NA and NP is presumed. However, if a far higher vapor transport from the NA to NP is allowed, a rather strong conveyor can emerge, showing that the southern impact is a supporting factor, rather than a driving force by itself. As SEIDOV ET AL., 2001A, 2001B have shown, the southern impacts on the THC are caused by restraining or magnifying the NADW formation, rather than by imposing its own direct control on the THC.

2. Experimental design and model setup

We have used the Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model version 2.2 (MOM2) (PACANOWSKI 1996), with a global resolution of 6° (zonal) by 4° (meridional) and 12 vertical levels (see HAUPT ET AL. (2001) for a more detailed description). GENT & MCWILLIAMS (1990) parameterization for isopycnal mixing is used as implemented in MOM-2 (1996). This scheme is considered to be the most advanced mixing scheme in numerical ocean circulation modeling and used for removal of hydrostatic instability (ocean models without isopycnal mixing show exaggerated vertical convection, especially in the northern NA and Southern Ocean region around Antarctica; a detailed discussion is given by MCWILLIAMS (1998)). This chosen model setup is similar to those used by HAUPT ET AL. 2001 and SEIDOV & HAUPT, 1999A, 1999B. Similar models have been used numerous times for successful modeling the global thermohaline circulation (e.g., MANABE & STOUFFER 1995, RAHMSTORF 1995, SEIDOV & MASLIN 1999, TOGGWEILER ET AL. 1989, WEAVER ET AL. 1994).

The world ocean bathymetry extends in the Northern Hemisphere to 80°N and 40ºE, which excludes the Arctic Ocean but includes the Barents Sea (HAUPT ET AL. 2001). The Mediterranean Sea is also excluded. Thus, the Bering and Gibraltar Straits are closed (e.g., LARGE ET AL. 1997). In the Southern Hemisphere the ocean is bounded by the Antarctica. Our bathymetry is taken from the ETOPO5 (1986) dataset. It is quite realistic despite the coarse resolution being used here. It includes the main gateways, like the Indonesian Throughflow, and major features of the world ocean bathymetry, such as the mid-ocean ridges and sea slopes.

Here we have abandoned restoring boundary conditions employed in a series of our previous work (e.g., HAUPT ET AL. 2001, SEIDOV ET AL., 2001A, SEIDOV & HAUPT 1999A), which may be a source for controversy in a sensitivity study targeting the role of vapor flux in the THC dynamics. We use a so-called mixed boundary conditions, with restoring ocean upper temperature to specified sea surface temperature (SST), and by prescribing freshwater fluxes calculated from a present-day (control) experiment (Exp. 1, Table 1). This control run has been performed using restoring to both SST and SSS from LEVITUS ET AL. (1994) and using wind stress from HELLERMAN & ROSENSTEIN (1983) (see SEIDOV & HAUPT (1999A, 2002 for details). Freshwater fluxes across the sea surface have been diagnosed using the ocean upper layer salinity in the steady state and SSS. The use of freshwater fluxes instead of fixed SSS allows upper ocean salinity to be redistributed by the ocean circulation.

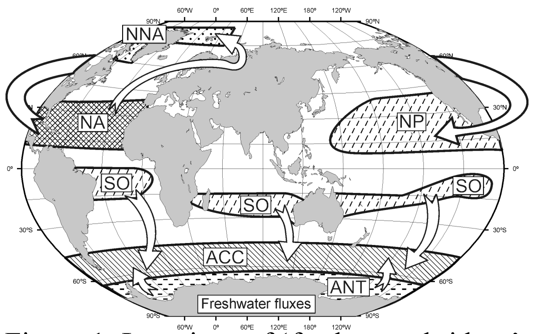

Figure 1. Locations of ‘freshwater bridges’; see Table 1 for applied freshwater fluxes

Ocean-atmosphere models undergo a rapid drift from an initial non-equilibrium condition (e.g., MANABE & STOUFFER 1994) toward the condition with total freshwater flux P-E+R (E-P plus river runoff) becoming globally balanced, i.e. vanishing when averaged over the entire world ocean. To assure that the ocean is in complete equilibrium, all runs last for 2000 ‘normal’ years for the upper ocean, and 10000 years ‘effective’ integration for the abyssal ocean due to the use of the acceleration technique provided by MOM2 (PACANOWSKI 1996). As the global average salinity we use 34.25 psu. It has also been verified that the diagnosed freshwater flux in the control run did not have any residual (there must be no such residual in a completely steady state).

In our experiments with the study of simplified moisture fluxes, we have replaced the effective fluxes by fluxes between different oceans. The following freshwater bridges have been used: (i) central NA - central NP, (ii) central NA - NNA, (iii) subtropical Southern Ocean - Antarctic Circumpolar Current (ACC) region (50º-60ºS), and (iv) ACC region - Antarctic shelf region (south of 60ºS) including Weddell and Ross Sea (see Figure 1 for locations and Table 1 for applied fluxes in Sv (1 Sv = 10^6 m^3s^{-1})). The regions have been chosen in imitation of observed E-P field and continental runoff.

Table 1 shows in addition to used 'atmospheric' and 'oceanic' freshwater fluxes the maximum northward oceanic heat flux in the NA, the maximum meridional overturning in the Atlantic and Pacific Ocean, and the inflow of Antarctic Bottom Water (AABW) into the Atlantic Ocean at 30°S.

Exp.	Freshwater flux from central NA into central NP [Sv]	Freshwater flux from central NA into NNA [Sv]	Freshwater flux from SO into Antarctica [Sv]	Freshwater flux from SO into ACC-region [Sv]	Freshwater flux from Antarctica into ACC-region [Sv]	Maximum northward oceanic heat flux in the NA [PW]	Maximum meridional overturning in the Atlantic Ocean [Sv]	Maximum meridional overturning in the NP [Sv]	AABW inflow into the Atlantic Ocean at 30°S [Sv]
1	-	-	-	-	-	0.85	20	-5	6
2	-	-	-	-	-	0.7	15	15	2
3	0.0345	-	-	-	-	0.8	20	10	4
4	0.15	0.015	-	-	-	1.5	55	-5	0
5	0.1	-	-	-	0.08	1.45	35	0	10
6	0.075	0.015	-	-	0.06	1.07	30	-5	8
7	0.05	0.0075	-	-	0.04	1.25	30	0	10
8	0.03	0.0075	0.015		0.025	1.16	30	0	5

Table 1. Freshwater fluxes between ocean regions in Sv (1 Sv = 10^6 m^3s^{-1}). Note that the area from where the water is taken becomes saltier while the second receiving region becomes fresher. Experiment 1 was driven with effective freshwater fluxes calculated from an experiment spun up with restoring boundary conditions from LEVITUS & BOYER (1994), LEVITUS ET AL. (1994), and HELLERMAN & ROSENSTEIN (1983) (see text for details). Also shown are the maximum northward heat flux in PW (1 PW = 10^{15}W) in the Atlantic Ocean and the meridional overturning in Sv for the Atlantic Ocean and NP. Note that the North Pacific's meridional overturning is calculated as the difference between global overturning and overturning in the NA.

3. Results and Discussion

To be concise, we show only the meridional overturning stream function between 30°S and the the northern boundary in the Atlantic Ocean. Meridional overturning is considered to be the most indicative THC parameter and descriptor of the global conveyor. The global THC functionality depends crucially on the NADW outflow from the Atlantic Ocean (e.g., BROECKER 1991, BROECKER ET AL. 1998, GORDON 1986, 2001, RINTOUL ET AL. 2001, SCHMITTNER ET AL. 2002, SCHMITZ 1995, SEIDOV & HAUPT 1999A, AND STOCKER & BROECKER 1992). Figure 2 shows meridional overturning and northward heat transport in the Atlantic Ocean the control run (Exp. 1, Table 1). All quantities like heat transport, meridional overturning, temperatures, salinities, and velocities agree well with analogous THC simulations (see a discussion in SEIDOV & HAUPT 1999A).

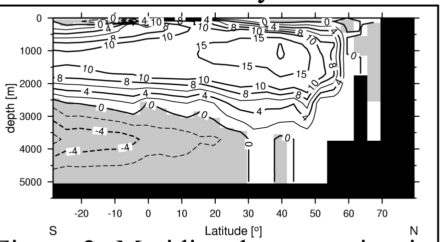

Figure 2. Meridional overturning in the Atlantic Ocean, both from Exp. 1 (see Table 1). Streamfunction is shown in Sv (1 Sv = 10^6 m^3s^{-1}).

Exp. 2 illustrates a response of the ocean to the shutting off of any freshwater exchange between the atmosphere and the ocean. In this experiment, salinity in the ocean and SSS are set to a constant (here 34.25 psu; see above) everywhere. Thus, the ocean circulation is only thermally- and wind-driven. Meridional overturning in the Atlantic Ocean slows down and no NADW is exported into the ACC region (Figure 3a) accompanied by a reduced northward heat flux in the North Atlantic Ocean (see Exp. 2, Table 1). Furthermore, the neglect of the low salinity signal in the NP allows convection in the northern North Pacific (see Table 1).

As soon as a small freshwater flux from the NA into the NP is applied to the conditions in Exp.2, the thermohaline circulation speeds up (Exp. 3, Figure 3b). However, the overall THC in this run is far from satisfactory. Although the "NA-NP air freshwater bridge" is very small, as little as 0.0345 Sv, i.e., ten times smaller than 0.45 Sv suggested by BAUMGARTNER & REICHEL (1975), BROECKER (1989), and WEAVER ET AL. (1999), the North Atlantic Current transports already too much salt from the central NA into the Greenland-Norwegian (GIN) Seas, where surface salinity has reached 35 psu. Diverting some of the moisture surplus of the water vapor from the NA (GORDON & FINE 1996) into the Arctic Ocean, as in Exp. 4, has helped to reduce surface salinity in the GIN Sea to more realistic values. This reduction occurred despite a more realistic freshwater bridge between NA and NP has been used in Exp.4 (0.15 Sv, which is only three time lower than the estimated 0.45 Sv). However, for such increase of the re-distribution of freshwater between the NA and NP led to an overwhelmingly strong conveyor (Figure 3c), which, in turn, led to a too warm abyssal ocean. Thus, the atmospheric freshwater bridge between NA and NNA alone causes too strong overturning, and the THC driven only by the NA-NP SSS contrasts needs a kind of "brakes" to slow down to a realistic present-day operation mode that would keep the deep ocean cooler than 4°C below 3 km.

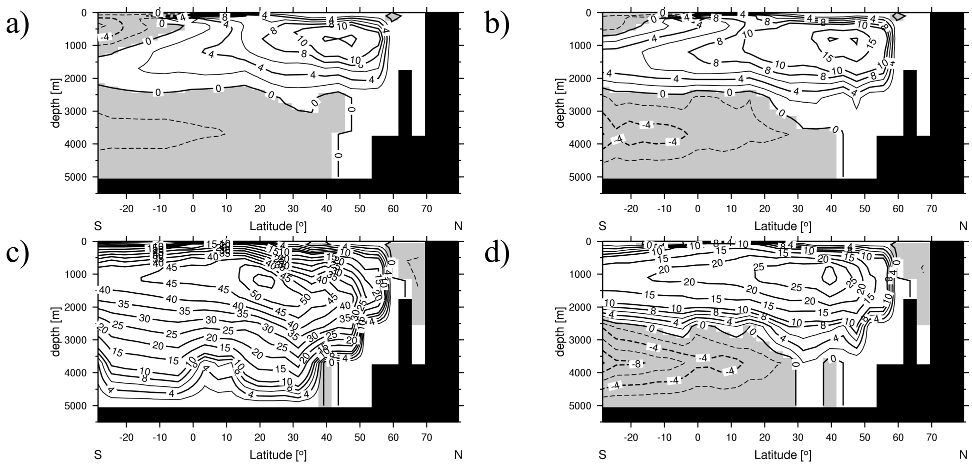

Figure 3. Meridional overturning in the Atlantic Ocean for (a) Exp. 2, (b) Exp. 3, (c) Exp. 4, and Exp. 8 (see Table 1 for details). Streamfunction is shown in Sv (1 Sv = 10^6 m^3s^{-1}).

Since the only concurrent deepwater mass that can limit the NADW dominance is the Antarctic Bottom Water (AABW), we presume that an additional "freshwater bridge" must exist in the southern high latitudes to put some brakes on the NADW speed-up. The logic suggests that some amount of freshwater has to be removed around the Antarctica to increase AABW formation. The interplay of NADW and AABW, known also as an "ocean seesaw" is essential for the THC dynamics (e.g., BROECKER 2000 and SEIDOV & MASLIN 2001). However, the freshwater re-distribution here is mostly oceanic (caused by brine rejection due to seawater freezing), rather than atmospheric, as in the Atlantic-Pacific teleconnections. In Exp. 6-8 we freshened either the band from 50-60°S (ACC area in Figure 1) or a 4° wide band around Antarctica including the Weddell and Ross Seas, respectively. The most realistic results were received in Exp. 8 where we freshened the ACC area removing freshwater from the Antarctic shelf area to mimic sea ice formation and from the subtropical Southern Ocean (SO area in Figure 1) as water vapor. The chosen combination of freshwater fluxes, which mimics only part of the global water cycle, reproduces the NADW-AABW water mass structure in the Atlantic Ocean as well the temperature and salinity distribution of the global ocean simulated in our control experiment Exp. 1 (Figure 3, compare with Figure 2).

Despite the fact that our chosen freshwater fluxes are small and not very realistic, depicting only the most prominent features of real-world hydrological cycle, it has been possible to mimic a THC very close to those in the control run, provided both NA-NP and the southern "freshwater bridges" have been effected. However, even with the southern component of hydrological partially included, the overturning rates and heat flux (both in the Atlantic Ocean and global) are higher than in reality. Further investigation will show to which extent the hydrological cycle can be simplified and which branches of freshwater cycling cannot be neglected in sensitivity ocean climate studies.

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