**Atmospheric circulation anomalies due to 115 kyr BP climate forcing
dominated by changes in the North Pacific Ocean***

Francis O. Otieno¹, David H. Bromwich^{1,2}, and Robert Oglesby³

¹Polar Meteorology Group, Byrd Polar Research Center

The Ohio State University

108 Scott Hall, 1090 Carmack Road

Columbus, OH 43210

²Atmospheric Sciences Program, Department of Geography

The Ohio State University

Columbus, OH

³Department of Earth and Atmospheric Sciences

University of Nebraska

221 Bessey Hall

Lincoln, NE 68588

Submitted to *Climate Dynamics*, Aug 2010. Revised April and June 2011

*Contribution 1389 of Byrd Polar Research Center

Corresponding Author: Francis Ochieng Otieno

E-mail: otieno.1@osu.edu

Tel: (614)-292-6318

Fax: (614) 292-4697

Abstract

Climate at the time of inception of the Laurentide Ice Sheet (LIS) at ~115 kyr BP is simulated with the fully coupled NCAR Community Climate System Model (CCSM3) and compared to a simulated preindustrial climate (circa 1870) in order to better understand land surface and atmospheric responses to orbital and greenhouse cooling at inception. The interaction between obliquity and eccentricity produces maximum decrease in TOA insolation in JJA over the Arctic but increases occur over the tropics in DJF. The land surface response is dominated by widespread summer cooling in the Northern Hemisphere (NH), increases in snowfall, and decreases in melt rates and total precipitation. CCSM3 responds to the climate forcing at 115 kyr BP by producing incipient glaciation in the areas of LIS nucleation.

We find that the inception of the LIS could have occurred with atmospheric circulation patterns that differ little from the present. The location of the troughs/ridges, mean flow over the Canadian Arctic and dominant modes of the atmospheric circulation are all very similar to the present. Larger changes in mean sea level pressure occur upstream of the inception region in the North Pacific Ocean and downstream in Western Europe. In the North Pacific region, the 115 kyr BP anomalies weaken both the Pacific high and Aleutian low making NH summers look more like the PREIND winters and vice versa. The occurrence of cold JJA anomalies at 115 kyr BP favors outbreaks of cold air not in the winter as in contemporary climates but during the summer instead and reinforces the cooling from orbital and GHG reductions. Increased poleward eddy transport of heat and moisture characterizes the atmospheric response in addition to reduced total cloud cover in the Arctic.

Key Words: Laurentide Ice Sheet, Inception, CCSM3, North-Pacific

1.0 Introduction

The inception of the Laurentide Ice Sheet (LIS) over the Canadian Arctic has been the subject of many previous studies. It is now generally accepted that the inception required orbital and GHG forcings acting together along with various feedback processes. Sea ice, snow albedo, sea surface temperature and vegetation changes all play important roles (Khodri et al. 2001; Meisner and Gerdes 2002; Meisner et al. 2003; Kageyama et al. 2004; Desprat et al. 2005; Roe 2006; Crucifix et al. 2006; Kubatzki et al. 2006; Krinner et al. 2006; Schurgers et al. 2007; Calov et al. 2009, Gorbarenko et al. 2010, Kleman et al. 2011). Only a few of these studies attempted to find links between glacial inception and changes in global atmospheric circulation, particularly over the Pacific Ocean. Also most of the studies used simplified models to examine the impacts of orbital, greenhouse gas (GHG) and surface forcings. The earlier results suggested that neither orbital nor GHG forcings alone were sufficient for inception. However, more recent studies that include detailed physics and account for model bias, in for example sea surface temperatures (Vettoretti and Peltier 2003a), are able to simulate perennial snow build-up, a prerequisite for LIS inception.

Debate on the glacial inception problem has been reignited recently by suggestions that humans have via agricultural activities reversed a natural decrease in CO₂ (Ruddiman 2003), causing a warming sufficient to offset glaciation in the near future. Kutzbach et al. (2009) demonstrate differences in CO₂ and CH₄ trends between the current interglacial and previous ones. In support of the anthropogenic hypothesis, Müller and Pross (2007) find that in the absence of anthropogenic greenhouse gas emissions the present insolation has the potential to terminate the Holocene interglacial, but Kutzbach et al. (2009) acknowledge that the present small eccentricity is not sufficient for full glaciation, while others (Berger and Loutre

2002) suggest that glaciation is unlikely for another 50,000 years (50 kyr). An even longer duration before the next glacial is suggested by Archer and Ganoploski (2005) who predict that future carbon release from fossil fuels or methane hydrate deposits of 5000 Gton C could prevent glaciation for a period of ~500 kyr.

Because of the earlier mixed results and limited information regarding the role of atmospheric circulation at the inception of the LIS, Otieno and Bromwich (2009) examined what role if any was played by the interactions between the atmosphere and the land surface. A limited domain centered in the Canadian Arctic was simulated using a land surface model with idealized atmospheric forcing. In the study, the atmospheric forcing (wind speeds, temperature, precipitation and specific humidity) was held constant while the insolation and CO₂ concentrations were modified to conform to the end of the Eemian ~116 kyr before present (116 kyr BP). The study found that even if an atmospheric circulation pattern similar to the one that produced the coldest summer in the last 50 years was maintained perpetually in a simulation, the local reduction in insolation and CO₂ was insufficient for LIS inception. Even in this highly idealized environment, additional cooling of 4° C was still needed for inception. Conversely, no inception occurred when the summer air temperature was specified using representative present-day values even under the perpetual extreme winter snowfall, decreased insolation and lower GHG concentrations. How the widespread cooling could have occurred in reality was not addressed. In this study we expand on the earlier work by examining not only the response of the climate system (land, ocean and atmosphere) on a global scale but also allowing the atmospheric circulation to respond to the forcing that produced the LIS. We demonstrate that a strong atmospheric response occurs over the North Pacific Ocean. Furthermore, changes in atmospheric circulation over the Canadian Arctic are weak and clouds appear to have played a secondary role in the LIS

inception. In order to place our work in perspective we briefly discuss some earlier studies mentioned above.

Since the early 1990s (e.g., Oglesby 1989, 1990, Dong and Valdes 1995 among many others) the role of surface conditions in the LIS inception has been well established both on land and in the ocean. Colder temperatures limiting summer snowmelt and permitting perennial snow are of primary importance. Many studies have since used climate models in various configurations to demonstrate how the cooling is achieved or enhanced at the last glacial inception (Kageyama et al. 1999; Yoshimori et al. 2002; Vettoretti and Peltier 2003a, 2003b, 2004; Hall et al. 2005; Calov et al. 2005; Kubatzki et al. 2006; Kasper et al. 2007; Calov et al. 2009; Otieno and Bromwich 2009). Yoshimori et al. (2002) used two models, a coupled climate model of intermediate complexity and the Canadian Centre for Climate Modeling and Analysis general circulation model (CCCma), to study the atmospheric response to lower boundary conditions at 116 kyr BP. They found that colder surface air temperatures caused by colder sea surface conditions reduce the energy available for melting of snow and also increase the snow fraction in precipitation. This leads to longer periods of snow cover and higher surface albedo which favor the expansion of snow cover as in earlier studies (Oglesby 1990). Earth system models of intermediate complexity (EMIC) studies using dynamic vegetation models forced with Eemian climate (~115 kyr BP) show that retreating boreal forests amplify glacial inception through corresponding increases in albedo (de-Noblet et al. 1996, Calov et al. 2005). Ice-sheet expansion in North America at the end of the Eemian was reduced or even completely suppressed depending on whether pre-industrial or Eemian ocean/vegetation was prescribed (Kubatzki et al. 2006). But the initial size of Greenland was found to have no effect on the expansion of northern hemisphere ice sheets.

Inception is characterized in a number of these studies by either the occurrence of perennial snow as done here or a rapid increase in terrestrial ice volume. However most of these studies focus not on the atmospheric circulation in the North Pacific Ocean but on the surface response over land in the Canadian Arctic (nucleation region). The atmospheric forcing on the land surface processes in many of them is either simplified or specified as in Otieno and Bromwich (2009). The studies are therefore unsuitable for diagnosing changes in atmospheric circulation. Vettoretti and Peltier (2003a, 2003b, and 2004) are among the first to pay close attention to atmospheric processes at the time of the inception. However by limiting their analysis to north of 55°N they excluded important changes in the atmospheric circulation occurring over the North Pacific Ocean. Changes in this region have the potential to impact global climates via teleconnections to the tropical Pacific and have been found at the Last Glacial Maximum (LGM, Yanase and Abe-Ouchi 2010) and in future climate scenarios (Salathe' 2006). Section 2 describes our model and experimental configuration. The results are presented in section 3 followed by discussions and conclusions in section 4.

2. Model and experimental configuration

2.1 The Community Climate System Model

We use the fully coupled Community Climate System Model (CCSM3; Collins et al. 2006a) from the National Center for Atmospheric Research (NCAR) at spectral truncation wave number 42 (T42). At this resolution, the model has a horizontal grid spacing of $\sim 2.8^\circ$ shared by both the land and atmospheric components. The atmosphere and the soil are divided vertically into 26 and 10 levels, respectively, with up to five possible layers in snow cover on land. While a higher spatial resolution is desirable in the Canadian Arctic, we are constrained in our choice of resolution in a fully-coupled-climate-system simulation by available

computing resources and knowledge of boundary conditions. Even more important, however, is the fact that the changes in the North Pacific Ocean are occurring on a very large-scale and therefore resolvable at T42. The evolution of the ocean is simulated using the Parallel Ocean Program (POP; Gent et al. 1998) from Los Alamos National Laboratory and the Community Sea Ice Model (CSIM; Briegleb et al. 2004). These two components share the same variable horizontal resolution (gx1v3), ~ 1 degree spacing near the poles and ~ 0.3 near the equator. The representation of the North Atlantic Meridional Overturning Circulation (MOC) that is important for climate change has also been found to be adequate at T42 (Otto-Bliesner et al. 2006). To our knowledge this is the first time that the fully dynamic CCSM3 model has been used to simulate the Eemian climate and at a resolution higher than is typical in many previous inception studies and paleoclimate modeling in general (Vettoretti and Peltier 2003a, 2003b; Groll et al. 2005).

Details regarding CCSM3 biases that can influence its simulations are discussed in Collins et al. (2006a). For LIS inception, biases in the JJA surface temperature, downwelling radiation, sea ice concentration as well as thickness, and in the ocean heat transport are especially important. A substantial cold bias (-4°C) occurs during JJA over Alaska. The annual mean bias is small but this is due to a compensating warm bias during winter in the model. Over the Canadian Arctic where inception is thought to have nucleated, a smaller cold bias ($\sim 1^{\circ}\text{C}$) exists. The model also has a deficit in the downwelling shortwave radiation, which is enhanced by a higher than observed surface albedo, but counteracted by an excess of downwelling longwave radiation. The zonally averaged ocean heat transport is also slightly higher than that inferred from the NCEP-NCAR Reanalysis, while the atmospheric transports are lower. The sea ice extent compares favorably with satellite estimates but is too thick (up to 3 m) in the Arctic Ocean. Although the

biases are important to consider, whether or not these biases, obtained with respect to present-day climate, remain the same in a 115 kyr BP simulation cannot be determined because the responses are nonlinear. But we have shown previously (Otieno and Bromwich 2009) and will show here that in the absence of a cooling which exceeds the bias in the nucleation region, the model does not spontaneously produce an ice sheet in non-glaciated areas.

2.2 Experimental strategy and configuration

Two main objectives are addressed here. The first is an assessment of how realistic the CCSM3 simulations of the end of the Eemian climate are relative to the earlier studies. Second to demonstrate and diagnose the significant changes in atmospheric circulation that are found upstream of the inception region over the North Pacific Ocean in response to the Eemian climate forcing using a more comprehensive model than in Otieno and Bromwich (2009). Two experiments, one representative of ~115 kyr BP and the other representative of preindustrial conditions were performed. The insolation as well as the concentrations of CH₄ and CO₂ appropriate for each period is specified in CCSM3. In our previous study, the Community Land Model (CLM3) was driven in offline simulations by a specified idealized atmospheric forcing derived from the European Centre for Medium-Range Weather Forecasts reanalysis (ERA-40; Uppala et al. 2005). Although the forcing in the previous study was derived so as to be representative of the climate at the end of the Eemian epoch (~116 kyr BP), the atmosphere was not free to respond to the developing perennial snow. Here the atmospheric forcing on the land surface is interactively derived from the Community Atmosphere Model (CAM3; Collins et al. 2006b) and will be influenced by the coupled dynamics.

2.2.1 Preindustrial control experiment (PREIND)

A number of studies compare paleoclimates with the present with its large amplitude of GHG forcing; this study uses the preindustrial period in order to keep the GHG forcing within the range of natural variability prior to the “anthropocene”. Although the insolation at inception is comparable, the preindustrial GHG concentrations are lower than present but still higher than those at the glacial inception. Our control experiment is derived from an existing preindustrial simulation which is part of a series of multi-century runs conducted by NCAR using CCSM3. The GHG and orbital parameters were held constant at their respective 1870 values (Table 1) in those simulations as the model was integrated forward in a 700-year long simulation. Years 400 to 599 (200 years beyond the point of model quasi-equilibrium discussed later) of this run constitute our control experiment, representing an equilibrium preindustrial simulation (PREIND henceforth).

The snow hydrology parameterization used by CLM3 in the standard CCSM3 configuration capped snow depths at 1000 kg m^{-2} and any excess snow was converted to runoff in order to produce a realistic present-day simulation of seasonal snow cover (Oleson et al. 2004). In preliminary analysis we have found only minimal changes in atmospheric circulation as a result of removing the snow capping in the land surface parameterization, but changes in snow hydrology were significant. We briefly describe the impact that this parameterization change has on snow depths in the results section.

2.2.2 Last glacial inception experiment (BP115K)

The last glacial inception experiment (hereafter BP115K) had 115 kyr BP orbital configuration, CO_2 and CH_4 concentrations. Top of the Atmosphere (TOA) insolation at 115 kyr is specified in CCSM3 using a formulation which depends on a

predefined reference date (Berger and Loutre 1992). The exact concentrations of CO₂ and CH₄ between 130 and 100 kyr BP are not easy to ascertain and range from 240 to 280 for CO₂ (Barnola et al. 1987) and 700 to 400 for CH₄ (Spahni et al. 2005). We choose to use values close to the lowest for this period for both (see Table 1) to enhance the GHG forcing in our experiments. We found that with even lower CO₂ concentration (180 ppmv), the spatial patterns remained the same but the temperature decreases were larger. Larger cooling is also obtained when comparing BP115K with present day (1990) conditions. We note here that some earlier studies did not decrease CH₄ concentrations (Vettoretti and Peltier 2004; Claussen et al. 2005), which would have reduced the overall GHG forcing given the potent impact of CH₄. Other GHG concentrations, specifically those of CFC11 and CFC12, are assumed to be negligible at 115 kyr BP as well as at 1870 because they arise mostly from anthropogenic sources.

Earlier generations of coupled models needed flux adjustments (normally applied to ‘oceanic heat transport’) to prevent model errors in simulating the present-day climate from becoming too large (model drift). Significant improvements in contemporary models have minimized the need for these adjustments, but they still show drift problems. The drift characteristics of CCSM3 arising from slow diffusive processes in the ocean component for short duration runs (~50 years) were presented in Danabasoglu et al. (2006). They found similar globally averaged cooling trends of -0.15 Wm^{-2} after 40 years from two experiments: a default configuration of CCSM3 and an experiment examining the sensitivity to idealized diurnal cycle of solar forcing. We therefore consider 50 years as the minimum duration of integrations necessary to minimize the impact of drift in our results. Even after 700 years in the PREIND there is still some drift hence the “quasi-equilibrium” description in the previous subsection. We start (branch) the

BP115K experiment from PREIND so that both experiments have the same initial state and integrate BP115K forward for a total of 200 simulated years. The first 100 years of BP115K are not used in the analysis in order to achieve comparable drift between BP115K and PREIND. The ideal situation is to have no drift at all but a small drift is not critical for this study. First the drift in the two experiments is comparable after 50 years. Second the ocean response is manifested primarily by large changes near the surface (above 1 km), and is less influenced by the small long-term cooling trend in the deep ocean. Finally, our focus is in the atmospheric and surface changes and these respond to BP115K forcing on a much shorter time scale.

2.3 Equilibrium state of the CCSM3

As a first step in our analysis, we examine the equilibrium state, as characterized by a reduced cooling trend following imposition of the 115 kyr BP perturbation. Figure 1 show twelve-month running global-area-weighted-averages of surface temperatures for 200 years from both BP115K and PREIND. Trend lines have been fitted to the first 100 and the last 100 years of the both experiments. In the first 100 years BP115K cools rapidly at about $0.004 \text{ K per year (yr}^{-1}\text{)}$ in response to the BP115K forcing. The rate decreases to only $\sim 0.002 \text{ K yr}^{-1}$ during the last 100 years indicating that the model is approaching a new equilibrium. The PREIND curve shows that even after 500 years there are still variations that are due to internal model variability. The globally-averaged temperature change over the last 100 years in both experiments remains less than 0.2 K implying that the atmosphere in the coupled climate system is indeed in a quasi-equilibrium state.

2.4 Definition of seasons

We use present-day season names for simplicity but our season definitions are based on the true longitude, the angular location of the earth in its orbit with

respect to the vernal equinox. The location influences the speed of revolution and hence the length of the seasons. Seasonal differences between BP115K and PREIND cannot therefore be computed directly using present day calendar definitions (Crucifix et al. 2006). Defining seasons using the true longitude allows for comparisons in which the earth is within the same segment of its orbit (season). CCSM3 uses a no-leap year calendar; we divided the year into four parts each representing a season. Thus the March-April-May season (MAM: spring) is the 91 days starting March 21st (vernal equinox). The other seasons (summer: June-July-August –JJA; fall: September-October-November-SON; and winter: December-January-February-DJF) are similarly defined so as to follow each other as in the present day calendar. Because of the quarter of a day that results when 365 days is divided by 4, our December-January-February (DJF) season is 92 days long instead of 91 days for convenience.

3. Results and discussion

We will first assess in sections 3.1 through 3.4 how realistic (physically) CCSM3 simulates the Eemian climate using the last 100 years of each simulation. The discussion will be limited to the Northern Hemisphere (NH) because both the LIS and the large changes in the North Pacific Ocean develop only in the NH. A diagnosis of the surface and atmospheric circulation changes over the North Pacific follows in section 3.5 onward. For completeness the smaller changes occurring in the SH will be mentioned in the discussion section. All differences are taken with respect to the preindustrial and thus show how the 115 kyr BP (BP115K) climate differed from that in 1870 (PREIND). Our earlier work suggested that changes in winter and (especially) summer are the most critical for the development of perennial snow; we therefore focus the discussion on these two seasons. In the

second half of this section we diagnose how the anomalous circulation over the North Pacific Ocean is established.

3.1 Surface temperatures

To be suitable for simulating the LIS inception, CCSM3 must show cooling and development of terrestrial perennial snow in the nucleation sites in response to the BP115K forcing. Figure 2 shows the differences (BP115K minus PREIND) in the surface radiative temperature. Clearly, the cooling impact of the BP115K forcing is largest during JJA and exceeds 3.0 K over extensive areas of the Northern Hemisphere (NH) Polar Regions. This reduces the melting of snow in summer and therefore favors the growth of the LIS as has been reported in previous studies (e.g., Oglesby 1990 and many others). Note that the Arctic Ocean remains colder by at least 3.0 K in winter (DJF) as well. A striking seasonal difference is seen over land in the Asian continent ($\sim 90^\circ$ E) during DJF, where the temperatures in BP115K are warmer by up to 2 K as far as 60° N both over land and in the North Pacific Ocean. Farther north surface temperatures remain colder establishing a strong meridional temperature gradient especially over Asia. An even stronger temperature contrast is created in DJF over the North Pacific Ocean by the cooling north of 60° N and warming (> 1.0 K) to the south. Considering that the JJA temperature changes in the North Pacific Ocean are small (Fig. 2a), the results suggest that the mid-latitude regions of the North Pacific Ocean remained relatively warm throughout the year in BP115K. We discuss the impact of this strong temperature gradient anomaly on atmospheric circulation in the North Pacific Ocean in Section 3.3.

3.2 Interplay between orbital and greenhouse gas forcing

In this section we discuss how the 115 kyr BP orbital configuration supports the widespread JJA cooling in the Northern Hemisphere (Fig. 2a), and substantial DJF warming over land areas in Asia and the North Pacific Ocean (Fig. 2b). The

longitude of perihelion at 115 kyr BP differs from that in PREIND by only ~8 degrees; hence the lower obliquity (22.58°) and higher eccentricity (0.04) dominate the orbital forcing at 115 kyr BP (see Table 1). Decreases in summer insolation at the Top of the Atmosphere (TOA) require a smaller obliquity, which tilts the NH less towards the sun in JJA and an increased eccentricity which increases the aphelion distance. At 115 kyr BP, the two reinforced each other both during JJA when the net effect reduced TOA insolation and during DJF when increases occurred in the sub-polar latitudes of the NH. The reductions in GHG concentrations however produce cooling everywhere and year round.

The influence of 115 kyr BP orbital configuration on TOA insolation is summarized by Fig. 3. Reductions (up to 25 Wm^{-2}) in insolation between April and August coincide with what is otherwise a time of maximum NH isolation and extend into the snow melt season. But it is not all reductions; there is a short period in August-September when the Polar Region received $10\text{-}15 \text{ Wm}^{-2}$ more insolation in BP115K than in PREIND. From October through March a substantial increase in TOA insolation also occurs in the NH but it is largest in the tropics. To explain these differences, consider the role of obliquity and eccentricity separately. In both PREIND and BP115K NH summer occurs near the aphelion minimizing the impact of the longitude of perihelion on the TOA insolation. On the other hand, lower obliquity in BP115K points the NH less towards the sun during summer compared PREIND (See Table 1). Coupled with a higher eccentricity which increases the aphelion distance, the two reinforce each other to cool the NH during JJA and warm it in DJF. However unlike in summer the combined effect is smaller in the Polar Region because little if any insolation reaches these higher latitudes in winter.

In summary, the orbital changes at 115 kyr BP impact the NH more because they coincide with the peak in summer insolation. In JJA the largest reductions

occur north of 60°N (*Arctic Region*) but in DJF the largest increases are centered at approximately 5°S (*Tropical Region*). Since the land areas over Asia and the North Pacific Ocean show increased surface temperatures in DJF, the response to orbital forcing dominates that from reduced GHG concentrations. Near the North Pole the impact of reduced GHG concentrations becomes dominant because little insolation reaches these latitudes in DJF. The synergy between orbital and greenhouse gas effects provides a mechanism for seasonal and regional amplification of orbital forcing during the NH summer and is critical for ice sheet inception at 115 kyr BP. Its absence coupled with warm biases in the models may have had a bearing on the difficulties in simulating glacial inception in earlier studies using obliquity or eccentricity forcing alone. Although some studies (Cubasch et al. 2006) suggest that orbital forcing alone may be sufficient for inception, proxy records indicate decreases in both insolation and GHG concentrations as applied here occurred at 115 kyr BP.

3.3 Radiative fluxes and clouds

Without atmospheric influence, changes in insolation of equal magnitude to those at the TOA discussed above would occur at the surface yet the downwelling solar flux at the surface depicted in Fig. 4a shows that this does not happen. Therefore the atmosphere in BP115K does modulate the changes in TOA radiation noted above. In the Arctic JJA downwelling solar flux increases at the surface in BP115K compared to PREIND. Because NH JJA TOA insolation decreased (Fig. 3), the only way for the surface fluxes to be higher is if the atmosphere becomes more transparent to the incident shortwave radiation. Such a scenario can occur from changes in optical properties of clouds or in the amounts of water vapor and clouds present in the atmosphere. We diagnosed changes in the latter using vertically integrated total (low, medium and high) cloud cover.

Most regions north of 60° N show significant (in a student t-test, $p < 0.05$, stippled) decreases in total cloud amount (Fig. 4b). Therefore the Arctic increases in downwelling solar fluxes (Fig. 4a) result primarily from decreases in cloud fraction in BP115K. The decrease in downwelling solar fluxes over land areas in the Mid-latitudes can also be explained by the increases in cloud fraction over the same regions. The increase in downwelling solar flux over Baffin Island accompanied by increases in cloud fraction (Fig. 4b) at first appears counter intuitive. However, the development of perennial snow discussed next can lead to multiple reflections of shortwave radiation between the land and bottom of the clouds through a positive feedback involving snow albedo. Wendler et al. (2004) found that the presence of a highly reflecting surface under overcast conditions can increase incoming global radiation by 85% due to multiple reflections. However, the net solar flux (Fig. 4c) difference at the surface remains negative over much of the NH and therefore favors development of perennial snow on land.

3.4 Perennial snow cover and depth

We pointed out that in PREIND the snow hydrology parameterization in CCSM3 limited snow depths to 1000 kg m^{-2} . The main impact of removing this limit is found in a steady increase in snow depths in areas having glaciers in the initial conditions in the Community Land Model (CLM3). Other fields and the atmospheric circulation in particular are largely insensitive to the limit removal, thus demonstrating that the PREIND can be used to assess the BP115K experiment in spite of the snow-depth parameterization issue. However, differences in snow depths over Greenland are overestimated because the depth was limited in PREIND but not in the BP115K experiment. Because expansion of the LIS requires growth in the areas without existing glaciers, the PREIND snow depths are still useful over such areas. The time evolution of area-averaged August snow depths over the LIS

nucleation region bound by latitudes 55° and 70° N and longitudes 120° to 60° W, chosen to exclude most of Greenland for the aforementioned reason, is shown in Fig. 5. After ~30 years a substantial increase in perennial snow depths develops over this area and continues to grow throughout the integration.

Figure 6 show the spatial distribution of averaged seasonal snow height for JJA and SON, seasons which are typically associated with higher melt rates and no accumulation in the present climate. It is evident that CCSM3 does indeed develop perennial snow in response to the BP115K forcing over Baffin Island, Alaska and eastern Siberia (Fig. 6a and 6c) but no accumulation of snow occurs in PREIND over the same areas (Fig. 6b and 6d). Therefore in spite of the uncertainty in the snow depths over glaciated regions in the PREIND (discussed above), the temperature and snow depths in this study confirm that CCSM3 responds to the climate forcing at ~115 kyr BP by producing incipient glaciation in the areas of LIS nucleation. The occurrence of simulated perennial snow in Alaska shown here is contrary to some other studies and uncertain from the proxy records. Vettoretti and Peltier (2003b) attribute the absence in their study to increased atmospheric transport of latent and sensible heat into the region. A later explanation (Krinner et al. 2006) suggests a lowering of snow albedo during the melt season by dust deposition. We think that increased snowfall coupled with reduced melt rates (Section 3.5) and the snow depth parameterization in CLM3 that is discussed earlier are responsible for the excessive snow thickness in Alaska. In addition the model has a substantial cold bias over Alaska during JJA (section 2.1).

3.5 Snowfall and snow melt

Decreases in accumulated precipitation occur over the Arctic north of 45° N in JJA (Fig.7a). During both JJA (Fig. 7a) and DJF (Fig.7b) the accumulated precipitation increases over the North Pacific Ocean, but changes over Baffin Island

and large areas of the Canadian Arctic do not exceed 5 mm. In spite of the decreases in accumulated precipitation, significant increases in snowfall occur in JJA (Fig. 7c) and are consistent with a change in phase of precipitation in the generally colder conditions discussed earlier. Thus at 115 kyr BP a larger fraction of the decreased precipitation falls as snow (Fig. 7c). Note the striking resemblance between the areas of increased JJA snowfall (Baffin Island, Alaska and northeastern Russia) and those that show perennial snow in Fig. 6a and 6c. Although most present-day snow falls in DJF in the NH, the DJF increases in BP115K are small. However, the season of increased snowfall is less important because the NH Polar Regions remained cold throughout the year (Fig. 2). Furthermore in Otieno and Bromwich (2009) we also found that extreme winter snowfall alone was insufficient for inception, a point also noted in the seminal work of Oglesby (1990).

Decreased JJA snowmelt exceeds snowfall by nearly an order of magnitude in all the regions that develop perennial snow (Fig. 7e). Over the nucleation region ~20 mm of additional snow falls in BP115K but reduced melting leaves behind ~200 mm. The growth of the LIS can therefore be explained primarily by a change in snow hydrology resulting from increased summer snowfall and decreased snow melt induced by the summer cooling. Over northern Asia between 60° and 75° N statistically significant increases in snow melt in JJA contrasts sharply with the decreases seen over Baffin Island, Alaska and Siberia. Colder temperatures are critical because they increase snowfall while reducing melt rates by keeping the NH around the freezing point. We show later (Fig. 9) that much of the nucleation area remains north of the zero line while most of Northern Asia is south of it.

3.6 Sea ice

An additional consequence of the colder temperatures in the NH is an expansion of sea ice coverage that is depicted in Fig. 8. The fraction of the surface covered by sea ice does not change substantially in the central Arctic in both JJA and DJF. However statistically significant increases occur in the North Pacific and North Atlantic, indicating that sea ice is getting thicker at the margins in response to 115 kyr BP forcing. Because the Arctic Ocean is colder in BP115K in both seasons, the southward flowing Oyashio current favors expansion of sea ice into the Bering Sea, but a warm surface anomaly in DJF (Fig. 2b) dominates the area of the Kuroshio, a warm ocean current originating in the tropical Pacific Ocean. We hypothesize that these warmer DJF temperatures limited the southward expansion of the sea ice setting up a strong meridional temperature gradient capable of altering the atmospheric circulation in the North Pacific. Over the North Atlantic DJF increases in temperature are modest and therefore exert only a moderate influence on the meridional temperature gradients.

3.7 Circulation changes

We have shown that the temperature and snow accumulation response to 115 kyr BP forcing in CCSM3 is consistent with the earlier studies. We now turn to what we consider the most interesting finding of this study; that a drastic change in atmospheric circulation over the Canadian Arctic is not necessary for inception under 115 kyr BP forcing and that the largest changes in atmospheric circulation occur instead over North Pacific Ocean. Otieno and Bromwich (2009) suggested that a change in atmospheric circulation may be necessary for the additional cooling needed for LIS inception. We diagnose the circulation differences using monthly mean sea-level pressure (MSLP), the daily mean surface pressure and its variance, 500 hPa geopotential heights, and wind speeds at the lowest model level (992 hPa). A comparison of the spatial patterns in the mean flow (ridge/trough) and the large-

scale stationary eddies over the nucleation region are essentially the same between BP115K and PREIND and are not shown. Therefore changes in the regional circulation over the Canadian Arctic are likely to have played a minor role in the inception of the Laurentide Ice Sheet.

The vectors which represent the magnitude and direction of the changes in the 100-year mean zonal and meridional wind speeds at each model grid point, show changes in mean flow over the North Pacific that are nearly twice as large as over the nucleation region. Table 2 summarizes the flow differences over the North Pacific (40-90° N and 150° E to 120° W) and Canadian Arctic (40-90° N and 120° W to 30° W) in both JJA and DJF. During JJA, the critical season for inception, the magnitudes of the change vector over the North Pacific Ocean is twice that in the nucleation area. In DJF on the other hand the changes in mean flow resulting from 115 kyr BP forcing are nearly equal. The anomalies are reversed between DJF and JJA, consistent with the sea-level pressure anomalies discussed next.

3.7.1 Sea-level pressure

Anticyclonic anomalies occur over both the nucleation region and in Siberia during JJA (Fig. 10a) in response to 115 kyr BP forcing. In the North Pacific Ocean on the other hand a cyclonic anomaly weakens the Pacific anticyclone. These changes in MSLP support the reversal in mean flow anomalies depicted above by the change vector (Fig. 9). The increase in MSLP over NH land areas together with decrease in the North Pacific represents a mass transfer from ocean to land during JJA. With this configuration the cold polar air masses would be directed southwards, expanding the influence of the BP115K cold anomalies. In the North Pacific Ocean this would expand the area of sea ice as we have already seen. The JJA MSLP anomalies over the North Atlantic are small (less than 1.5 hPa) compared to those in the North Pacific Ocean.

During DJF the differences in MSLP over the North Pacific are even more pronounced with increases exceeding 4 hPa south of the Bering Strait compared to a decrease of less than 2 hPa in the nucleation region (Fig. 10b). The larger anticyclonic anomaly in the North Pacific draws cold polar air southward over western North America and warm tropical maritime flow northward towards East Asia. This configuration is consistent with the DJF differences in both the mean air flow (Fig. 9) and surface air temperatures anomalies (Fig. 2). Warmer temperature anomalies occur where the air flows northward and vice versa. Note the near exact opposite patterns in MSLP anomalies between JJA and DJF. The strong cyclonic anomaly that developed in the North Pacific Ocean ($\sim 45^\circ$ N and 190° W) in JJA is now replaced with an anticyclonic anomaly in DJF ($\sim 45^\circ$ N and 160° W). Figure 10 shows that in the North Pacific region, the 115 kyr BP anomalies weaken both the Pacific high and Aleutian low making NH summers look more like the PREIND winters and vice versa. This has significant implications for cold temperature anomalies in NH at 115 kyr BP. In contemporary investigation of the amplification of the Siberian high, Takaya and Nakamura (2005) found that cold air outbreaks follow when anomalously surface cold air reaches the northeastern slope of the Tibetan Plateau. Therefore the occurrence of cold JJA anomalies at 115 kyr BP favors outbreaks of cold air not in the winter as in contemporary climates but during the summer instead and further reinforces the orbital and GHG cooling. This additional mechanism was lacking in the specified atmospheric forcing in Otieno and Bromwich (2009) and can explain why an additional cooling was necessary in that study but not here.

Finally, Fig. 11 shows the spatial patterns of the daily surface pressure variance (associated with the passage of storms) between BP115K and PREIND. In JJA increases in variance exceeding 40 hPa^2 occur only in the North Pacific Ocean,

while changes over the North Atlantic and the nucleation region are again relatively small. The increased variance suggests increased frequency of JJA storm activity in the North Pacific Ocean. While not as large, increased activity extends into Hudson Bay. But during DJF not only does the enhanced variance stretch from the North Pacific across western Canada into the nucleation region; the activity is also shifted poleward in the North Pacific Ocean. This suggests more vigorous winter storms were influencing the nucleation region at 115 kyr BP. We examined changes in the eddy meridional transport of heat and moisture during JJA and DJF (Fig. 12). The figure shows increases between 30 and 60° N during JJA. In DJF an increase moisture transport by transient eddies occurs in the same latitude band but increased heat transport occurs further north (centered at 60° N), consistent with the poleward shift in surface pressure variances (Fig. 11). While increased moisture transport supports the increased snowfall (Fig. 7c), increase in eddy transports is essentially a negative feedback. Specifying the atmospheric forcing in Otieno and Bromwich (2009) prevented such increases in eddy transport.

The change in surface pressure variance and eddy transports suggest that our imposed 115 kyr BP forcing can also alter high latitude climate variability. We examined the Northern Annular Mode (NAM), North Atlantic Oscillation (NAO) and Pacific Decadal Oscillation (PDO) using Empirical Orthogonal Functions (EOFs) with no rotation for DJF. The spatial patterns of the first dominant SLP (26.5° N to 90° N) mode associated with NAM look very similar (spatial correlation 0.90) and explain 40 and 30 % of the temporal variance in BP115K and PREIND respectively. The dominant NAO patterns (26° N to 80° N and 70° W to 40° E) are also very similar with corresponding explained variances of 60 and 47%. Over on the North Pacific where the PDO is dominant, we also find spatial structures from monthly SST's from BP115K and PREIND that are very similar and explain 39 and

33 % of the variance respectively. A frequency analysis of the NAO index indicates a shift towards lower frequencies but our simulated time series is too short for a meaningful analysis of variations in PDO frequency because of its longer period (~30 years).

We find dominant modes in both BP115K and PREIND DJF that are similar to each other. Hall et al. (2005) examined the atmospheric variability in a general circulation model for the last 165 kyr BP. They found as we have, that in JJA the atmospheric response is primarily thermal. The similarity in spatial patterns of the dominant DJF modes (associated with atmospheric variability), together with the small JJA changes in the mean flow over the Canadian Arctic supports our finding that the atmospheric circulation need not have changed substantially for inception at 115 kyr BP. While the spatial patterns are similar, the temporal variability which can be impacted by both internal model variations and the imposed forcing is different at 115 kyr BP. A detailed analysis of the cause of changes in NAO frequency is beyond the scope of the present study.

3.7.2 Upper level circulation

The sharp meridional temperature gradient that develops in the North Pacific Ocean influences atmospheric circulation in layers far above the surface via the thermal wind effect. Zonal cross sections of averaged zonal wind speed (Fig. 13) in the West (120° E to 180° W and East (180° W to 120° W) Pacific Ocean as well as the Atlantic Ocean (60° W to 0° E) show large changes in NH zonal wind. Westerlies increase near 40° N in both the West and East Pacific Ocean in JJA from the surface to 100 hPa. These increases ($>3 \text{ ms}^{-1}$) are collocated ($\sim 40^\circ \text{ N}$) with the core of strong westerlies ($>15 \text{ ms}^{-1}$) implying that no shift in the location of the core occurs. The pattern of increases in the North Atlantic is however very different; the largest increases are centered near 12 km above the equator. During DJF by contrast the

decreases in zonal wind at the core of the mean westerlies are accompanied by increases poleward in both the Pacific and Atlantic Oceans. The adjacent reductions and increases in wind speeds suggest a poleward shift in the center of westerlies to near 50 °N.

To explain the JJA intensification and DJF shift, consider the surface temperature patterns (Fig. 2) again. In JJA the area of greatest cooling ($> 3.0\text{K}$) expands to cover much of the area north of 40° N. Although regions south of 40° N also cool down somewhat, the greater cooling to the north enhances the gradients and strengthens the westerlies with no significant shift in their locations (Fig. 13; top panel). We postulate that a stronger meridional temperature gradient in the West Pacific Ocean is created by the strong contrast between warm water associated with the Kuroshio Current and colder anomalies of the Oyashio Current emanating from the Arctic. Over the East Pacific Ocean (Fig. 13; top middle) the cold temperature anomalies extending southward (Fig.2a) weakens the gradient, this leads to weaker westerlies. A sharp meridional temperature gradient does not develop in the North Atlantic Ocean and changes in westerlies there are not substantial ($<1 \text{ ms}^{-1}$). During DJF, the core of westerlies shows a northward shift in both the North Pacific and Atlantic Oceans (Fig. 13, bottom panel). We hypothesize that as the area of cooling migrates poleward in DJF (Fig. 2) so does the core of westerlies that depend on the horizontal temperature gradient.

Figure 14 shows the changes in zonal flow at 200 hPa (level of the subtropical jet stream core) are not limited to the Pacific and Atlantic Oceans. During JJA the westerlies are stronger over Asia and show no shift. Over Canada they are weaker and changes in the nucleation region are less than $\pm 1 \text{ ms}^{-1}$. The area of strongest zonal winds is shifted to the tropics in JJA over North America. The figure confirms that no major changes in westerlies occur over the North Atlantic in

JJA at this level. Changes in easterlies are evident in equatorial Africa and extend across the Atlantic Ocean. Anomalous westerlies in the central Atlantic increase by more than 3 ms^{-1} and are responsible for the equatorial positive anomalies in Fig. 13. In DJF on the other hand (Fig. 14b) westerlies intensify over southern Alaska and also over the nucleation region. The increases over the North Pacific are accompanied by decreases centered near 30°N consistent with a northward shift.

These changes in upper level westerlies and in surface flow associated with the MSLP shown earlier confirm that few if any changes occur in the atmospheric circulation over the LIS inception region in JJA. The changes in atmospheric circulation due to 115 kyr BP forcing are therefore dominated by those occurring in the North Pacific Ocean and are driven primarily by the horizontal temperature gradients which are enhanced by the expansion of sea ice. This finding is also consistent with the conclusion of Hall et al. (2005) that the meridional temperature gradients in the NH are key to understanding model response to orbital forcing.

3.8 Southern hemisphere and ocean responses

The simulated changes in surface temperature and sea ice concentrations over the Southern Hemisphere (SH) are not as pronounced as those in the NH and are therefore not shown. Low latitude land areas (Australia, South America, and South Africa) show colder JJA anomalies and are slightly warmer in DJF. Unlike the NH, the decrease in TOA insolation does not coincide with Austral summer season (Fig. 3) but instead occurs in austral spring. Note that maximum insolation increase during DJF (austral summer) occurs near the equator. Also the eccentricity and obliquity effects counter each other in the SH. During boreal summer when the earth is near the aphelion, the obliquity tilts the SH more towards the sun (increasing TOA insolation) but the higher eccentricity increases the aphelion distance (decreasing insolation) and vice versa. In the presence of large oceanic areas with

higher thermal capacity, the net effects on the SH of the 115 kyr BP forcing are therefore subdued.

Analysis of the oceans shows that the response in BP115K is limited to depths above 1 km. More (0.5 petawatts) poleward heat transport occurs in the Atlantic compared to the Pacific Ocean. Variations within the basin result in averaged anomalies that are very small but in the North Pacific Ocean temperature anomalies in the area of the Kuroshio Current exceed 2° C. The temperature profiles in the Atlantic suggest an unstable stratification, with cold above warmer water. Salinity changes in the North Pacific Ocean however make the basin more stable (less dense near the surface). The enhancement of the global meridional overturning circulation north of 30 ° N above depths of 2000 m and the small changes in average ocean conditions do not appear to have any obvious impact on glacial inception.

4.0 Conclusions and Discussion

The NCAR fully coupled community climate system model at T42 is used to simulate the climate at 115 kyr BP in a 200-year long simulation. The last 100 years of the simulation is compared to a preindustrial run from NCAR with 1870 orbital forcing and preindustrial CH₄ and CO₂ concentrations. To make seasonal comparisons the true longitude of the earth is used to define the seasons. In spite of the inadequate parameterization in the CLM3 land model during the preindustrial run, which limits snow depths over glaciers, CCSM3 is capable of simulating the cooling and development of perennial snow over the nucleation region of the LIS. The main findings can be summarized as follows:

1. Most of the Northern Hemisphere land responds to lower greenhouse gas concentrations and reduced top of the atmosphere insolation at 115 kyr BP by cooling down in JJA. The region where surface cooling exceeds 3.0 K (needed for inception) is more widespread in JJA compared to DJF when the

temperatures are marked by warm anomalies south of 60° N. A marked meridional temperature gradient occurs over the North Pacific but not the Atlantic Ocean, which strengthens the westerlies in JJA and also shifts them poleward in DJF. We hypothesize that weak horizontal temperature gradient occurs in the North Atlantic because the Gulf Stream is able to penetrate farther north into the Norwegian and Barents Seas in the North Atlantic Ocean while the Kuroshio Current in the North Pacific is constrained to a more zonal direction by southward expansion of sea ice and the cold Oyashio current.

2. The climate response to 115 kyr BP forcing varied by season and region. Eccentricity and obliquity reinforce each other in the Northern Hemisphere in both summer and winter but oppose each other in the Southern Hemisphere during both seasons. Coincident NH reduction in top of the atmosphere insolation with peak summer months produces maximum cooling in the Arctic Region in JJA. Within each hemisphere, contrasts are influenced by the thermal capacity of the surface. Land areas, and in particular the Asian continent shows the greatest sensitivity to the BP115K forcing due to large size and lower heat capacity.
3. In Otieno and Bromwich (2009) we found that local cooling was insufficient for inception. With an interactive atmosphere, the initial cooling from orbital and GHG forcing is amplified by increased snowfall which becomes perennial on land and expanded sea ice cover (almost into the North Pacific). The enhanced land cooling favors cold air outbreaks in summer in contrast to contemporary climates in which outbreaks occur in the winter. The outbreaks and increased winter storm activity reinforce the 115 kyr BP forcing and allow the atmospheric anomalies to influence a much wider area of the Northern Hemisphere. CCSM3 shows that decreased snow melt was of primary

importance for the development of the Laurentide Ice Sheet (LIS). The ice sheet expands southward over the Canadian Arctic but not in Asia because temperatures in Asia, although colder at 115 kyr BP are still above freezing in the summer. Perennial snow develops only in areas that remain below freezing all year round. These results provide a physical mechanism for the occurrence of the idealized conditions that were specified in our earlier study (perpetual extreme winter snowfall and summer cold anomalies).

4. Inception of the LIS does not require a major reorganization in the mean circulation over the Canadian Arctic. Inception occurs under mean circulation patterns that do not differ much from the present day configuration, at least in the basic Rossby wave patterns. However eddy transports become more vigorous, enhancing poleward transport of heat and moisture and are in essence negative feedbacks. The largest atmospheric circulation changes occur upstream of the nucleation region of the Laurentide Ice Sheet in the North Pacific Ocean. Both the Aleutian low and Siberian high are considerably weaker in BP115K winter compared to PREIND. The JJA anomalies make the BP115K summers more like the preindustrial winters, allowing cold Arctic air to influence larger areas of the Northern Hemisphere especially in the North Pacific Ocean where sea ice expands south of Bering Strait.
5. We conclude that the North Pacific Ocean is a key place to be watching for significant climate change signals, because of both large present day anomalies in sea ice and the occurrence of larger circulation anomalies in other drastically different climates (past and future) including at 115 kyr BP. Yanase and Abe-Ouchi (2010) found somewhat different pressure anomalies in the North Pacific Ocean at the Last Glacial Maximum, with a strengthened Aleutian low (DJF, opposite to here) and a weakened Pacific High (JJA, same as here). Francis and

Hunter (2007) found that contemporary anomalies in the zonal wind component significantly impact the sea ice edge in the Bering Sea that we have associated with development of strong horizontal temperature gradients at 115 kyr BP. Even in global warming scenarios the Aleutian Low and associated storm tracks appear to intensify (Salathé 2006). The record 2007 summer sea ice loss was greatest over the Pacific side of the Arctic and the loss was associated with anomalous circulation (L'Heureux et al. 2008)

6. The simulated anomalous circulations over the North Pacific Ocean and Asia can be explained using basic concepts, but have not been previously reported by other modelers or supported by paleorecords. A paleoclimate model inter-comparison study with the same 115 kyr BP conditions (similar to PMIP3) could demonstrate that these anomalies are not unique to CCSM3. The main challenge is to conclusively determine if these anomalies did indeed occurred at 115 kyr BP. This can only be done by comparing simulations with quality paleorecords and falls in the purview of the paleodata collectors not the climate modeling community. The number of paleo-records in the North Pacific is almost an order of magnitude smaller than in the North Atlantic Ocean. Harada et al (2008) show a cold anomaly at 115 kyr BP in the Okhotsk Sea but key to explaining the CCSM3 circulation anomalies is the warm sea surface temperatures in the area of the Kuroshio Current. Our results also show clearly a need for verification on the seasonal time scale, therefore we encourage the paleodata community to develop other proxy records with seasonal-scale resolution that can be used in model verification especially for the area centered near 40° N in the North Pacific Ocean.

Acknowledgments

This research was funded by National Science Foundation grant OPP-0352865. We are grateful to the National Center for Atmospheric Research (NCAR) for providing the technical support needed to optimize CCSM3 on the high performance computing platforms and other resources provided through the Scientific Computing Division (SCD-36091015). We also thank the three anonymous reviewers for their comments and suggestions that helped to substantially improve the manuscript.

References

- Archer D, Ganoploski A (2005) A movable trigger: Fossil fuel CO₂ and the onset of the next glaciation. *Geochem Geophys Geosys* 6, Q05003, doi:10.1029/2004GC000891
- Barnola JM, Raynaud D, Korotkevich YS, Lorius C (1987) Vostok ice core provides 160,000-year record of atmospheric CO₂. *Nature* 329:408-414 doi:10.1038/329408a0
- Berger A, Loutre MF (2002) An exceptionally long interglacial ahead? *Science* 297:1287-1288 doi:10.1126/science.1076120
- Berger A, Loutre MF (1992) Astronomical solutions for paleoclimate studies over the last 3 million years. *Earth and Planet Sc Lett* 111:369-382
- Briegleb BP, Blitz CM, Hunke EC, Lipscomb WH, Holland MM, Schramm JL, Moritz RE (2004) Scientific description of the sea ice component in the Community Climate System Model Version 3. Technical Note NCAR/TN463+STR, National Center for Atmospheric Research
- Calov R, Ganopolski A, Kubatzki C, Claussen M (2009) Mechanisms and time scales of glacial inception simulated with an Earth system model of intermediate complexity. *Clim Past Discuss* 5:245-258
- Calov R, Ganopolski A, Petoukhov V, Claussen M, Greve R (2005) Transient simulation of the last glacial inception. Part I: Glacial inception as a bifurcation of the climate system. *Clim Dyn* 24:545-561. doi:10.1007/s00382-005-0007-6
- Claussen M, Brovkin V, Calov R, Ganopolski A, Kubatzki C (2005) Did humankind prevent a Holocene glaciation? Comment on Ruddiman's hypothesis of a pre-historic Anthropocene. *Clim Change* 69:409-417. doi:10.1007/s10584-005-7276-2

- Collins WD, Bitz CM, Blockmon ML, Bonan GB, Bretherton CS, Carton JA, Chang P, Doney SC, Hack JJ, Henderson TB, Kiehl JT, Large WG, McKenna DS, Santer BD, Smith RD (2006a) The Community Climate System Model, version 3 (CCSM3). *J Clim* 19:2122-2143
- Collins WD, Rasch PJ, Boville BA, Hack JJ, McCaa JR, Williamson DL, Briegleb BP, Bitz CM, Lin S-J, Zhang M (2006b) The formulation and atmospheric simulation of the Community Atmosphere Model version 3 (CAM3). *J Clim* 19:2144-2161
- Crucifix M, Loutre MF, Berger A (2006) The climate response to astronomical forcing. *Space Sci Rev* 125:213-226
- Cubasch U, Zorita E, Kaspar F, Gonzalez-Rouco JF (2006) Simulation of the role of solar and orbital forcing on climate. *J Adv Space Physics* 37:1629-1634
- Danabasoglu G, Large WG, Tribbia JJ, Gent PR, Briegleb BP, McWilliams JC, (2006) Diurnal Coupling in the Tropical Oceans of CCSM3. *J Climate* 19: 2347-2365. doi: 10.1175/JCLI3739.1
- de Noblet, NI, Prentice IC, Joussaume S, Texier D, A. Botta A, Haxeltine A (1996) Possible role of atmosphere-biosphere interactions in triggering the Last Glaciation. *Geophys Res Lett* 23:3191-3194, doi:10.1029/96GL03004.
- Desprat S, Sanchez-Goni MF, Turon JL, McManus JF, Loutre MF, Duprat J, Malaize B, Peyron O, Peypouquet J-P (2005) Is vegetation responsible for glacial inception during periods of muted insolation changes? *Quaternary Sci Rev* 24:1361-1374
- Dong B, Valdes, PJ (1995) Sensitivity studies of Northern Hemisphere glaciation using an atmospheric general circulation model. *J Climate* 8:2471-2496

- Francis JA, Hunter E (2007) Drivers of declining sea ice in the Arctic winter: A tale of two seas. *Geophys Res Lett.* doi:10.1029/2007GL030995
- Gent PR, Bryan FO, Danabasoglu G, Doney SC, Holland WR, Large WG, McWilliams JC, (1998) The NCAR Climate System Model global ocean component. *J Clim* 11:1287-1306
- Gorbarenko, SA, Harada N, Malakhov MI, Vasilenko YP, Bosin AA, Goldberg EL (2010) Orbital and millennial-scale environmental and sedimentological changes in the Okhotsk Sea during the last 350 kyr. *Global and Planetary Change*, 72:79-85
- Groll N, Widmann M, Jones J, Kaspar F, Lorenz S (2005) Simulated relationships between regional temperatures and large-scale circulation: 125 kyr BP (Eemian) and the preindustrial period. *J Clim* 18:4032-4045
- Hall A, Clement A, Thompson DWJ, Broccoli A, Jackson C (2005) The importance of atmospheric dynamics in the northern hemisphere winter climate response to changes in the earth's orbit. *J Clim* 18:1315-1325
- Harada, N., M. Sato, and T. Sakamoto (2008) Freshwater impacts recorded in tetraunsaturated alkenones and alkenone sea surface temperatures from the Okhotsk Sea across millennial-scale cycles. *Paleoceanography*, 23:1-14. doi:10.1029/2006PA001410
- Kageyama M, Charbit S, Ritz C, Khodri M, Ramstein G (2004) Quantifying ice-sheet feedbacks during the last glacial inception. *Geophys Res Lett.* doi:10.1029/2004GL021339
- Kageyama M, Valdes PJ, Ramstein G, Hewitt C, Wyputta U (1999) Northern Hemisphere storm tracks in present day and Last Glacial Maximum climate

- simulations: A Comparison of the European PMIP Models. *J Clim* 12:742-760
- Kasper F, Spangehl T, Cubasch U (2007) Northern Hemisphere winter storm tracks of the Eemian interglacial and the last glacial inception. *Clim. Past*, 3:181-192 doi:10.5194/cp-3-181-2007
- Khodri M, Leclainche Y, Ramstein G, Braconnot P, Marti O, Cortijo E (2001) Simulating the amplification of orbital forcing by ocean feedbacks in the last glaciation. *Nature* 410:570-574. doi:10.1038/35069044
- Kleman J, Jansson K, De Angelis H, Stroeve AP, Hättestrand C, Alm G, Glasser N (2011) North American Ice Sheet build-up during the last glacial cycle, 115–21 kyr. *Quaternary Sci Rev* 29:2036-2051
- Krinner G, Boucher O, Balkanski Y (2006) Ice-free glacial northern Asia due to dust deposition on snow. *Clim Dyn* 27:613–625. doi:10.1007/s00382-006-0159-z
- Kubatzki C, Claussen M, Calov R, Ganopolski A (2006) Sensitivity of the last glacial inception to initial and surface conditions. *Clim Dyn* 27:333-344. doi:10.1007/s00382-006-0136-6
- Kutzbach JE, Ruddiman WF, Vavrus SJ, Philippon G (2009) Climate model test of anthropogenic influence on greenhouse-induced climate change (early agriculture to modern): the role of ocean feedbacks. *Clim Change* 99:351-381
- L'Heureux ML, Kumar A, Bell GD, Halpert MS, Higgins RW (2008) Role of the Pacific-North American (PNA) pattern in the 2007 Arctic sea ice decline. *J Geophys Res Lett* 35:doi:10.1029/2008GL035205

- Meisner KJ, Weaver AJ, Matthews HD, Cox PJ (2003) The role of land surface dynamics in glacial inception: a study with the UVic earth System model. *Clim Dyn* 21:515-537
- Meisner KJ, Gerdes R (2002) Coupled climate modelling of ocean circulation changes during ice age inception. *Clim Dyn* 18:455-473
- Müller UC, Pross J (2007) Lesson from the past: present insolation minimum holds potential for glacial inception. *Quaternary Sci Rev* 26:3025-3029. doi:10.1016/j.quascirev.2007.10.006
- Oglesby RJ (1989) A GCM study of Antarctic glaciations. *Clim Dyn* 3:135-156
- Oglesby RJ (1990) Sensitivity of glaciation to initial snow cover, CO₂, snow albedo, and oceanic roughness in the NCAR CCM. *Clim Dyn* 4:219-2351
- Oleson KW, Dai Y, Bonan G, Bosilovich M, Dickinson R, Dirmeyer P, Hoffman F, Houser P, Levis S, Niu G-Y, Thornton P, Vertenstein M, Yang Z-L, Zeng X (2004) : Technical description of the Community Land Model (CLM), NCAR Technical Note NCAR/TN-461+STR
- Otieno FO, Bromwich DH (2009) Contribution of atmospheric circulation to inception of the Laurentide Ice Sheet at 116 kyr BP. *J Clim* 22:39–57
- Otto-Bliesner BL, Tomas R, Brady EC, Ammann C, Kothavala Z (2006) Climate sensitivity of moderate- and low-resolution versions of CCSM3 to preindustrial forcings. *J Clim* 19:2567-2583
- Roe G, (2006) In defense of Milankovitch. *J Geophys Res.* doi:10.1029/2006GL027817
- Ruddiman W (2003) The anthropogenic greenhouse era began thousands of years ago. *Climatic Change* 61:261-293

- Salathe' EP Jr, (2006) Influences of a shift in North Pacific storm tracks on western North American precipitation under global warming. *Geophys Res Lett*. doi:10.1029/2006GL026882
- Schurgers G, Mikolajewicz U, Grogger M, Maier-Reimer E, Vizcaíno M, Winguth A (2007) The effect of land surface changes on Eemian climate. *Clim Dyn* 29:357-373
- Spahni R, Chappellaz J, Stocker TF, Loulergue L, Hausammann G, Kawamura K, Flückiger J, Schwander J, Raynaud D, Masson-Delmotte V, Jouzel J (2005) Atmospheric methane and nitrous oxide of the late Pleistocene from Antarctic ice cores. *Science* 310:1317-1321
- Takaya K, Nakamura H (2005) Mechanisms of intraseasonal amplification of the cold Siberian High. *J. Atmos. Sci.* 62:4423–4440
- Uppala SM, Kallberg PW, Simmons A J, Andrae U, Bechtold VD, Fiorino M, Gibson JK, Haseler J, Hernandez A, Kelly GA, Li X, Onogi K, Saarinen S, Sokka N, Allan RP, Andersson E, Arpe K, Balmaseda MA, Beljaars ACM, Van De Berg L, Bidlot J, Bormann N, Caires S, Chevallier F, Dethof A, Dragosavac M, Fisher M, Fuentes M, Hagemann S, Holm E, Hoskins BJ, Isaksen I, Janssen PAEM, Jenne R, McNally AP, Mahfouf JF, Morcrette JJ, Rayner NA, Saunders RW, Simon P, Sterl A, Trenberth KE, Untch A, Vasiljevic D, Viterbo P, Woollen J (2005) The ERA-40 re-analysis. *Q J Roy Meteor Soc* 131:2961-3012
- Vettoretti G, Peltier WR (2003a) On Post-Eemian glacial inception. Part I: the impact of summer seasonal temperature bias. *J Clim* 16:889-911
- Vettoretti G, Peltier WR (2003b) On Post-Eemian glacial inception. Part II: Elements of a cryospheric moisture pump. *J Clim* 16:912-927

- Vettoretti G, Peltier WR (2004) Sensitivity of glacial inception to orbital and greenhouse gas climate forcing. *Quaternary Sci Rev* 23:499-519
- Wendler G, Moore B, Hartmann B, Stuefer M, Flint R (2004) Effects of multiple reflection and albedo on the net radiation in the pack ice zones of Antarctica. *J Geophys Res* 109. doi:10.1029/2003JD003927
- Yanase W, Abe-Ouchi A (2010) A Numerical study on the atmospheric circulation over the midlatitude North Pacific during the Last Glacial Maximum. *J Clim* 23:135-151
- Yoshimori M, Reader MC, Weaver AJ, McFarlane NA (2002) On the causes of glacial inception at 116 ka BP. *Clim Dyn* 18:383–402. doi:10.1007/s00382-001-0186-8

Table 1: Experimental configuration

	115 kyr BP (BP115K)	Pre Industrial (PREIND)
Obliquity	22.58 Degrees	23.45 Degrees
Eccentricity	0.0414	0.0167
Longitude of Perihelion	290.9	282.7
CO₂	250 ppm	289 ppm
CH₄	500 ppbv	901 ppbv
N₂O	281 ppbv	281 ppbv
CFC11	0	0
CFC12	0	0
Vegetation	PD	PD
Elevation	PD	PD
SST	POP	POP
Duration	200 yrs	200 yrs

PD- Present-day; POP-Parallel Ocean Program

Table 2: Seasonal changes in mean flow over the North Pacific (40-90° N and 150° E to 120° W) and Canadian Arctic (40-90° N and 120° W to 30° W). The mean change is defined as the magnitude of the vector difference between BP115K and PREIND at the lowest model level.

	JJA		DJF	
	N. Pacific	Canada	N. Pacific	Canada
Mean change (m/s)	0.68	0.34	0.84	0.78

Fig. 1: Global-area-weighted 12-month running averages of surface air temperature (K) and linear trends for the first and last 100 years of BP115K (red) and PREIND (black) The temperature change in the last 100 years is ~ 0.2 K in both experiments but internal model variability results in an upward trend in PREIND while BP115K is still cooling.

Fig. 2: BP115K minus PREIND June-July-Aug (JJA) and (left) December-January-February (DJF; right) surface temperature (K). Summer differences are largest in the polar regions of the Northern Hemisphere. Asia and much of the North Pacific are warmer in NH winter compared to the preindustrial. The latitudes are every 20° starting at the equator.

Fig. 3 Incident solar radiation at the top of the atmosphere in PREIND (contours) and difference (BP115K minus PREIND, colors). PREIND contours are drawn at 25 Wm⁻² intervals.

Fig. 4: (a) Downwelling surface solar flux (b) vertically integrated cloud fraction and (c) net solar flux at the surface in PREIND (contours; every 0.2) along with difference between BP115K and PREIND (colors). Stippled areas have significant ($p < 0.05$) differences.

Fig.5: August area-weighted (over the LIS nucleation region 55° N-70° N and 120° W-60° W) mean snow height (m) for BP115K (red curve) and PREIND (black curve). After 30 years of integration substantial perennial snow occurs in the BP115K run marking inception in these simulations.

Fig. 6: Seasonal averages (100 years) of snow height (m) for summer (a) and (b), and fall (c) and (d) from the BP115K experiment (left panels) and the PREIND (right panels). Perennial snow is indicated by the accumulations in summer and fall seasons in BP115K. Spatial patterns of BP115K winter and spring season distributions are very similar (not shown) to the left panels.

Fig. 7: Differences (color) in seasonally accumulated total precipitation (top), snowfall (middle) and snow melt (bottom) during JJA (left) and DJF (right). Contours are the seasonal means in PREIND (mm). Stipples show areas where the differences are statistically significant at $p < 0.05$. Latitudes are spaced at 15° starting at the North Pole.

Fig. 8: Differences (color) in NH fraction of the surface covered by sea ice (sea ice varies from 0 to 1) in the summer and winter. Sea increases around the margins in both JJA and DJF. The red and blue contours mark the 0.1 fraction in BP115K and PREIND respectively.

(a) JJA Difference vectors and BP115K Surface temperature

(b) DJF Difference vectors and BP115K Surface temperature

Fig. 9: Vectors representing the change in mean zonal and meridional near surface (992mb) wind flow between BP115K and PREIND. Temperatures (K) in the area enclosed by the red contour are below freezing at BP115K. The reference vector represents 2 ms⁻¹.

Fig. 10: Mean preindustrial sea-level pressure (contours) and the BP115K minus PREIND differences (colors) for (a) JJA and (b) DJF. Stippled areas have significant differences at $p<0.05$.

Surface pressure

Fig. 11: Daily surface pressure variances (contours) for JJA (top) and DJF (bottom)

and the difference in variances (BP115K minus PREIND; color). Due to BP115K forcing the surface pressure variance increases indicating more storm activity in the Canadian Arctic in DJF.

Fig. 12: Zonally averaged moisture (left) and heat (heat) transport differences between BP115K and PREIND. The atmosphere responded to 115 kyr BP forcing by increasing both moisture and heat transport (negative feedback).

Fig. 13: Zonally averaged zonal wind separated by regions; West Pacific Ocean (120°E to 180°W), East Pacific (180°W to 120°W) and the Atlantic (60°W to 0°E). BP115K wind anomalies are colored. Line contours are the mean PREIND wind (m/s). The alternating pattern over all northern oceans suggests a poleward shift during winter in the region of highest zonal wind.

Fig. 14: Mean PREIND zonal wind speed at 200 hPa (contours) in JJA (upper) and DJF (lower). The differences in the mean (BP115K minus PREIND) are overlaid as colors.