Effects of the Bering Strait closure on AMOC and global climate under different background climates

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Abstract:

Previous studies have suggested that the status of the Bering Strait may have a significant influence on global climate variability on centennial, millennial, and even longer time scales. Here we use multiple versions of the National Center for Atmospheric Research (NCAR) Community Climate System Model (CCSM, versions 2 and 3) to investigate the influence of the Bering Strait closure/opening on the Atlantic Meridional Overturning Circulation (AMOC) and global mean climate under present-day, 15 thousand-year before present (kyr BP), and 112 kyr BP climate boundary conditions. Our results show that regardless of the version of the model used or the widely different background climates, the Bering Strait's closure produces a robust result of a strengthening of the AMOC, and an increase in the northward meridional heat transport in the Atlantic. As a consequence, the climate becomes warmer in the North Atlantic and the surrounding regions, but cooler in the North Pacific, leading to a seesaw-like climate change between these two basins. For the first time it is noted that the absence of the Bering

Strait throughflow causes a slower motion of Arctic sea ice, a reduced upper ocean water exchange between the Arctic and North Atlantic, reduced sea ice export and less fresh water in the North Atlantic. These changes contribute positively to the increased upper ocean density there, thus strengthening the AMOC. Potentially these changes in the North Atlantic could have a significant effect on the ice sheets both upstream and downstream in ice age climate, and further influence global sea level changes.

1. Introduction:

The Bering Strait is a shallow and narrow strait connecting the North Pacific and the Arctic. Presently, about 0.8 Sv of relatively fresher North Pacific water (Woodgate and Aagaard, 2005) is transported through this strait into the Arctic, and subsequently into the North Atlantic. Studies show that this strait opened about 4.8~7.4 million years ago (Marincovich and Gladebkov, 2001). The opening of the Bering Strait allows a short-cut communication between the Pacific and the Atlantic via the Arctic. A series of previous studies indicates that changes of the Bering Strait status, such as an opening or closing of this strait due to the raising and lowering of sea level associated with the retreat or growth of the land-based ice sheets, may have played an important role in modulating glacial-interglacial climate variability (Shaffer and Bendtsen, 1994; Nof, 2000; Nof and Van Gorder, 2003; De Bore and Nof, 2004a,b; Hu and Meehl, 2005; Hu et al., 2007, 2008, 2010, 2011, 2012a,b; Keigwin and Cook, 2007; Sandal and Nof, 2008a,b,c; Sarinthein et al., 2009; Ortiz et al., 2012). This is because the opening or closing of the Bering Strait affects the water mass exchange, especially freshwater, between the Pacific and the North Atlantic, leading to changes of deep convection in the subpolar North Atlantic. This affects the production of North Atlantic Deep Water (NADW), thus resulting in changes to

the global scale oceanic circulation involving the Atlantic Meridional Overturning Circulation (AMOC) or the Thermohaline Circulation (THC).

AMOC transports upper ocean warm and saline water into the subpolar North Atlantic where this water loses its heat to the overlying atmosphere and becomes dense, sinking to depth and flowing southward, and upwelling elsewhere in the world ocean (Dickson and Brown, 1994; Stocker and Broecker, 1994). This circulation is important to the Atlantic and global climate because the AMOC carries a significant amount of heat to the North Atlantic (Ganachaud and Wunsch, 2000), resulting in a mild climate downstream in Europe. Numerical simulations (e.g., Keigwin and Jones, 1994; Sakai and Peltier, 1997; Stouffer et al., 2006; Hu et al., 2013) and paleo-proxy records (Alley and MacAyeal, 1994; Broecker, 1994, Zahn et al., 1997; Clark et al., 2007; Gutjahr et al., 2010) suggest that the collapse of this circulation could induce significant cooling events, such as the Heinrich events (Heinrich, 1988; Hemming, 2004). Studies also suggest that variations of this circulation can affect the response of the climate system to greenhouse gas forcing in the future (e.g., Manabe et al., 1991; Delworth et al., 1993; Timmermann et al., 1998; Schmittner and Stocker, 1999; Hu et al., 2004a; Gregory et al., 2005; Schmittner et al., 2005; Timmermann et al., 2005a, b; Dahl et al., 2005; Zhang and Delworth, 2005; Levermann et al., 2005; Broccoli et al., 2006).

Previous studies indicate that the response of the AMOC to buoyancy forcing in the North Atlantic will affect the strength of the Bering Strait throughflow, but changes of the Bering Strait throughflow can also feed back to the strength of the AMOC (Shaffer and Bendtsen, 1994; De Bore and Nof, 2004a,b; Hu and Meehl, 2005; Hu et al., 2007, 2008, 2009, 2010, 2011, 2012a,b, 2013). Under present day climate, the relatively fresh Pacific water flows through this strait into the Arctic (Woodgate and Aagaard, 2005). Then the freshwater carried by this

throughflow is transported into the subpolar North Atlantic as liquid water or sea ice via the Fram Strait. This freshwater can influence the deep convection in the North Atlantic subpolar seas, and thus the strength of the AMOC. When the AMOC is stronger, it pulls more fresher North Pacific water into the subpolar North Atlantic, which tends to weaken the AMOC. When the AMOC is weak, less fresh North Pacific water flows into the subpolar North Atlantic, resulting in a strengthening of the AMOC on decadal to centennial time scales (Hu et al., 2011). In extreme cases, when the AMOC is really weak or collapsed in response to strong buoyancy forcing in the North Atlantic, the Bering Strait throughflow reverses direction. Instead of transporting Pacific water into the Atlantic, it transports North Atlantic water into the North Pacific (De Bore and Nof, 2004a,b; Hu and Meehl, 2005, Hu et al., 2007, 2008, 2012a,b, 2013), resulting in further changes of the global oceanic freshwater budget and circulation .

One of the plausible mechanisms proposed by previous studies (e.g., Manabe and Stouffer, 1988; Rahmstorf, 1996, 2002, 2005; Stocker and Wright, 1991; Stocker and Schmittner, 1997; Stocker, 2000; Stocker and Marchal, 2000; Stouffer et al., 2006, 2007; Mikolajewicz et al., 2007; Hu et al., 2012a,b) to explain the abrupt climate transitions during the last glacial period is AMOC hysteresis. Theoretical studies suggest that in response to slowly increasing buoyancy forcing in the North Atlantic, the AMOC gradually weakens at first (e.g., Stocker and Wright, 1991; Stocker and Schmittner, 1997; Rahmstorf, 1995, 2005; Stocker, 2002). When the buoyancy forcing becomes strong enough, the AMOC collapses suddenly and shuts off the oceanic meridional heat transport into the North Atlantic, inducing a significant cooling in the North Atlantic and surrounding regions, resulting in a cold event. On the other hand, when the buoyancy forcing reduces slowly, the AMOC stays in the collapsed state until the buoyancy forcing becomes small enough, then the AMOC suddenly restarts again, and the oceanic

meridional heat transport into the North Atlantic resumes, leading to a warm event, such as the Dansgaard/Oeschger events (Dansgaard et al., 1993; Ditlevsen et al., 2005). A recent coupled climate model study suggests that this AMOC hysteresis can only exist when the Bering Strait is closed, indicating that a sudden collapse of the AMOC can only occur during glacial times when the Bering Strait is closed or nearly closed (Hu et al., 2012a). This change at the Bering Strait is due to the buildup of land-based ice mass during the glacial times, which moves water from the ocean to land and lowers the global sea level, resulting in the closure of the Bering Strait. This closed Bering Strait effectively shuts off the water mass exchange from the North Pacific to North Atlantic via the Arctic. As a result, the external buoyancy forcing added into the North Atlantic, such as ice sheet melt water, is not transported out of the North Atlantic efficiently, leading to an accumulation of the buoyancy flux in the subpolar North Atlantic and the Arctic, resulting in a shutoff of the NADW formation, and the sudden collapse of the AMOC. On the other hand, an open Bering Strait reduces the flow of fresher North Pacific water into the North Atlantic when the AMOC weakens. The added buoyancy flux in the subpolar North Atlantic can then be exported into the Pacific via the Bering Strait when the AMOC is very weak, thus preventing a sudden collapse of the AMOC (Hu et al., 2012a). Instead, the AMOC can only slowly weaken/strengthen as the freshwater forcing in the North Atlantic is enhanced/weakened.

Previous studies also suggest that the closure or opening of the Bering Strait itself could affect the glacial climate by modifying the strength of the AMOC (Hu et al., 2010). This has been supported by recent observational evidence (Grützner and Higgins, 2010). During the early last glacial period, the closure of the Bering Strait enhanced the AMOC via preventing the fresher North Pacific water from being transported into the North Atlantic via the Arctic. This strengthened AMOC then transported more heat into the North Atlantic, and thus warmed the

North Atlantic and the surrounding regions which led to the initial melt-back of the ice sheets in North America. Once the ice sheets start to melt, the ice sheet surface albedo is reduced, allowing more solar radiation to be absorbed by the ice sheet and enhancing the melting process. At the same time, if the solar radiation also increases, it will lead to a further enhancement of the ice sheet melting. Hu et al. (2010) have proposed that these processes might be the mechanism to explain the significant sea level oscillations during the early last glacial period.

In this study, we explore the Bering Strait's opening/closure on the mean climate under vastly different climate background conditions, which has not been the focus of many previous studies (e.g., Hu and Meehl, 2005; Hu et al., 2007, 2008, 2010, 2012a, b, 2013). Here we use two versions of the National Center for Atmospheric Research (NCAR) Community Climate System Model – version 2 (CCSM2) and version 3 (CCSM3). Although these are two older versions of the CCSM models, they have been widely used for climate studies under a variety of climatic conditions and different climate applications in the last 10 years, especially CCSM3 (e.g., Otto-Blietner et al., 2006a, b, 2007, Meehl et al., 2005, 2010; Hu et al., 2008, 2010, 2011, and many others). Our results derived from simulations using these two models indicate that the climate impact of the opening/closure of the Bering Strait is very similar under widely different climate backgrounds. In addition, preliminary results from the newest version of the CCSM series (CCSM4 under early Pliocene background conditions with an open/closed Bering Strait) are consistent with our findings using CCSM2 and CCSM3. This leads us to conclude that the impact of the Bering Strait opening/closure on the global climate is independent of the versions of the CCSM models used in this study, and also is independent of the background climates. These results imply that the underlying physical mechanisms explored in this study are very likely to be at work for other coupled models as well based on the vast difference in physics and

numerics among these three versions of the CCSM model. Of course, to test these mechanisms further, a comparison study using different coupled models is ideal for the future, but out of the scope of this study. The rest of the paper is organized as follows: Section 2 describes the models and experiments; Section 3 shows the results; and Section 4 gives the discussion and conclusions.

2. Models and experiments:

2.1 Models

Here we use two versions of the NCAR Community Climate System Model – CCSM2 (Kiehl and Gent, 2004) and CCSM3 (Collins et al., 2006). The atmospheric components in CCSM2 and CCSM3 are the NCAR Community Atmospheric Model version 2 (CAM2) and version 3 (CAM3, Collins et al., 2004), respectively. Both use spectral dynamics at T42 resolution and 26 hybrid levels vertically. The ocean component of both models is a version of the Parallel Ocean Program (POP) developed at Los Alamos National Lab with 1° horizontal resolution and enhanced meridional resolution ($1/2^\circ$) in the equatorial tropics and the North Atlantic and with 40 vertical levels (Smith and Gent, 2002). The sea ice component is the Community Sea Ice Model version 4 (CSIM4) for CCSM2 and Community Sea Ice Model version 5 (CSIM5; Briegleb et al., 2004) for CCSM3 with elastic-viscous-plastic dynamics, a subgrid-scale thickness distribution, and energy conserving thermodynamics. The land component is the Community Land Model version 2 (CLM2) for CCSM2 (Bonan et al., 2002), and the Community Land Model version 3 (CLM3; Oleson et al., 2004) for CCSM3.

In comparison with CCSM2, improved physics and numerics have been implemented in each of the component models of CCSM3. The following is a list of the major physical and numerical improvements: 1. Better treatments of the cloud physics, precipitation processes,

radiation processes, aerosols and atmospheric dynamics ; 2. a more efficient solver for the barotropic continuity equation, a non-uniform absorption of solar radiation based on in situ chlorophyll and satellite ocean color observations (Ohlmann, 2003), and a parameterized double diffusive mixing scheme in the ocean component (the Parallel Ocean Program, POP1.4.3); 3. a modified formulation of the biogeophysics to increase the sensible and latent heat fluxes over sparsely vegetated surfaces, and a scheme to account for the effects of competition for water among plant functional types in the Community Land Model version 3 (CLM3); and 4. modifications to the sea ice dynamics, sea-ice albedos, and exchanges of salt between sea-ice and the surrounding ocean in Community Sea Ice Model version 5 (CSIM5).

Although both versions of the CCSM model used in this study produce stable global mean climate comparable to observations (Kiehl and Gent, 2004; Otto-Bliesner et al., 2006b; Collins et al., 2006), with all of the above-mentioned improvements in physics and numerics, the simulated climate in the CCSM3 control experiment run under 1990s conditions is better than that in CCSM2 when compared with observations (Collins et al., 2006, e.g. their Figures 5 and 6, and see also the references therein). The major improvements of the simulated climate in CCSM3 relative to CCSM2 include a reduced cold bias in the polar regions of the upper troposphere and in the tropical tropopause, a better energy balance at both surface and top of the atmosphere, a reduced sea surface temperature and salinity bias in many regions of the world ocean, a closer to observed Atlantic meridional heat transport, and a better sea ice thickness distribution. A more detailed description of these improvements can be found in Collins et al. (2006) and the references therein.

We run experiments, described below, in both model versions with their various differences to document how consistent are the responses to a closed and open Bering Strait

between two different models. This will provide context for interpretation of the processes and mechanisms and how robust those responses are to the Bering Strait throughflow.

2.2 Experiments:

Here we analyze 5 sets of experiments under different climate boundary conditions which are summarized in Table 1. Within each set of the experiments, everything is identical except one has an open Bering Strait and the other has a closed Bering Strait (i.e. continental outlines are the same in all experiments so there are no effects of sea level changes on continental shorelines). Each set of the simulations analyzed here runs parallel for 350 years and the last 300 years are used in this study. The closed Bering Strait simulations are branched from the corresponding open Bering Strait simulations after those open Bering Strait simulations reach quasi-equilibrium states. When the Bering Strait is closed, the AMOC normally takes about 50 years to reach a quasi-equilibrium state (Hu et al., 2007; 2008). Therefore we used the simulated results of the last 300 years.

The first two sets of the experiments are under present day climate boundary conditions in which the CO₂ concentration is fixed at 355 parts per million by volume (ppmv) using both CCSM2 and CCSM3 (hereafter referred as CCSM2 PD and CCSM3 PD, respectively). The second two sets of the experiments are under 112 thousand year before present (kyr BP) climate boundary conditions with a concentration CO₂ fixed at 257 ppmv and methane (CH₄) at 456 parts per billion by volume (ppbv, Petit et al., 1999; Loulergue et al., 2008). This is the first time during the last glacial period that the Bering Strait was closed due to the intensified growth of the North American ice sheets. In one of this second two sets of experiments, the Larentide Ice Sheet is included, but excluded in the other. Hereafter these two sets of experiments are referred

as 112kyr IS and 112kyr NIS, respectively. The final set of experiments uses the climate boundary conditions of 15 kyr BP with a CO₂ concentration of 214 ppmv, CH₄ at 350 ppbv (Petit et al., 1999; Loulergue et al., 2008), and the Larentide Ice Sheet is included. The 15 kyr BP condition is chosen for this study because it represents the mid-glacial condition which may characterize the general glacial condition better than the more widely used last glacial maximum (LGM). This set of simulations will aid us to further investigate how the Bering Strait status would affect the AMOC hysteresis under mid-glacial conditions which will be a follow up to the study of Hu et al. (2012a). This set of experiments will be referred as 15kyr IS later.

The ice sheet information for 112 kyr BP is based on the ice model of the Research School of Earth Sciences at the Australian National University (ANU). This model is developed from the inversion of sea-level and shoreline data constrained by the geological evidence for the ice margin migrations through time, and subject to the requirement that the changes in total ice volume are consistent with the analyses of sea level change at sites far from the former ice margins (Lambeck and Chappell, 2001). The analyses for the LGM and post-LGM (Marine Isotope Stage 2 (MIS-2) to MIS-1) ice are carried out separately for the individual major ice sheets in the northern hemisphere where there is a reasonable data base for these inversions (Lambeck et al., 1998; Fleming and Lambeck, 2004; Lambeck et al., 2010). The ice information for 15 kyr BP is from the ICE-5G model (Peltier, 2004; Peltier and Solheim, 2004) which is a well tested ice model for the inversion of the past ice sheet conditions. The data from this model has been adopted for the paleo-model intercomparison project phase 3 (PMIP3).

3. Results:

3.1 Changes of the AMOC, meridional heat and freshwater transport

Figure 1 shows the Atlantic meridional streamfunction from the 5 sets simulations, and the anomaly of the streamfunction between the closed and open Bering Strait simulations. In general, under present day climate conditions, the AMOC is stronger in CCSM3 than in CCSM2. As shown in Table 2, the mean strength of the AMOC in both CCSM3 and CCSM2 with an open Bering Strait is similar to that suggested by observations (e.g., Ganachaud and Wunsch, 2000; Cunningham et al., 2007, 2010; Johns et al., 2011). Under glacial conditions, the AMOC with an open Bering Strait is as strong as or even stronger than the AMOC under present day conditions. It is still under debate whether the AMOC was stronger or not under glacial conditions relative to the AMOC under present day conditions (Yu et al., 1996; McManus et al. 2004). Results from the paleo-model intercomparison project phase 2 (PMIP2) suggest that the AMOC during the last glacial maximum was neither appreciably weaker or stronger than modern (Weber et al., 2007; Otto-Bliesner et al., 2007). When the Bering Strait is closed, all simulations show that the AMOC is about 2~3 Sv stronger than that in the open Bering Strait simulations. Thus, it is clear that a strengthened mean AMOC is a robust result across all experiments in spite of different models and drastically different background climate conditions.

Figure 2a confirms this result and shows that the AMOC is more than 10% stronger with a closed Bering Strait than with an open Bering Strait in all of the simulations regardless of the versions of CCSM model used or the background climate conditions (The absolute values of the AMOC in all simulations are shown in Table 2). As the AMOC becomes stronger in all cases with a closed Bering Strait, warmer tropical-subtropical water is pulled into the subpolar North Atlantic, leading to a strengthened northward meridional heat transport (MHT) at all latitudes in the Atlantic basin. At 24°N, the MHT increases by at least 10% relative to the MHT with an open Bering Strait (Figure 2b, Table 1). On the other hand, the northward MHT in the North Pacific is

reduced and southward MHT increases in the South Pacific (Figure 2c, d). This increase/decrease in Pacific MHT represents an overall reduction of the tropical heat gain in the Pacific being transported into the North Pacific with a decrease by approximately 10% or more at 24°N, and greater heat transport into the South Pacific by roughly 10% or more at 30°S. This increased southward Pacific MHT is further transported into the Atlantic basin by the AMOC. Thus our results suggest that as the AMOC becomes stronger and the northward Atlantic MHT increases due to the closure of the Bering Strait, more heat is pulled from the Pacific and transported into the North Atlantic. Therefore, the closure of the Bering Strait can induce global scale changes of the oceanic circulation and meridional redistribution of heat, agreeing with Hu et al. (2010). In addition, our results also indicate these oceanic changes due to a Bering Strait closure or opening are independent of the background climate conditions.

As shown in Table 2, the simulated Bering Strait throughflow varies from 0.78 to 1.02 Sv in our five open Bering Strait simulations, agreeing well with modern observations (e.g., Aagaard and Carmack, 1989; Woodgate and Aagaard, 2005). Table 2 also shows that the strength of the Bering Strait throughflow is, in general, positively related to the strength of the AMOC. A stronger AMOC leads to a stronger Bering Strait throughflow. As suggested by previous studies, the strength of the Bering Strait throughflow is mainly dynamically controlled and determined by the sea level difference between the North Pacific and the North Atlantic (Shaffer and Bentsen, 1994; Hu et al., 2008; 2009, 2011), and may also partly influenced by the southern ocean winds (Nof, 2000; Ortiz et al. 2012). As the AMOC becomes stronger, the water in the subpolar North Atlantic becomes denser and the sea level becomes lower, leading to an increased sea level contrast between the subpolar North Pacific and subpolar North Atlantic, and a stronger Bering Strait throughflow, and vice versa for a weaker AMOC (Hu et al., 2009, 2011).

In our simulations, the liquid freshwater transport via the Bering Strait from the Pacific into the Arctic varies from 0.062 to 0.093 Sv which is also within the uncertainty range of the modern observations (Woodgate and Aagaard, 2005). As suggested by Aagaard and Carmack (1989), the freshwater imported from the Pacific via the Bering Strait makes up roughly 30% of the total freshwater gain in the Arctic. When the Bering Strait is closed under glacial conditions, this freshwater transport from the North Pacific into the Arctic via the Bering Strait is restricted. The lack of freshwater transport from the North Pacific into the Arctic leads to a significantly reduced freshwater export from the Arctic into the North Atlantic in all our simulations (Figure 2e, Table 2). This reduction, percentage wise, becomes extremely large when ice sheets are present in the North American continent under glacial conditions. When the Bering Strait is closed under glacial conditions with the presence of the North American Ice Sheets, the Arctic's role in the freshwater budget in the North Atlantic actually changes from a freshwater source to a freshwater sink (Table 2). In the other words, instead of exporting liquid freshwater from the Arctic into the North Atlantic, now the freshwater is being transported from the North Atlantic into the Arctic. This freshwater import into the Arctic is balanced by the sea ice export from the Arctic into the Atlantic, in addition to the evaporation, precipitation and river runoff in the Arctic. In fact, the sea ice export from the Arctic into the North Atlantic is also reduced in all our simulations with a closed Bering Strait relative to the simulations with an open Bering Strait, indicating an overall reduction of the freshwater export from the Arctic into the North Atlantic, leading to a saltier North Atlantic, denser sea surface water, stronger deep convection in the subpolar North Atlantic, and a more vigorous AMOC.

On the other hand, the closure of the Bering Strait leads to an increased surface freshwater gain in the North Pacific because this closure prevents the relatively fresh North Pacific water

from exiting into the Arctic. This freshwater gain results in an increased southward meridional freshwater transport from the subpolar North Pacific towards the equatorial Pacific. At 40°N, this increase with a closed Bering Strait is at least 25% of that with an open Bering Strait (Figure 2f, Table 2). The absolute values increase from about 0.23~0.25 Sv with an open Bering Strait to 0.32~0.33 Sv with a closed Bering Strait. This increased southward meridional freshwater transport with a closed Bering Strait is roughly equivalent to the freshwater water transport from the North Pacific into the Arctic via the Bering Strait (Table 2). These results indicate the closure/opening of the Bering Strait during glacial times can potentially cause not only significant changes of the freshwater budget in the Atlantic and the Pacific, but also further affect the strength of the AMOC and the MHT. These changes may affect global scale surface climate as suggested by Hu et al. (2010). We will address these corresponding global changes in the next subsection with a focus on the North Pacific and North Atlantic.

3.2 Changes of the surface climate with a focus on the North Atlantic and North Pacific

Figure 3 shows the surface temperature changes in the simulations with a closed Bering Strait in comparison to the simulations with an open Bering Strait. In general, compared to the open Bering Strait simulations, the North Atlantic becomes warmer and the North Pacific becomes cooler when the Bering Strait is closed. In the Southern Hemisphere, the South Atlantic becomes cooler and the South Pacific between 10 and 40°S becomes warmer. This pattern of surface temperature change in these basins is consistent with changes of the oceanic MHT as discussed in the previous subsection and suggested by Hu et al. (2010). As the AMOC strengthens, more heat is transported into the North Atlantic inducing a warming there. At the same time, in the Pacific basin, there is reduced northward MHT from the tropics into the

subpolar North Pacific, and the southward MHT into the South Pacific increases (Figure 2b, c, Table 2). This excess heat is further transported into the Atlantic basin via the Drake Passage. This is why there is surface warming in the Drake Passage and the Atlantic sector of the Southern Oceans in all simulations with a closed Bering Strait (Figure 3).

Earlier studies based on standalone ocean model simulations (Saenko et al., 2004) indicate that the climate changes between the North Atlantic and the North Pacific could oppose each other, with seesaw-like climate fluctuations between the two basins. This seesaw-like climate change is associated with variations of the AMOC. By using a standalone ocean model with a closed Bering Strait, Saenko et al. (2004) found that by increasing the freshwater input into the North Atlantic, the AMOC weakens and the North Atlantic cools. However, the Pacific MOC (PMOC) strengthens and the North Pacific warms up. If the freshwater input in the North Pacific increases, the AMOC strengthens and the PMOC weakens, resulting in a warming in the North Atlantic and a cooling in the North Pacific. By employing a coarse resolution ocean model, Seidov and Haupt (2003, 2005) suggested that the sea surface salinity (SSS) changes between the North Atlantic and North Pacific are opposite to each other in association with the AMOC fluctuations. They proposed that the SSS contrast between these two basins can be used as an indicator of the AMOC changes. Later this idea was proved in two fully coupled climate models (Hu et al., 2004a,b). An observation-based modelling study (Okazaki et al., 2010) showed that the paleo proxy record indicates that during Heinrich event 1 (H1), the North Pacific was warmer. They used an Earth System Model of Intermediate Complexity (EMIC) to illustrate that this warmer North Pacific during H1 is due to a setup of the PMOC when the AMOC collapsed. A recent coupled model study (Hu et al., 2012b) illustrated that these seesaw-like SSS and surface temperature changes between the two basins only work if the Bering Strait is closed.

The results from all our simulations show that closing the Bering Strait can induce a seesaw-like surface temperature change between these two basins in comparison to the simulations with an open Bering Strait (Figures 3, 4 and Table 3), and that this too is a robust result.

Further, we find that the closing of the Bering Strait not only causes a seesaw-like surface temperature change between the North Pacific and North Atlantic, but also induces seesaw-like changes in SSS, precipitation, net surface heat flux, and net surface freshwater input in these two basins (Figures 4-9, and Table 3). The SSS changes show increases in the North Atlantic, Arctic and the regions in the Southern Oceans where sea ice is present, and decreases in other basins, especially the North Pacific in all closed Bering Strait simulations in comparison to the ones with an open Bering Strait (Figures 4, 5). The precipitation patterns between the closed and open Bering Strait simulations show a reduction in the North Pacific and an increase in the North Atlantic (Figures 6, 7), which is opposite to the SSS changes in these two basins (Figures 4, 5), suggesting that other processes may be responsible for the SSS changes. On the other hand, the tropical precipitation changes show a southward movement of the inter-tropical convergence zone (ITCZ) in the Pacific and a northward movement in the Atlantic associated with the cooling in the North Pacific and the warming in the North Atlantic. This is consistent with previous studies (e.g., Hu et al., 2010; Brady and Otto-Blietner, 2010).

The surface freshwater input includes precipitation, evaporation, river runoff, and sea ice melt-freeze (Figure 8). In general, with a closed Bering Strait, the surface freshwater input in the North Atlantic is reduced (Figures 7, 8). But the changes of surface freshwater input are mixed in the North Pacific with a Bering Strait closure (Figures 7, 8). As shown in Table 3 and Figure 7, the net surface freshwater input in the North Pacific between 40 and 60°N is actually increased

in the simulations with a closed Bering Strait compared to the simulations with an open Bering Strait, with corresponding decreases in SSS. In the North Atlantic (40-80°N), the surface freshwater input with a closed Bering Strait is decreased, contributing positively to the SSS increase there (Table 3, Figure 7). This indicates that other components in the surface freshwater input overcome the reduction of precipitation in the North Pacific and the increase in precipitation in the North Atlantic. An analysis of each surface freshwater component shows that as the Bering Strait is closed, surface evaporation and river runoff into the North Pacific decrease, but freshwater input from sea ice melt increases (Table 3, Figure 7). Thus it is the increased sea ice melt and especially the reduced evaporation in this basin that cause the overall increase in surface freshwater input. In the North Atlantic basin, evaporation increases and sea ice melt decreases when the Bering Strait is closed in all simulations, with mixed changes of river runoff into this basin. Therefore, the overall reduction of surface freshwater input is caused primarily by the increased evaporation and reduced sea ice export from the Arctic (Table 3, Figure 7). The role of sea ice to the fresh water budget and associated changes in AMOC have been noted by previous studies (e.g., Hu et al., 2005; Jahn and Holland, 2013) for present and future climates, but are documented here for the first time in the context of a closed or open Bering Strait. The increased evaporation in the North Atlantic and the decreased evaporation in the North Pacific are related to the changes of surface temperature due to alterations of the AMOC and MHT induced by the closure of the Bering Strait. As the surface temperature increases in the subpolar North Atlantic due to the increased northward MHT, the overall heat loss also increases (Figures 7 and 9, Table 3). This increased heat loss includes contributions from greater sensible and latent heat flux. The increased loss of latent heat flux leads to increased evaporation in this basin. In the subpolar North Pacific, the cooling due to a reduced northward

MHT with a closed Bering Strait produces a reduced total surface heat loss, resulting in a reduced evaporation there. Therefore, the closure of the Bering Strait not only causes changes of the AMOC and MHT, but also modulates the surface freshwater and heat balance, especially in the subpolar North Pacific and North Atlantic. These changes induce a seesaw-like climate change between these two basins.

It is worth noting that as the Bering Strait is closed, the freshwater and sea ice exchanges between the North Pacific and the Arctic are restricted. The increased ice melt in the North Pacific with a closed Bering Strait is a result of the cooler climate which causes increased sea ice formation during the winter season. Therefore this increased ice melt does not contribute much to the annual mean SSS change there, and the increased surface freshwater input in this basin is mainly due to the reduction of evaporation more than the decrease of precipitation in the North Pacific. On the other hand, in the subpolar North Atlantic basin, the sea ice melt includes two parts: the sea ice export from the Arctic and the local sea ice formation in winter season. As shown in Table 3 and Figure 7, the sea ice export from the Arctic into the North Atlantic is reduced when the Bering Strait is closed in all closed Bering Strait simulations, thus contributing positively to the net reduction of the surface freshwater input and the increase of the SSS there.

3.3 Changes in Arctic Sea ice

In this subsection, we focus more on this new aspect of how Arctic sea ice can play a crucial role in the climate changes associated with either a closed or open Bering Strait. Figures 10 and 11 show the March and September mean sea ice thickness and velocity in the open Bering Strait simulations under different background climates. Under present day conditions, the March mean sea ice thickness is about 2 to 3 meters in CCSM2 (Figure 10a), and roughly 3 to 6

meters in CCSM3 (Figure 10b). The simulated sea ice thickness agrees with observations better in CCSM3 than in CCSM2 (Holland et al., 2006). The sea ice is much thicker, up to 10 meters thick, under 112 kyr BP conditions (Figure 10c and d). This thicker sea ice is due to much reduced solar radiation reaching the earth at that time. At 65°N, the June solar radiation is about 35 W/m² less at 112 kyr BP than under present day conditions (Berger, 1978; Loutre et al., 2004; Hu et al., 2010). However, for 15 kyr BP, the reduction of the June solar radiation at 65°N is only 5 W/m² relative to that of present day. Resulting from this, the March mean sea ice thickness (Figure 10e) under 15 kyr BP climate conditions is only a couple of meters thicker than that of the present day simulation (Figure 10b). In September (Figure 11), the sea ice extent shrinks and the sea ice also becomes thinner than in March (Figure 10). The simulated sea ice velocity in all cases shows an anticyclonic motion in both March and September, agreeing with observations (e.g., Kwok, 2008). This anticyclonic sea ice motion helps to export Arctic sea ice into the subpolar North Atlantic, contributing to the freshwater balance there.

When the Bering Strait is closed, both March and September sea ice become thicker in the Canadian basin of the Arctic, but thinner in part of the Eurasian basin of the Arctic (Figures 12-13). On average, the basin mean sea ice becomes thicker in all closed Bering Strait simulations relative to the open Bering Strait simulations (Table 4). The changes of the basin mean sea ice extent, concentration, and total volume are mixed. There are more cases with an increased sea ice extent, concentration and total volume in general (Table 4). For the simulations under present day conditions, the sea ice extent, thickness, concentration, and total volume are all increased with the closure of the Bering Strait. For the cases under glacial conditions with North American Ice Sheets, the sea ice extent is slightly increased under 112 kyr BP conditions, but decreased somewhat under 15 kyr BP conditions. On the other hand, the sea ice concentration is

decreased under 112 kyr BP conditions, and increased under 15 kyr BP conditions. With the increase of the mean sea ice thickness, the total sea ice volume increases in both of these cases with a closed Bering Strait relative to the open Bering Strait simulations (Table 4). Without the North American ice sheets, although the sea ice thickness and concentration are both increased under 112 kyr BP conditions, the more significant reduction of the sea ice extent leads to a reduction of total sea ice volume when the Bering Strait is closed (Table 4).

By comparing sea ice velocity fields in March and September in the open Bering Strait simulations, the anomalous sea ice velocity shows a cyclonic pattern when the Bering Strait is closed. This pattern of sea ice velocity results in two major changes of the Arctic sea ice: 1. The sea ice export from the Arctic via Fram Strait is reduced (Table 1, Figure 7); 2. The overall sea ice motion becomes slower in the Arctic (Figures 10-13). The slower sea ice motion and reduced sea ice export both contribute to the increase of sea ice thickness. The changes of the Arctic sea level pressure field show mixed changes in our simulations when the Bering Strait is closed under different background climates, suggesting that the changes of the sea level pressure field and the associated changes of winds may not be the primary cause of the slower sea ice motion in the closed Bering Strait simulations relative the open Bering Strait simulations (Figure not shown). As indicated by Shaffer and Bentsen (1994), the Bering Strait throughflow is driven primarily by oceanic dynamics associated with the sea level difference between the North Pacific and the Arctic (subpolar North Atlantic). A larger sea level difference between these ocean basins will result in a stronger Bering Strait throughflow. As indicated by many previous studies, the Arctic currents are a complicated system which is affected by multiple processes, such as the net input of freshwater from evaporation minus precipitation, river runoff (e.g., Aagaard and Carmack, 1989; Carmack et al., 2008), the overlying atmospheric variability (Dickson et al.,

2000), etc., and interact with the sea ice export and the thermohaline circulation. However, this net water mass inflow from the North Pacific via the Bering Strait is at least one of the driving forces of the Arctic surface flow, and makes it flow toward the subpolar North Atlantic as shown in Figure 14 in the simulations with an open Bering Strait. When the Bering Strait is closed, this driving force is absent, leading to a much slower motion of the surface water (Figure 15) through a set of complicated processes in responding to the absence of this throughflow. This reduced surface water motion leads to slower sea ice motion in the closed Bering Strait simulations. The water mass exchange in the upper 100 meters of the ocean between the Arctic and the subpolar North Atlantic is reduced when the Bering Strait is closed (see right most panels in Figure 1). The reduction of this exchange is equivalent to the strength of the Bering Strait throughflow. This reflects the fact that the export of both the upper ocean water and sea ice in the Arctic decreases with the closure of the Bering Strait.

Figures 16 and 17 show the simulated snow depth on sea ice in March and September with an open Bering Strait. Under present day and 15 kyr BP climate conditions, the simulated snow depth is only a few tenths of a meter with an open Bering Strait, but reaches up to 2 meters under 112 kyr BP climate conditions in both March and September. With a closed Bering Strait, the pattern of snow cover change is similar to the sea ice change (Figures 18, 19). Where the sea ice becomes thicker, the snow depth also becomes deeper, and vice versa. This suggests that with thicker ice, less heat from underneath the ocean reaches the ice surface, thus reducing the melting of snow on top of the sea ice. Conversely, more snow on top of sea ice will increase the albedo during summer, leading to a reduced melt of sea ice and snow.

4. Conclusions

In this study, we investigated the changes of the AMOC, North Pacific and North Atlantic climate, and Arctic sea ice due to opening or closing of the Bering Strait under different climate boundary conditions using two global coupled climate models, the NCAR CCSM2 and CCSM3. Our results suggest that, in general, the response of the AMOC to the Bering Strait closure is qualitatively the same across the two model versions and across quite different climate boundary conditions, with the AMOC becoming stronger in all simulation pairs. This stronger AMOC transports more heat from the rest of the ocean into the North Atlantic, leading to a warming in the North Atlantic and a cooling in the North Pacific in comparison to the mean climate with an open Bering Strait, a seesaw-like climate change between the North Atlantic and the North Pacific. This seesaw-like change can not only be found in SST, but also in SSS, precipitation, evaporation, surface freshwater input and total heat loss in these two basins. Furthermore, we also show that the lack of Bering Strait throughflow due to a Bering Strait closure causes a slower upper Arctic water and sea ice motion, resulting in thicker sea ice and deeper snow on top of sea ice in the Arctic, and reduced sea ice export from the Arctic into the North Atlantic. This helps to maintain a more saline and dense subpolar North Atlantic, and a more vigorous AMOC.

From our results, it is clear that although the Bering Strait is a shallow and narrow strait, the status of this strait – open or closed – possibly has played not only a vital role in modulating past global climate (Shaffer and Bendtsen, 1994; De Bore and Nof, 2004a,b; Hu and Meehl, 2005; Hu et al., 2007, 2008, 2010, 2012a,b; Keigwin and Cook, 2007; Sandal and Nof, 2008a,b,c; Sarnthein et al., 2009), but also could play a key role in understanding the mechanisms involved with the future climate response to increasing greenhouse gases (Hu et al., 2009, 2011, 2013). Our study also suggests that it is important that a proper Bering Strait

condition is used for studies of paleoclimate, as well as modern and future climates. Biases will arise if an open Bering Strait is used for studies of past glacial climates when the Bering Strait was actually closed, or if a closed Bering Strait is used for present day or future climate simulations when the Bering Strait is actually open (e.g. Hu et al., 2012a, b).

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Acknowledgements: A portion of this study was supported by the Office of Science (BER), U.S. Department of Energy, Cooperative Agreement No. DE-FC02-97ER62402. The National Center for Atmospheric Research is sponsored by the National Science Foundation.

Tables:

Table 1 A summary of the experiments

Experiment name	Model Version	Background climate	Bering Strait	North American Ice Sheets
CCSM2 PD OBS	CCSM2	Present day	open	No
CCSM2 PD CBS	CCSM2	Present day	closed	No
CCSM3 PD OBS	CCSM3	Present day	open	No
CCSM3 PD CBS	CCSM3	Present day	closed	No
112kyr NIS OBS	CCSM3	112 thousand year before present day	open	No
112kyr NIS CBS	CCSM3	112 thousand year before present day	closed	No
112kyr IS OBS	CCSM3	112 thousand year before present day	open	Yes
112kyr IS CBS	CCSM3	112 thousand year before present day	closed	Yes
15kyr IS OBS	CCSM3	15 thousand year before present day	open	Yes
15kyr IS CBS	CCSM3	15 thousand year before present day	closed	Yes

OBS is the open Bering Strait simulation, and CBS is the closed Bering Strait simulation. PD represents present day; kyr BP represents thousand year before present; NIS represents “no North America ice sheets; and IS represents “with North America ice sheets”.

Table 2. The oceanic mean state of the open and closed Bering Strait simulations

	CCSM2 PD		CCSM3 PD		112kyr NIS		112kyr IS		15kyr IS	
	OBS	CBS	OBS	CBS	OBS	CBS	OBS	CBS	OBS	CBS
AMOC	15.7	18.3	19.4	22.2	19.2	24.5	24.9	27.9	19.5	22.1
NAMHT 24°N	0.84	0.98	1.05	1.19	1.04	1.26	1.25	1.39	1.19	1.32
NPMHT 24°N	0.91	0.77	0.88	0.76	0.87	0.74	0.82	0.79	0.78	0.71
SPMHT 30°S	-0.45	-0.60	-0.61	-0.76	-0.77	-0.97	-0.92	-1.08	-0.86	-0.96
Arc-Atl FW	-0.171	-0.114	-0.218	-0.150	-0.072	-0.047	-0.055	0.007	-0.017	0.043
NPMHF 40°N	-0.23	-0.32	-0.25	-0.32	-0.26	-0.32	-0.23	-0.33	-0.23	-0.30
BS transport	0.81	--	0.86	--	0.78	--	1.02	--	0.83	--
BS FW transport	0.062	--	0.081	--	0.067	--	0.093	--	0.063	--
Arc-Atl SIE	-0.129	-0.124	-0.181	-0.167	-0.349	-0.317	-0.412	-0.344	-0.402	-0.364

The units for Atlantic meridional overturning circulation (AMOC), Arctic to Atlantic freshwater transport (Arc-Atl FW), North Pacific freshwater transport at 40°N (NPMHF 40°N), Bering Strait mass (BS transport) and freshwater transports (BS FW transport), and Arctic to Atlantic sea ice export (Arc-Atl SIE) are Sv ($10^6 \text{ m}^3 \text{ s}^{-1}$), and for the North Atlantic meridional heat transport at 24°N (NAMHT 24°N), North (South) Pacific meridional heat transport at 24°N (30°S) (NPMHT 24°N or SPMHT 30°S) are PW (10^{15} W). OBS is the open Bering Strait simulation, and CBS is the closed Bering Strait simulation. PD represents present day; kyr BP represents thousand year before present; NIS represents “no North America ice sheets; and IS represents “with North America ice sheets”.

Table 3 Changes of surface property in the North Pacific and North Atlantic

	CCSM2 PD		CCSM3 PD		112kyr NIS		112kyr IS		15kyr IS	
	OBS	CBS	OBS	CBS	OBS	CBS	OBS	CBS	OBS	CBS
SST NP	9.33	8.41	6.83	6.34	4.37	4.16	4.11	3.49	3.71	3.50
SST NA	8.04	8.51	6.85	7.30	4.36	5.75	5.03	5.69	3.01	3.68
SSS NP	32.89	32.36	32.55	31.94	32.27	31.75	31.95	31.35	32.13	31.86
SSS NA	32.24	32.85	32.55	33.20	32.58	33.46	33.53	33.92	33.29	33.68
SPD NP	25.28	25.00	25.40	24.98	25.46	25.07	25.25	24.84	25.45	25.25
SPD NA	24.89	25.29	25.32	25.76	25.66	26.19	26.36	26.58	26.44	26.67
SHF NP	-19.47	-14.06	-15.46	-11.23	-15.30	-10.67	-13.72	-8.95	-12.11	-9.53
SHF NA	-28.95	-35.20	-38.69	-45.02	-45.57	-54.32	-57.18	-63.64	-45.56	-56.29
SFWF NP	.3377	.3537	.3670	.3781	.3706	.3801	.3806	.3952	.3503	.3525
SFWF NA	.3275	.3111	.3501	.3290	.4507	.4002	.4296	.3867	.3838	.3655
EVAP NP	-.3179	-.2904	-.2613	-.2416	-.2224	-.2056	-.2156	-.1923	-.1963	-.1858
EVAP NA	-.3237	-.3537	-.3263	-.3558	-.2800	-.3416	-.2847	-.3177	-.2252	-.2539
PREC NP	.5559	.5341	.5172	.4988	.4593	.4433	.4604	.4364	.4132	.4026
PREC NA	.4409	.4568	.4451	.4600	.3847	.4271	.3345	.3554	.2964	.3173
ROFF NP	.0887	.0861	.0896	.0877	.0826	.0822	.0919	.0891	.0793	.0784
ROFF NA	.1019	.1023	.1019	.1017	.0931	.0950	.1134	.1147	.0932	.0950
MELT NP	.0180	.0239	.0249	.0382	.0576	.0689	.0500	.0704	.0618	.0647
MELT NA	.0899	.0867	.1197	.1121	.2574	.2192	.2739	.2360	.2198	.2047
Denmark ice exp	0.044	0.043	0.065	0.063	0.149	0.144	0.181	0.161	0.176	0.173

The unit for sea surface temperature (SST) is °C, for sea surface salinity (SSS) is psu, for surface potential density (SPD) is kg/m³, for surface heat flux (SHF) is W/m², and for surface freshwater flux (SFWF), evaporation (EVAP), precipitation (PREC), river runoff (ROFF), sea ice melt (MELT) and Denmark Strait sea ice export (Denmark ice exp) is Sv (10⁶ m³/s). NA represents North Atlantic (40-80°N), and NP North Pacific (40-60°N). OBS is the open Bering Strait simulation, and CBS is the closed Bering Strait simulation. PD represents present day; kyr BP represents thousand year before present; NIS represents “no North America ice sheets; and IS represents “with North America ice sheets”.

Table 4 Changes of the Arctic surface properties

	CCSM2 PD		CCSM3 PD		112kyr NIS		112kyr IS		15kyr IS	
	OBS	CBS	OBS	CBS	OBS	CBS	OBS	CBS	OBS	CBS
Ice extent	11.33	11.93	12.80	13.21	17.45	16.68	16.55	16.73	16.55	16.39
Thickness	1.34	1.38	2.13	2.34	5.27	5.40	5.37	5.41	3.16	3.28
Concentr.	81.24	81.35	84.52	84.96	87.27	87.92	87.65	87.57	85.88	86.33
Volume	15.17	16.39	26.53	29.97	87.81	85.89	85.62	86.97	49.86	51.04
SST	-0.540	-0.629	-0.903	-1.010	-1.655	-1.594	-1.684	-1.686	-1.627	-1.645
SSS	29.42	29.57	29.29	29.46	30.06	30.83	30.19	31.07	29.64	30.02
SPD	23.60	23.73	23.52	23.66	24.19	24.81	24.30	25.02	24.02	24.33
SHF	-12.77	-11.86	-13.17	-11.69	-3.73	-5.62	-2.91	-4.15	-3.89	-3.75
SFWF	.1280	.1287	.1331	.1393	.0125	.0384	-.0141	.0122	-.0364	-.0166
EVAP	-.0375	-.0352	-.0313	-.0278	-.0084	-.0111	-.0078	-.0083	-.0082	-.0076
PREC	.0478	.0449	.0365	.0326	.0098	.0121	.0079	.0078	.0075	.0070
ROFF	.1191	.1173	.1404	.1409	.1221	.1254	.1411	.1404	.0982	.0989
MELT	-.0201	-.0179	-.0384	-.0232	-.1500	-.1272	-.2020	-.1727	-.1713	-.1514

The unit for sea ice extent is 10^6 km^2 , for ice thickness is meter, for sea ice concentration is percent (%), for sea ice volume is 10^3 km^3 , for sea surface temperature (SST) is $^{\circ}\text{C}$, for sea surface salinity (SSS) is psu, for surface potential density (SPD) is kg/m^3 , for surface heat flux (SHF) is W/m^2 , and for surface freshwater flux (SFWF), evaporation (EVAP), precipitation (PREC), river runoff (ROFF), and sea ice melt (MELT) is Sv ($10^6 \text{ m}^3/\text{s}$). OBS is the open Bering Strait simulation, and CBS is the closed Bering Strait simulation. PD represents present day; kyr BP represents thousand year before present; NIS represents “no North America ice sheets; and IS represents “with North America ice sheets”.

Figure Captions:

Figure 1. The Atlantic meridional streamfunction (AMSF). The left panels are for the open Bering Strait simulations, and the mid-panels are for the closed Bering Strait simulations, and the right panels are the changes of the AMSF between closed and open Bering Strait simulations. PD represents present day; kyr BP represents thousand year before present; NIS represents “no North America ice sheets; and IS represents “with North America ice sheets”. CCSM2 PD or CCSM3 PD represent the simulation using CCSM2 or CCSM3 with present day climate background conditions. 112kyr BP NIS represents CCSM3 simulations under 112 thousand years before present climate conditions without the North America ice sheets. 112kyr IS and 15kyr IS are the simulations using CCSM3 under 112 and 15 thousand year before present climate background conditions with the North America ice sheets. The same abbreviations are used in all figures.

Figure 2. Percentage changes of the Atlantic Meridional Overturning Circulation (AMOC), Atlantic meridional heat transport (MHT) at 24°N, Pacific MHT at 24°N and 30°S, the freshwater transport (FWT) from Arctic to the North Atlantic, and the FWT in the Pacific at 40°N.

Figure 3. Surface temperature changes when the Bering Strait is closed under different background climate conditions. The contour interval is 0.1°C. Stippling indicates the statistically significant changes by using a double sided t-test.

Figure 4. Regional mean sea surface temperature (SST), sea surface salinity (SSS), sea potential surface density (SPD), and total surface heat flux changes in the closed Bering Strait simulations relative to the open Bering Strait simulations in the North Pacific (NP, 40-60°N), North Atlantic

(NA, 40-80°N), and the Arctic. The unit for SST is °C, for SSS psu, for SPD kg/m³, and for surface heat flux W/m².

Figure 5. Sea surface salinity changes between the closed and open Bering Strait simulations. The contour interval is 0.1 psu. Stippling indicates the statistically significant changes by using a double sided t-test.

Figure 6. Annual precipitation changes between the closed and open Bering Strait simulation. Contour interval is 0.03 m/yr. Stippling indicates the statistically significant changes by using a double sided t-test.

Figure 7. Total surface freshwater input, precipitation, evaporation, river runoff, sea ice melt anomaly in the closed Bering Strait simulations relative to the open Bering Strait simulations in the North Pacific (NP, 40-60°N), North Atlantic (NA, 40-80°N), and the Arctic. And, the sea ice export anomaly from the Arctic into the North Atlantic in the closed Bering Strait simulations relative to the open Bering Strait simulations. The units for all of these are Sv (Sv=10⁶m³/s).

Figure 8. Surface freshwater input anomaly between the closed and open Bering Strait simulations. Stippling indicates the statistically significant changes by using a double sided t-test. Contour interval is 0.05 m/yr.

Figure 9. Annual mean surface heat flux anomaly between the closed and open Bering Strait simulations. Stippling indicates the statistically significant changes by using a double sided t-test. Contour interval is 1 W/m².

Figure 10. The March mean sea ice thickness (shading) and velocity (vector) in the Arctic with an open Bering Strait simulation under different climate boundary conditions. The contour

interval for ice thickness is 0.5 meter. The scale vector for velocity shown in the lower left corner of each panel is 5 cm/s.

Figure 11. The same as in Figure 10, but for September.

Figure 12. The March mean sea ice thickness (shading) and velocity (vector) anomalies in the closed Bering Strait simulations relative to the open Bering Strait simulations. The contour interval is 0.1 meter for the ice thickness. The scale vector shown at the upper right of each panel is 1 cm/s for panels b, c, and d, but 5 cm/s for panels a and e.

Figure 13. The same as in Figure 12, but for September.

Figure 14. Simulated sea surface velocity in the open Bering Strait simulations. The vector unit is cm/s, and the scale is 5 cm/s.

Figure 15. Simulated sea surface velocity difference between the closed Bering Strait (CBS) and open Bering Strait (OBS) simulations. The scale vector is 2 cm/s.

Figure 16. March mean snow depth in the Arctic in the open Bering Strait simulations. The contour interval is 0.1 meter.

Figure 17. The same as in Figure 16, but for September.

Figure 18. The March mean snow depth anomaly in the closed Bering Strait simulation relative to the open Bering Strait simulation. The contour interval is 0.05 meter.

Figure 19. The same as in Figure 18, but for September.

Figures:

Figure 1. The Atlantic meridional streamfunction (AMSF). The left panels are for the open Bering Strait simulations, and the mid-panels are for the closed Bering Strait simulations, and the right panels are the changes of the AMSF between closed and open Bering Strait simulations. PD represents present day; kyr BP represents thousand year before present; NIS represents “no North America ice sheets; and IS represents “with North America ice sheets”. CCSM2 PD or CCSM3 PD represent the simulation using CCSM2 or CCSM3 with present day climate background conditions. 112kyr BP NIS represents CCSM3 simulations under 112 thousand years before present climate conditions without the North America ice sheets. 112kyr IS and 15kyr IS

are the simulations using CCSM3 under 112 and 15 thousand year before present climate background conditions with the North America ice sheets. The same abbreviations are used in all figures.

Figure 2. Percentage changes of the Atlantic meridional overturning circulation (AMOC), Atlantic meridional heat transport (MHT) at 24°N, Pacific MHT at 24°N and 30°S, the liquid freshwater transport (FWT) from Arctic to the North Atlantic, and the FWT in the Pacific at 40°N.

Sea surface temperature anomaly

Figure 3. Surface temperature changes when the Bering Strait is closed under different background climate conditions. The contour interval is 0.1°C . Stippling indicates the statistically significant changes by using a double sided t-test.

Figure 4. Regional mean sea surface temperature (SST), sea surface salinity (SSS), sea surface potential density (SPD), and total surface heat flux changes in the closed Bering Strait simulations relative to the open Bering Strait simulations in the North Pacific (NP, 40-60°N), North Atlantic (NA, 40-80°N), and the Arctic. The units for SST is $^{\circ}\text{C}$, for SSS psu, for SPD kg/m^3 , and for surface heat flux W/m^2 .

Sea surface salinity anomaly

Figure 5. Sea surface salinity changes between the closed and open Bering Strait simulations. The contour interval is 0.1 psu. Stippling indicates the statistically significant changes by using a double sided t-test.

Precipitation anomaly

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Figure 7. Total surface freshwater input, precipitation, evaporation, river runoff, sea ice melt anomaly in the closed Bering Strait simulations relative to the open Bering Strait simulations in the North Pacific (NP, 40-60°N), North Atlantic (NA, 40-80°N), and the Arctic. And the sea ice export anomaly from the Arctic into the North Atlantic in the closed Bering Strait simulations relative to the open Bering Strait simulations. The units for all of these are Sv ($Sv=10^6 m^3/s$).

Surface freshwater flux anomaly

Figure 8. Surface freshwater input anomaly between the closed and open Bering Strait simulations. Stippling indicates the statistically significant changes by using a double sided t-test. Contour interval is 0.05 m/yr.

Surface heat flux anomaly

Figure 9. Annual mean surface heat flux anomaly between the closed and open Bering Strait simulations. Stippling indicates the statistically significant changes by using a double sided t-test. Contour interval is 1 W/m^2 .

March Mean Sea Ice Thickness/Velocity

Figure 10. The March mean sea ice thickness (shading) and velocity (vector) in the Arctic with an open Bering Strait simulations under different climate boundary conditions. The contour interval for ice thickness is 0.5 meter. The scale vector for velocity shown in the upper right corner of each panel is 5 cm/s.

September Mean Sea Ice Thickness/Velocity

Figure 11. The same as in Figure 10, but for September.

March Sea Ice Thickness/Velocity Difference

Figure 12. The March mean sea ice thickness (shading) and velocity (vector) anomalies in the closed Bering Strait simulations relative to the open Bering Strait simulations. The contour interval is 0.1 meter for the ice thickness. The scale vector shown at the upper right of each panel is 1 cm/s for panels b, c and d, but 5 cm/s for panels a and e.

September Sea Ice Thickness/Velocity Difference

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Figure 15. Simulated sea surface velocity difference between the closed Bering Strait (CBS) and open Bering Strait (OBS) simulations. The scale vector is 2 cm/s.

March Mean snow depth on sea ice

CCSM2 PD

CCSM3 PD

112kyr NIS

112kyr IS

15kyr IS

Figure 16. March mean snow depth in the Arctic in the open Bering Strait simulations. The contour interval is 0.1 meter.

September Mean Snow Depth on Sea Ice

CCSM2 PD

CCSM3 PD

112kyr NIS

112kyr IS

15kyr IS

Figure 17. The same as Figure 16, but for September.

March Snow Depth Difference

Figure 18. The March mean snow depth anomaly in the closed Bering Strait simulation relative to the open Bering Strait simulations. The contour interval is 0.05 meter.

September Snow Depth Difference

CCSM2 PD

CCSM3 PD

112kyr NIS

112kyr IS

15kyr IS

m

Figure 19. The same as in Figure 18, but for September.