Removing the North Pacific halocline: Effects on global climate, ocean circulation and the carbon cycle

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Abstract

A well-pronounced halocline is a key feature of today’s subarctic North Pacific. There is indirect paleo-evidences from the last glacial termination as well as from the early and middle Pliocene that this halocline has not always been there. To study the effects of North Pacific salinity on global climate, ocean circulation and the marine carbon cycle, we perform idealized experiments using an Earth system model that is made of intermediate complexity (LOVECLIM). Imposing a negative freshwater flux in the northern North Pacific, the halocline vanishes and a deep Pacific meridional overturning circulation (PMOC) establishes. The associated increase of meridional heat transport in the Pacific leads to a bipolar seesaw response in temperature, with warming in the North Pacific and over North America and cooling in the Southern Ocean. As a result of the formation of North Pacific deep water (NPDW), the surface branch of the global conveyor belt circulation weakens. Transport through the Indonesian Seas decreases by 50% as the warm and saline waters of the equatorial Pacific are diverted into the North Pacific.

In our idealized experiments, the enhanced global deep water formation is balanced by an increase in diapycnal mixing. As a result nutrient concentrations in the euphotic zone increase by about 25% globally, leading to a 20% increase in global export production. The effect of greater export production on atmospheric pCO$_2$ is, however, compensated by the enhanced transport of dissolved inorganic carbon (DIC) to the surface. As a result, the atmospheric CO$_2$ concentration increases by only 5 ppmv. Our results further suggest that the absence of the subarctic halocline for instance during Heinrich event 1 and the Pliocene may have exerted a strong influence on global climate and the carbon cycle.

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1. Introduction

Under present-day conditions, deep water is formed in the North Atlantic and not in the North Pacific. In the North Pacific surface salinity is relatively low (~ 32.8 psu) compared to that of the underlying deep water (~ 34.6 psu) (Warren, 1983), thereby preventing the formation of deep waters. The presence of a well-pronounced subarctic North Pacific halocline (NPH) is controlled by a number of different processes (see Emile-Geay et al., 2003; Kiefer, 2010 for an overview):

- the atmospheric export of moisture from the Atlantic to the Pacific by trade winds crossing Central America,
- a limited meridional cross-gyre flow of salty subtropical waters and
- the positive freshwater balance associated with the East Asian summer monsoon.

One key question to address is what long-term processes control the strength of the NPH. During the early to middle Pliocene for example, when the Panama Seaway was still open and allowed for the exchange of surface waters and thermocline waters between the Atlantic and the Pacific, paleo-data reveal a strong level of surface productivity in the subarctic North Pacific which has been interpreted (Haug et al., 1999) in terms of the absence of a NPH. This result was corroborated by a modeling study (Motoi et al., 2005) in which an open Panama Isthmus caused an increase of surface salinity in the North Pacific and even led to the formation of a deep Pacific meridional overturning cell. Motoi et al. (2005) furthermore argued that this may have further delayed the build-up of Northern Hemispheric ice-sheets during the Pliocene. Similar conclusions were drawn by Haug et al. (2005).

Palo-data and recent modeling experiments (Okazaki et al., 2010) document that also during the last glacial termination the
NPH may have disappeared twice—during Heinrich event 1 and the Younger Dryas Period. While during these periods the Atlantic meridional overturning circulation (AMOC) was substantially weaker than at present (McManus et al., 2004), there is indirect paleo-evidence for enhanced deep water ventilation in the North Pacific (Ahagon et al., 2003; Ohkushi et al., 2004; Sagawa and Ikehara, 2008; Okazaki et al., 2010) that may have been caused by increased surface salinities in the Bering Sea and the Gulf of Alaska. This scenario is also consistent with previous modeling studies (Mikolajewicz et al., 1997; Saenko et al., 2004; Krebs and Timmermann, 2007; Okazaki et al., 2010). The development of a deep Pacific meridional overturning circulation (PMOC) in these model solutions increases the poleward heat transport in the North Pacific, which leads to a warming of subpolar areas and parts of North America.

Influences of such dramatic reorganizations of the global conveyor belt circulation on the carbon cycle and resulting potential feedbacks between North Pacific climate and marine biogeochemical changes have, to our knowledge, not been studied in great detail yet.

The objectives of this study are to study the impacts of removing the NPH on ocean circulation, global climate and the carbon cycle. For that purpose, we perform idealized experiments using an Earth system model of intermediate complexity (EMIC), LOVECLIM version 1.1. (Goosse et al., 2010), in which we artificially impose negative freshwater fluxes in the northern North Pacific. In Section 2, we describe the experimental setup of our idealized modeling experiments in more detail. Section 3, discusses the climate and carbon cycle responses to the establishment of the PMOC. Finally Section 4 presents a summary and discussion of the main results.

2. Model and experimental setup

To study the impact of North Pacific salinity changes on the climate and the carbon cycle system, we perform idealized sensitivity experiments using the EMIC LOVECLIM, version 1.1. LOVECLIM is faster than GCMs because its atmospheric component is simplified and coarser than other GCMs. As a result, longer experiments can be performed with a full coupling between all the components of the climate and carbon cycle systems.

2.1. The Earth system model LOVECLIM

The atmospheric component of the coupled model LOVECLIM is ECBilt (Opsteegh et al., 1998), a spectral T21, three-level model, based on quasi-geostrophic equations extended by estimates of ageostrophic terms (Lim et al., 1991). The model contains a full hydrological cycle which is closed over land by a bucket model for soil moisture and a runoff scheme. Synoptic variability associated with weather patterns is explicitly computed. Diabatic heating due to radiative fluxes, the release of latent heat and the exchange of sensible heat with the surface are parametrized.

The sea–ice–ocean component of LOVECLIM, CLIO (Goosse et al., 1999; Goosse and Fichefet, 1999; Campin and Goosse, 1999) consists of a free-surface primitive equation model with 3° × 3° resolution coupled to a thermodynamic-dynamic sea ice model. Coupling between atmosphere and ocean is done via the exchange of freshwater and heat fluxes, rather than by virtual salt fluxes. To avoid a singularity at the north pole, the oceanic component makes use of two subgrids: The first one is based on classic longitude and latitude coordinates and covers the whole ocean except for the North Atlantic and Arctic Ocean. These are covered by the second spherical subgrid, which is rotated and has its poles at the equator in the Pacific (111°W) and Indian Ocean (60°E).

The terrestrial vegetation module of LOVECLIM, VECODE (Brovkin et al., 1997), computes once a year the evolution of the vegetation cover based on annual mean values of several climatic variables. The vegetation cover is described as a fractional distribution of desert, tree, and grass in each land grid cell. Within the version of LOVECLIM used here, simulated vegetation changes affect only the land-surface albedo, and have no influence on other processes such as evapo-transpiration.

LOCH is a three-dimensional global model of the oceanic carbon cycle simulating dissolved inorganic carbon (DIC), total alkalinity, 14C, phosphates (PO4^3−), organic products, oxygen and silica (Mouchet and Francois, 1996; Fichefet et al., 2007). LOCH is coupled to CLIO, with the same time step. In addition to their biogeochemical transformations, tracers in LOCH experience the advection and diffusion predicted by CLIO. The uptake of CO2 by the ocean is governed by the solubility and biological pumps. The partial pressure of CO2 in the surface waters is calculated from the total alkalinity and DIC tracers. The difference between the partial pressure of CO2 in the ocean and in the atmosphere, modulated by an exchange coefficient sets the CO2 air–sea exchange rate. LOCH computes the export production from the fate of a phytoplankton pool in the euphotic zone (0–120 m). The phytoplankton growth depends on the availability of nutrients (PO4^3−) and light, with a weak temperature dependence. A grazing process together with natural mortality limit the biomass of the primary producers and provide the source term for the organic matter sinking to depth. Remineralization of organic matter depends on oxygen availability, but anoxic remineralization can also occur. Depending on the silica availability, phytoplankton growth is accompanied by the formation of opal or CaCO3 (calcite and aragonite) shells, which then sink to depth. CaCO3 shells are dissolved depending on the calcite and aragonite saturation states, whereas a simple constant rate is used for opal. The organic matter that is not remineralized and the shells that are not dissolved are permanently preserved in the sediments. This leads to a loss of alkalinity, carbon, phosphates and silica, which is compensated by the river influx. The 14C content at any point in the ocean is expressed as a 14C/12C ratio. The production rate of 14C is held constant. The atmospheric CO2 content is predicted for each ocean timestep from the air–sea CO2 fluxes calculated by LOCH as well as from the air–terrestrial biomass CO2 fluxes provided by VECODE.

2.2. Experimental setup

The pre-industrial steady state (PIN) was obtained by forcing LOVECLIM with 278 ppmv of atmospheric CO2 for 500 years, then allowing the atmospheric CO2 to vary freely for 700 years (Menviel et al., 2008). The model is then forced with an atmospheric 14C content of 0 permil for about 10,000 years, after which it is allowed to vary freely. All sensitivity experiments start from this equilibrated simulation.

To remove the NPH and initiate the formation of North Pacific deep water (NPDW) we extract freshwater in the area 150°–230°E, 47°–63°N (Freshwater Experiment FE) under pre-industrial conditions and starting from experiment PIN. The overall duration of the applied forcing is 2000 years. The magnitude of the negative freshwater forcing increases linearly during the first 1000 years to −0.3 Sv [1 Sv = 10^6 m^3/s], a linear decrease at a similar rate is applied for the following 1000 years to return to the control background conditions (Fig. 1). Experiment FE is highly idealized and the negative freshwater forcing is chosen as a way to mimic paleo situations in which the NPH was weak or absent, as was hypothesized for Heinrich event 1 or the early Pliocene. The length of the forcing (2000 years) was chosen to
mimic the duration of Heinrich event 1, a time with probably higher salinities in the subpolar North Pacific (Okazaki et al., 2010). The amplitude of the forcing is set up such as to remove the haloline and generate a substantial PMOC that extends into deep waters. In experiment FE, we close the Bering Strait to prevent sea-level gradient-driven freshwater exchanges between the Arctic Ocean and the North Pacific (Hu and Meehl, 2005). Experiment FE was performed without global salinity compensation to avoid a biased temperature response in the Southern Ocean (Stocker et al., 2007).

3. Results

3.1. Oceanic response

In response to the anomalous freshwater extraction in the northern North Pacific, the sea surface salinity in the area 150−230° E, 50−60° N increases by up to 2 after 1000 years (Fig. 2) which is enough to remove the NPH. This positive salinity anomaly causes an increase in surface water density in the North Pacific by about 1.2 kg/m³. This positive density anomaly leads to the gradual development of a deep overturning circulation in the North Pacific (Fig. 1), as well as a slight weakening of the AMOC (Fig. 3, upper panels). After 1000 years, the maximum overturning strength in the Pacific (as measured by the maximum of the meridional streamfunction in the North Pacific) attains values of up to 30 Sv and the newly formed NPWD reaches a depth of 4000 m (Fig. 3, bottom right panel). A cross-equatorial transport of about 10 Sv establishes which leads to a reduction of the volume of Antarctic bottom water (AABW) in the Pacific.

As illustrated by the winter mixed layer depth anomalies (Fig. 4, right panel versus left panel) the major site of NPWD formation is the western part of the Bering Sea, close to the coast of Kamchatka, but intensified mixing also occurs elsewhere in the subarctic North Pacific. As a result of changes in the overturning circulation the model simulates an increased poleward transport of heat by 0.5 PW which is accompanied by a large-scale warming in the North Pacific reaching maximum values of up to 7 °C in the Sea of Okhotsk and by 2.5 °C, when averaged over 150−230° E, 50−60° N (Fig. 2). The warmer conditions at the ocean surface can be related to a strong poleward shift of the Kuroshio extension region (Fig. 4, right panel) creating conditions that are somewhat analogous to the Gulf-Stream-North Atlantic Drift configuration under present-day conditions. Moreover, the stronger meridional component of the Kuroshio extending into subpolar areas advects more saline subtropical surface waters into the sinking regions, thereby providing a positive feedback (Saenko et al., 2004; Okazaki et al., 2010). The upper ocean (1000 m) poleward salinity transport at 35° N increases by ~200% compared to PIN.

However, as soon as the imposed extraction of freshwater ceases (after model year 2000) NPWD formation stops. The inflow of warm water to the North Pacific does not advect enough saline waters to maintain high surface density. This suggests further that in our model solution the PMOC does not exhibit any considerable hysteresis or saddle node bifurcation behavior. As can be seen in Fig. 5 (left panel), the NPWD that is being formed in the Bering Sea is relatively warm, whereas the southern hemisphere waters cool down to 2000 m. Averaged over the entire Pacific, we observe a reduction in intermediate to deep level stratification in the Pacific Ocean, in particular between 40° S and 10° N (Fig. 5, right panel). Since surface winds do not change dramatically in experiment FE, the globally increased formation of deep water has to be balanced by enhanced diapycnal mixing. As will be shown below this is essential for the global biogeochemical response in experiment FE.

Due to the formation of NPWD, the flow of surface waters in the western tropical Pacific is diverted more towards the North Pacific, feeding the Kuroshio rather than the Indonesian throughflow. This leads to a 50% decrease in the transport through the Indonesian passages. A weakening of the Agulhas Current, the North Brazil Current as well as the gulf stream are also simulated (Fig. 4). The formation of NPWD basically weakens the whole surface branch of the global conveyor belt circulation (Fig. 4, right panel). This relates also to a reduction of North Atlantic deep water formation by 22% (from 27 to 21 Sv) (Fig. 3). In the Southern Ocean, the zonally integrated overturning circulation in the bottom cell increases by about 30% (from 17 to 22 Sv, not shown).

3.2. Atmospheric response

Due to the simulated North Pacific warming and reduced sea ice coverage, the surface air temperature in the subpolar Pacific and over northern North America increases by about 2 °C (Fig. 6). This warming pattern causes a reduction of the Subtropical High and an overall weakening of the tropical trade winds. Anomalously advected warm air over North America is likely to contribute to the warming over Canada. We also observe a reduction of the Westerlies in the Atlantic and an intensification of the Westerlies over the Southern Ocean. The latter may be explained by an increased meridional sea surface temperature (SST) gradient in this region (Fig. 2, right panel), associated with a −5 °C cooling equatorward of the Southern Ocean sea ice margin. The latter can be explained in terms of the heat piracy argument (Seidov and Maslin, 2001). The meridional gradient of equatorial Pacific SSTs and the resulting diabatic forcing (as characterized by the precipitation anomalies in Fig. 6) leads to an intensification of the southeasterly trade winds and a weakening of the northeasterly trades in the Northern Hemisphere in agreement with simplified equatorial atmospheric dynamics. The positive wind-evaporation SST feedback is likely to intensify the trade wind anomalies in the eastern equatorial Pacific. The tropical SST pattern is responsible for a northward shift of the intertropical convergence zone. Anomalous North Pacific precipitation patterns in the sub tropics and the extratropics can be directly linked to the weakened subtropical high.
Due to the stronger southeasterly trades and intensified southern hemispheric Westerlies (Fig. 6), the upwelling in the Eastern Equatorial Pacific and the Southern Ocean is enhanced, respectively. The globally reduced stratification and associated enhanced mixing cause an enhanced transfer of nutrients to the surface ocean. The globally averaged euphotic zone phosphate 3.3. Carbon cycle response

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content thus increases by 25% with regional values in the eastern equatorial and tropical Pacific attaining values of up to 60%. Changes in the North Pacific and Southern Ocean are smaller and amount to 35% and 20%, respectively (Fig. 7, top panel).

As a result of the enhanced delivery of nutrients to the euphotic zone, the export production increases globally by up to 20%. The strongest changes are seen in the Eastern Equatorial and Tropical Pacific (+60%), the Eastern Tropical Atlantic (+40%) and the North Pacific (+40%) (Fig. 7, middle panel). As light is the main limiting factor of the marine production in the polar Southern Ocean, despite the nutrient increase the export production slightly decreases there due to sea ice advance.
Due to the enhanced mixing the euphotic silicate content also increases by about 20%. This leads to 50% greater opal production in the North Pacific and an increase of 20% in the Eastern Equatorial Pacific (Fig. 7, bottom panel). In the Southern Ocean, the silicate content of the euphotic zone increases by 45%. As the export production is actually reduced south of Drake passage, the anomalously high silicate concentrations are advected to the southeastern Pacific and Atlantic. As a result, the opal production intensifies by 30% in these regions.

There is a slight increase in CaCO₃ sedimentation rate (not shown) in the Sea of Japan, the Kuril basin and off the west coast of North America between the latitudes 35 and 55°N. These changes are due to an enhanced export production and increased CaCO₃ preservation.

The increased ventilation of deep water leads to greater dissolved oxygen content in the deep part of the Atlantic and Pacific basins as well as in the intermediate layers of the North Pacific and Southern Ocean (Fig. 8, top panels). This version of the model calculates the ratio of ¹⁴C over ¹²C as a tracer from which the ventilation age of the water masses is calculated. After 1000 years of integration, the formation of NPDW leads to a substantial drop in simulated ventilation ages in the North Pacific between 1000 and 3500 m depth of about 1000 years (Fig. 8, bottom left panel).

The increased mixing brings also more DIC to the surface (+4%). Even though the enhanced export production leads to a greater flux of carbon from the surface to the deep ocean, the surface DIC content still increases. This leads to a greater CO₂ flux from the ocean to the atmosphere, namely in the Eastern Equatorial Pacific (not shown). As a result of the competitive effects of greater DIC transport to the surface and increased export production, the oceanic carbon reservoir loses only about 18 GtC. On the other hand, wetter conditions in the northern hemisphere induce a slightly greater carbon storage on land (+8 GtC, not shown). The atmospheric CO₂ thus increases by about 5 ppmv.

4. Summary and discussion

Using an Earth system model of intermediate complexity we quantified the effects of removing the subarctic North Pacific halocline on the large-scale ocean circulation, global climate and the carbon cycle. It was demonstrated that higher salinity in the North Pacific can trigger deep convection and the formation of North Pacific Deep Water in the Bering Sea. Large-scale adjustment processes in the ocean help to establish a Pacific meridional overturning circulation which advects heat and salinity poleward. This PMOC circulation does not exhibit any strong hysteresis behavior, which is indicative of a relatively weak Stommel feedback. The increased northward heat transport initiates warming in the North Pacific and adjacent Canada as well as a substantial cooling of the Southern Hemisphere. This behavior is consistent with the concept of the bipolar seesaw (Stocker, 1998). The North Pacific warming induces a reduction of the subtropical high which contributes to the warming of northern North America through anomalous warm air advection. Otherwise the global wind changes are relatively small.

In our experiments global deep-water formation increases which requires an intensification of diapycnal mixing of heat and tracers through a reduction in stratification. This response largely determines the biogeochemical changes simulated in our idealized sensitivity experiment. An enhanced mixing of nutrients to the surface stimulates productivity and also export production. In the Eastern Equatorial Pacific, export production is further amplified by an intensification of southeasterly trade winds that lead to an enhancement of upwelling. Changes of the surface DIC inventory partially compensate the carbon sink due to increased export production and higher terrestrial carbon uptake. It is concluded that the removal of the North Pacific halocline has only a small bearing on atmospheric CO₂.

The presented results on the physical response are not entirely without precedent. Using an OGCM coupled to an energy balance model (UVIC model) and salinity compensation, Saenko et al. (2004) also describe the formation of a PMOC in result to higher North Pacific surface salinities. In contrast to our results, Saenko et al. (2004) find a much stronger response of the AMOC to the establishment of the PMOC. We repeated experiment FE with salinity compensation (Stocker et al., 2007) but were not able to reproduce the UVIC model results. In a North Pacific “destratification” experiment performed with the Earth system of intermediate complexity CLIMBER-2, a substantial warming of the North...
Pacific and of Northern America is simulated, in accordance with our solution (Haug et al., 2005). It is further argued that the sudden build-up of the NPH may have helped to establish climate conditions that were more susceptible to ice-sheet growth. Our simulated temperature response to a removal of North Pacific surface stratification confirms their conjecture. Another important study that elucidates the effects of higher North Pacific salinities during the early and middle Pliocene is the modeling study by Motoi et al. (2005). Using an older version of the GFDL coupled general circulation model, the authors find that opening Panama Isthmus leads to a weakening of the AMOC, in agreement with Lunt et al. (2008) and Steph et al. (2010). The Panama gateway opening allows a salinity transport from the Atlantic into the Pacific which helps to establish a PMOC. Resulting changes in North Pacific heat transport (~ 0.4 PW at 45°N) generate a widespread North Pacific and North America warming that may have prevented the inception of glacial cycles in response to orbital forcing changes in the early Pliocene.

Opal accumulation rates and nitrogen-isotope data from a marine sediment core of the North Pacific (Haug et al., 1999) also suggest that the establishment of the NPH in the mid-Pliocene had a significant impact on the marine carbon cycle. Our model experiments presented here confirm the notion of strongly enhanced North Pacific opal production and reduced nutrient utilization in the North Pacific in the absence of a NPH (Haug et al., 1999); however, they do not support the claim that the formation of a NPH had a significant effect on atmospheric CO₂.

Different processes can be responsible for causing high stratification in the North Pacific. As suggested by Motoi et al. (2005) the closing of the Panama Isthmus prevented high saline waters from the Caribbean from leaking into the North Pacific. Moreover, it was demonstrated (Emile-Geay et al., 2003) that under present-day conditions the atmospheric moisture transport across Central America together with the east Asian monsoon play an important role in maintaining the salinity gradient between Atlantic and Pacific. Both of these atmospheric processes are strongly modulated through changes of the AMOC (Okazaki et al., 2010; Kiefer, 2010). It should be noted here that there exists quite some modeling uncertainty in regards to the question whether a weakened AMOC leads to a reduced (Krebs and Timmermann, 2007; Leduc et al., 2007) or an intensified model (Richter and Xie, 2010). It should be noted here that there exists quite some modeling uncertainty in regards to the question whether a weakened AMOC leads to a reduced (Krebs and Timmermann, 2007; Leduc et al., 2007) or an intensified model (Richter and Xie, 2010). It should be noted here that there exists quite some modeling uncertainty in regards to the question whether a weakened AMOC leads to a reduced (Krebs and Timmermann, 2007; Leduc et al., 2007) or an intensified model (Richter and Xie, 2010). It should be noted here that there exists quite some modeling uncertainty in regards to the question whether a weakened AMOC leads to a reduced (Krebs and Timmermann, 2007; Leduc et al., 2007) or an intensified model (Richter and Xie, 2010). It should be noted here that there exists quite some modeling uncertainty in regards to the question whether a weakened AMOC leads to a reduced (Krebs and Timmermann, 2007; Leduc et al., 2007) or an intensified model (Richter and Xie, 2010).

We realize that our study may well be relevant for the interpretation of Pliocene climate-biogeochemical processes but its initial motivation was to explain the global climate response to the potential onset of the PMOC during Heinrich event 1. Evidence of increased ventilation of intermediate to deep waters in the Northwestern Pacific during Heinrich event 1 and the YD was found in numerous marine sediment cores from the western North Pacific (Ahagon et al., 2003; Okhushi et al., 2004; Sagawa and Ikehara, 2008; Okazaki et al., 2010), indicating the formation of NPDW during periods of weak AMOC and the establishment of a Deep Western Boundary Current in the North Pacific. Such conditions require a significant salinity increase in the subtropical North Pacific which is in fact consistent with a salinity reconstruction from marine sediment core GH02-1030 (Sagawa and Ikehara, 2008). The LOVECLIM model results presented in Okazaki et al. (2010) and in Menviel et al. (in press) further support the scenario of a disappearing halocline during Heinrich event 1 and probably even during the Younger Dryas.

From recent modeling studies (Zhang and Delworth, 2005; Timmermann et al., 2007) it also becomes apparent that the climatic influence of an AMOC shutdown on the Pacific, in the absence of a PMOC, is opposite to the response described here. Given that PMOC and AMOC are tightly coupled in a series of model solutions (Mikolajewicz et al., 1997; Saenko et al., 2004; Krebs and Timmermann, 2007; Okazaki et al., 2010) it can be concluded that the detailed characteristics of millennial-scale North Pacific climate variations during the last glacial termination depend on a superposition of AMOC and PMOC-related processes.

Reconstructing the detailed spatio-temporal behavior of sea surface temperature, salinity and ocean ventilation in the North Pacific during glacial terminations will be an essential task that will further our understanding of the operation of the Atlantic-Pacific seesaw and its role in shaping glacial cycles.

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