True to Milankovitch: Glacial Inception in the new Community Climate System Model.

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Abstract: The equilibrium solution of a fully coupled general circulation model with present day orbital forcing is compared to the solution of the same model with the orbital forcing from 115,000 years ago. The difference in snow accumulation between these two simulations has a pattern and a magnitude comparable to the ones inferred from reconstructions for the last glacial inception. As postulated in Milankovitch’s hypothesis the only necessary positive feedback is the snow albedo feedback, which is initiated by reduced melting of snow and sea-ice in the summer. The ocean meridional heat transport in the two simulations is almost identical, and the atmospheric meridional heat transport and the sea-ice/low-cloud connection constitute negative feedbacks that almost fully compensate for the snow albedo feedback. The fact that the realistic difference in snow accumulation is achieved with the same model that is used for the fifth assessment report builds trust in the ability of climate models to anticipate the evolution of climate in the future.
1 Introduction

Over the last 400,000 years Earth’s climate went through several glacial and interglacial cycles, which are correlated with changes in its orbit and associated changes in insolation (Milankovitch 1941; Hays et al. 1976; Huybers and Wunsch 2004). Over this time changes in global ice volume, temperature and atmospheric CO$_2$ have been strongly correlated with each other (Petit et al. 1999). The connection between temperature and ice volume is not surprising, and a strong snow/ice-albedo feedback makes it plausible to connect both to changes in the Earth’s orbit (Huybers and Tziperman 2008, and references therein).

Models of intermediate complexity (e.g., Crucifix and Loutre 2001; Khodri et al. 2001; Wang and Mysak 2002; Khodri et al. 2003; Calov et al. 2005) and flux-corrected GCMs (Groeger et al. 2007) have typically been able to simulate a connection between orbital forcing and temperature or snow volume. So far, however, free running, fully coupled primitive equation General Circulation Models (GCMs) have failed to reproduce glacial inception, the cooling and increase in snow and ice cover that leads from the warm interglacials to the cold glacial periods. Milankovitch (1941) postulated that the driver for this cooling is the orbitally induced reduction in Northern Hemisphere summer time insolation and the subsequent increase of perennial snow cover. The increased perennial snow cover and its positive albedo feedback are, of course, only precursors to ice sheet growth. The GCMs failure to recreate glacial inception (see Otieno and Bromwich 2009 for a summary) indicates a failure of either the GCMs or of Milankovitch’s hypothesis. Of course, if the hypothesis would be the culprit, one would have to wonder if climate is sufficiently
understood to assemble a GCM in the first place. Either way, it appears that re-
producing the observed glacial/interglacial changes in ice volume and temperature
represents a good testbed for evaluating the fidelity of some key model feedbacks
relevant to climate projections. Note that from the GCM perspective glacial incep-
tion and glacial termination are not symmetric, because for the latter one needs
an ice sheet model, the development of which is still in its infancy (see Vizcaino et
al. 2008 for the description of a state of the art model), whereas the current GCMs
should contain all the necessary components of the climate system to reproduce the
former.

The potential causes for GCMs failing to reproduce inception are plentiful, rang-
ing from numerics (Vettoretti and Peltier 2003) on the GCMs side to neglected feed-
backs of land, atmosphere, or ocean processes (e.g.; Gallimore and Kutzbach 1996,
Hall et al. 2005, Jochum et al. 2010, respectively) on the theory side. It is encour-
aging, though, that for some GCMs it takes only small modifications to produce
an increase in perennial snowcover (e.g.; Dong and Valdes 1995). Nevertheless, the
goal for the GCM community has to be the recreation of increased perennial snow-
cover with a GCM that has been tuned to present day climate, and is subjected to
changes in orbital forcing only.

The present study is motivated by the hope that the new class of GCMs that
has been developed over the last 5 years (driven to some extent by the forthcoming
fifth assessment report, AR5), and incorporates the best of the climate community’s
ideas, will finally allow us to reconcile theory and GCM results. It turns out that
at least one of the new GCMs is true to the Milankovitch hypothesis. The next
section will describe this GCM and illustrate the impact of changing present day insolation to the insolation of 115,000 years ago (115 kya). At least to the present authors it was surprising that increased perennial snow cover was achieved with only the ice/snow albedo feedbacks postulated by Milankovitch, no other feedbacks seem to be necessary. In particular the stability of the Atlantic Meridional Over-turning Circulation (AMOC) under changes to the insolation is surprising, and will be discussed in Section 3. Section 4 analyzes in detail the Arctic heat budget with its multitude of positive and negative feedbacks. Section 5 concludes the study with a summary and implications for future model development.

2 Model Description and Results

The numerical experiments are performed using the National Center for Atmospheric Research (NCAR) latest version of the Community Climate System Model (CCSM) (version 4), which consists of the fully coupled atmosphere, ocean, land and sea ice models. A description of this version can be found in Gent et al. (2011). The ocean component has a horizontal resolution that is constant at 1.125° in longitude and varies from 0.27° at the equator to approximately 0.7° in high latitudes. In the vertical there are 60 depth levels; the uppermost layer has a thickness of 10 m, the deepest layer has a thickness of 250 m. The atmospheric component uses a horizontal resolution of 0.9° × 1.25° with 26 levels in the vertical. The sea ice model shares the same horizontal grid as the ocean model and the land model is on the same horizontal grid as the atmospheric model. The details of the different model components will be described in a series of forthcoming papers; for the present
purpose it is sufficient to know that CCSM4 is a state-of-the-art climate model that has improved in many aspects from its predecessor CCSM3 (Gent et al. 2011). For the present context the most important improvement is probably the increased atmospheric resolution and the switch from a spectral to a finite volume dynamical core, because both of these improvements allow for a more accurate representation of altitude and therefore land snow cover (e.g.; Vettoretti and Peltier 2003).

The subsequent sections will analyze and compare two different simulations: an 1850 control (CONT), in which earth’s orbital parameters are set to the 1990 values and the atmospheric composition are fixed at its 1850 values; and a simulation identical to CONT, with the exception of the orbital parameters, which are set to the values of 115 kya (OP115). CONT was integrated for 700 years, and OP115 was branched off from CONT after year 500 and continued for 300 years. Unless noted otherwise the comparisons will be done between the means of years 651-700 of CONT and the years 751-800 of OP115. Although CONT is considered fully spun-up the comparison is not quite clean, because there is a small drift in ocean temperature and salinity. The two different time intervals were chosen because at year 700 the AMOC in OP115 did not reach equilibrium yet, and we did not have the computational resources to extend CONT beyond year 700. However, in the present context the drift is negligible because it is mostly confined to the abyssal ocean (Danabasoglu et al. 2011).

The particular time period for the orbital forcing of OP115 is chosen because the solar forcing at 65°N during June differs from its present value by approximately 7%, more than during most other epochs (Berger and Loutre 1991). More-
over, observations suggest that the onset of the last ice age dates around this time (Petit et al. 1999; Capron et al. 2010). The shift in orbital parameters means that in particular at northern high latitudes there is less insolation during spring and early summer months, and more radiation later in the fall (Figure 1). This shift in the seasonal distribution of insolation is at the heart of the inception hypothesis of Milankovitch (1941), because it leads to increased snow and sea-ice cover during early summer, and a positive snow/ice albedo feedback. The two main goals of the present work are to demonstrate that the 115 kya orbital changes produce a realistic increase in snow cover in CCSM4, and to quantify the relevant climate feedbacks.

The difference in orbital parameters between OP115 and CONT lead to a warmer tropical band, and cooler northern high latitudes (Figure 2a). There is only little response in the southern high latitudes, and these are mostly confined to the ocean. To isolate the impact of the ocean on glacial/interglacial dynamics, we replaced the full ocean model with a slab-ocean model (for the detailed methodology, please see Danabasoglu and Gent 2009). The results are quite similar (Figure 2b; snow and sea-ice differences are similar too, but not shown), with the full ocean leading to a slightly weaker response. The exception is the Labrador Sea, which shows approximately $2^\circ$C more cooling with a full ocean model. While tropical and southern hemisphere differences in both sets of experiments are similar to the coarse resolution results of Jochum et al. (2010, JPLM from here on), the northern high-latitude response is significantly weaker. The relative weakness at high latitudes can be attributed to the different response of the AMOC, and will be focus of
The colder northern high-latitudes in OP115 are associated with significant increase in snow depth in the Canadian Archipelgo, and northern Siberia (Figure 2c), regions cited as the origination points for the glacial ice sheets in ice sheet reconstructions (Svendsen et al. 2004; Kleman et al. 2010). The increase in snow depth will be analyzed in section 4, but it is worth noting here that it has been achieved without modifying the underlying GCM. The CONT simulation is identical to the 1850 control simulation of Gent et al. (2011), which has been optimized to produce the observed 20th century climate. The fact that CONT can produce a realistic snow distribution for a very different mean climate is a major achievement for the CCSM, and instills confidence in the results of its future projections.

In accordance with the Milankovitch (1941) hypothesis (see also Huybers and Tziperman 2008), it can be seen that the reduced spring/summer insolation in OP115 leads to more perennial snow, and larger summer sea-ice concentration (Figure 2d), both of which contribute to the positive snow/ice albedo feedback. Section 4 will provide a quantification of the positive and negative feedback processes leading to the northern high latitude cooling.

3 The role of the AMOC and the Labrador Sea gyre

The meridional heat transport of the AMOC is a major source of heat for the northern North Atlantic ocean (e.g.; Ganachaud and Wunsch 2001), but it is also believed to be susceptible to small perturbations (e.g.; Marotzke 1990). This raises the
possibility that the AMOC amplifies the orbital forcing, or even that this amplification is necessary for the northern hemisphere glaciations and terminations (e.g.; Broecker 1998). In fact, JPLM demonstrates that at least in one GCM changes in orbital forcing can lead to a weakening of the MOC and a subsequent large northern hemisphere cooling. Here, we revisit the connection between orbital forcing and AMOC strength with the CCSM4, which features improved physics and higher spatial resolution compared to JPLM.

It turns out that for CCSM4 the OP115 scenario does not lead to a reduced AMOC in contrast to the JPLM results, which show a 30% reduction of the AMOC strength under OP115 forcing. After branching off at year 500 of CONT, the AMOC in OP115 immediately weakens, and continues to do so until it reaches a minimum after approximately 40 years (Figure 3a). It stays weaker for some 50 years, starts to recover around year 600, and reaches a mean value of 15.3 Sv (yrs 751-800), which is not significantly different from the mean of CONT. The depth of the AMOC has not changed significantly either (not shown). The meridional heat transport shows a similar evolution (Figure 3b), with final values of 0.74 PW. The strength of the subpolar gyre, too, mirrors this decline and subsequent recovery (to 52.7 Sv, Figure 3c). By the end of this 300 year development, the surface of the Irminger Sea and most of the subsurface subpolar Atlantic is warmer and saltier than in CONT, and the surface of the Labrador Sea is colder and fresher (Figure 3d).

Much of this reorganization of the subpolar Atlantic can be explained by the analysis of sea-ice extent and maximum ocean boundary layer depth, the latter being by construction the depth of a year’s deepest convective event (Large et al.
In CONT convective activity can be seen all along the edge of the sea-ice and south of Iceland (Figure 4a). In OP115 the different orbital forcing leads initially to more extensive sea-ice in the Labrador Sea (Figure 4b), because the melting season is shortened (Figure 1). The Irminger Sea, however, is under the influence of the warm North Atlantic Drift, which leads to only minor changes in the sea ice. The increase in sea-ice concentration insulates the Labrador Sea from atmospheric forcing, and therefore inhibits convective activity and reduces the strength of the subpolar gyre (Figure 3c). South of Iceland, without the effects sea-ice, the orbital forcing leads merely to a cooling of the surface water, and therefore a destabilization of the water column (Figure 4c) and deeper convection (Figure 4b). The intermediate depth warming seen in Figures 4c and 3d is the direct result of the weaker Labrador Sea gyre: the weaker gyre leads to a poleward migration of the Gulf Stream and the North Atlantic Drift (Figure 4d), with its associated influx of more spicy (warmer and saltier) water. For water of equal density, the atmosphere removes buoyancy more efficiently from spicier water (Jochum 2009), so that after 300 years the convective activity in the subpolar gyre is quite similar between CONT and OP115 (not shown), albeit in a different background state (Figure 3d). Thus, the connection between subpolar gyre strength and subtropical spiciness advection acts as a negative feedback that stabilizes the gyre.

Another source of North Atlantic Deep Water is convection in the Norwegian Sea, which enters the subpolar gyre through the Denmark Strait and Faroe Bank Channel overflows. In CONT, this is a 5.2 (± 0.4) Sv contribution to the AMOC. This value is determined by a parameterization and depends largely on the density
differences between the water masses on either sides of the ridges (Danabasoglu et al., 2010). North of the ridges, the changes in salinity and temperature largely compensate each other so that the maximum boundary depth and density changes are minor (not shown). Thus, the transient response is determined by the behaviour of the subpolar gyre, which leads to an initial cooling of the waters on the Atlantic side of the ridge, and then a recovery as the gyre and the AMOC strengthen (Figure 4c). The cooling leads to minimum overflow of 4.1 Sv after 120 years (about 40 years after the minimum in AMOC and gyre strength) and then recovers towards a transport of 5.4 Sv in years 751-800 (not shown). Thus, the variability of the overflow strength, too, is controlled by the subpolar gyre.

In principle, the two mechanisms described above should have come into play in JPLM as well. There, however, an increased freshwater export out of the Arctic led to a halocline catastrophe (see Bryan 1986 for a general discussion of this process). It turns out that in CCSM4 with its finer spatial resolution there is a crucial third process that allows the AMOC to recover: the freshwater flux through the Baffin Bay (Figure 5). An analysis of the subpolar freshwater budget shows that its largest anomalies are the liquid freshwater transport through the Nares Strait and Northwest Passage (via the Baffin Bay), and the sea-ice import through the Fram Strait (Figure 6): Immediately after changing the orbital forcing, the import of sea-ice through the Fram Strait increases, and continues to increase for another 100 years. This is also happening in the coarse resolution version and leads to the halocline catastrophe (JPLM). In the present experiment, however, a good part of the increased ice import is compensated for by reduced inflow of liquid freshwater into
the Baffin Bay. The strength of the flow into Baffin Bay is determined by the sea
surface height difference between the Arctic and the Labrador Sea (Prinsenberg
and Bennett 1987; Kliem and Greenberg 2003; Jahn et al. 2010). This difference
is reduced from 70 cm in CONT to 50 cm during the years 551-600 in OP115 (not
shown).

Thus, there are two negative feedbacks by which the effect of orbital forcing on
the AMOC is minimized. Both work through the subpolar gyre: Firstly, increased
sea-ice cover reduces its strength; this brings in spicier subtropical water, which is
more susceptible to convection; and secondly, the reduced gyre strength leads to a
reduced pressure difference between the Arctic and the Labrador Sea, thereby re-
ducing the import of freshwater through the Nares Strait and Northwest Passage.

To what extent the present feedbacks are relevant in the real world will depend
on an accurate representation of the Arctic and subpolar physics. A detailed dis-
cussion of model performance and biases is provided in Danabasoglu et al. (2011);
here we will limit ourselves to a quantification of some of the key processes. As
in the observations (Dickson and Brown 1994; Vage et al., 2009), deep water for-
mation occurs in the Irminger, Labrador and the Greenland Seas with maximum
boundary layer depths of approximately 1000 m in the latter two, and 800 m in
the former. However, deep convection is not observed south of Iceland and this is
a major bias in the model. The product of the Denmark Strait and Faroe Bank
Channel overflow water is well reproduced with 5.2 Sv when compared to the re-
cent observational estimates of 6.5 Sv (Girton and Sandford 2003; Mauritzen et al.
2005). The shutdown of the Labrador Sea convection in OP115 leads to 1 Sv reduc-
tion in the AMOC, which is consistent with recent evidence that the contribution of Labrador Sea convection to the AMOC is less than 2 Sv (Pickart and Spall 2007). With 19.5 Sv the strength of the AMOC at 26.5 N is close to the observed value of 18.7 Sv (Kanzow et al. 2009). The freshwater budget and the sea-ice properties of the Artic Ocean are also well represented in CCSM4 (Jahn et al., 2011, to be submitted). This general assessment of the model performance is encouraging, but it should be kept in mind that the response of the AMOC to disturbances can be critically dependent on poorly constrained mixing processes (e.g.; Schmittner and Weaver 2001; Saenko et al. 2003).

Observational evidence for the past AMOC strength comes from proxy tracers obtained from deep ocean cores for the Last Glacial Maximum (LGM, 21 kya). The most widely used tracers to infer ocean circulation patterns during the LGM are carbon isotopes ($\delta^{13}C$), Cadmium-Calcium ratios ($Cd/Ca$) in benthic foraminifera, Protactinium-Thorium ratios ($Pa/Th$) in benthic foraminifera, and estimates of the density gradient across the Atlantic basin based on oxygen isotope profiles ($\delta^{18}O$). Measurements of sortable silts have been used to infer rates of past deep-ocean flows. While for each of these tracer studies exist that indicate a weaker or shallower AMOC during the LGM (e.g.; Lynch-Stieglitz et al. 2007; Gherardi et al. 2009; Lynch-Stieglitz et al. 2006; McCave et al. 1995), the more recent application of rigorous statistical tools suggests that the available database is currently not sufficient to reject the hypothesis of an unchanged AMOC during the LGM (e.g.; Gebbie and Huybers 2006; Marchal and Curry 2008; Peacock 2010). The present authors are aware of only one observation based analysis of the AMOC strength.
that stretches past 115 kya, and it suggests that the AMOC started weakening only several thousand years after the beginning of the last glacial inception (Guichhou et al. 2010). Thus, the present model result of 115 kya orbital forcing having no impact on the strength of the AMOC is consistent with the available observations.

4 The Atmospheric Response

The pattern and amplitude of wind stress and precipitation response is similar to the one in the coarse resolution study of JPLM, involving minor changes with the exception of a stronger Indian Summer Monsoon and stronger westerlies over the North Pacific (not shown). In particular the zonally averaged wind stress over the Southern Ocean is identical to within 0.5 %, and its maximum is at the same latitude. As already illustrated in Figure 2 the main differences between OP115 and CONT are in the Arctic and will be analyzed here.

The difference in orbital forcing between OP115 and CONT leads to larger incoming radiation at the top of the atmosphere (TOA) in the tropics, and smaller radiation at high latitudes in the former compared to the latter (Figure 7a), with the total incoming solar radiation being 0.3 $W/m^2$ larger in OP115 than in CONT. Our focus will be on the northern hemisphere north of 60°N, which covers the areas of large cooling and increased snow cover (Figure 2). Compared to CONT, the annual average of the incoming radiation over this Arctic domain is smaller in OP115 by 4.3 $W/m^2$ (black line), but the large albedo reduces this difference at the TOA to only 1.9 $W/m^2$ (blue line, see also Table 1). This is only a minor change, and the core piece of the Milankovitch hypothesis is how this signal is spread across the
seasonal cycle and amplified (e.g.; Huybers and Tziperman 2008): reduced summer insolation (Figure 1) prolongs the time during which land is covered with snow, and ocean covered with ice (Figure 2d). In CCSM4 this larger albedo in OP115 leads to a TOA clearsky shortwave radiation that is 8.6 $W/m^2$ smaller than in CONT (red line; clearsky radiation is diagnosed during the simulation by omitting the clouds from the radiation calculation) - more than four times the original signal. The snow/ice albedo feedback is then calculated as 6.7 $W/m^2$ ($8.6 W/m^2 - 1.9 W/m^2$). Interestingly, the low cloud cover is smaller in OP115 than in CONT, reducing the difference in total TOA shortwave radiation by 3.1 $W/m^2$ to 5.5 $W/m^2$ (green line). Summing up, an initial forcing of 1.9 $W/m^2$ north of 60°N, is amplified through the snow/ice albedo feedback by 6.7 $W/m^2$, and damped through a negative cloud feedback by 3.1 $W/m^2$.

The regions of TOA shortwave difference are confined to areas of increased snow and sea-ice (compare Figure 7b with Figure 2d). The negative feedback of the clouds is mostly due to a reduction of low-cloud cover over the ocean during summer (Figure 7c). High latitude stratus clouds have a similar effect as snow or ice: they reflect sunlight. Unlike at lower latitudes, Arctic low cloud formation via coupling with the ocean is frequently inhibited by sea ice and surface temperature inversions. While the presence of open water in the Arctic does not guarantee atmosphere-ocean coupling and low cloud formation, open water has been shown to increase Arctic cloud cover when atmosphere-ocean coupling is strong (Kay and Gettelman 2009). Since the Arctic Ocean in OP115 is colder and has more sea ice cover than in CONT, especially in summer, reductions in low cloud amount are not
surprising, and serve to counteract the positive ice-albedo feedback from increased
sea ice cover.

Because of the larger meridional temperature (Figure 2a) and moisture (not
shown) gradient, the lateral atmospheric heatflux into the Arctic is increased from
2.88 to 3.00 PW. This 0.12 PW difference translates into an Arctic average of 3.1
$W/m^2$, a negative feedback as large as the cloud feedback, and ten times as large as
the changes to the ocean meridional heat transport (Figure 3b). Thus, the negative
feedback of the clouds and the meridional heat transport almost compensate for
the positive albedo feedback, leading to a total feedback of only 0.5 $W/m^2$. One way
to look at these feedbacks is that the climate system is quite stable, with clouds and
meridional transports limiting the impact of albedo changes. This may explain why
some numerical models have difficulties creating the observed cooling associated
with the orbital forcing (e.g.; Jackson and Broccoli 2003).

Ultimately, of course, a successful simulation of the inception does not necessar-
ily need cooling, but an increased snow and ice cover to build ice sheets. In principle
the increased snow depth seen in Figure 2b could be due to increased snow fall or
reduced snow melt. The global moisture budgets reveal that OP115 has a larger
poleward moisture transport than CONT (not shown), which is consistent with the
increased heat transport. This does lead to increased snow fall, but in contrast to
the results of Vettoretti and Peltier (2003) this is negligible compared to the reduc-
tion in snow melt (Figure 7d). The global net difference in melting and snowfall
between OP115 and CONT leads to a snow built-up, the volume increase of which
is equivalent to a sea-level drop of 20 m in 10.000 years. This is less than the 50 m
estimate based on sea-level reconstructions between present day and 115 kya (e.g.; Lambeck and Chappell 2001), but nonetheless it suggests that the model response is of the right magnitude.

5 Summary and Discussion

It is shown here that the same CCSM4 version that realistically reproduces the present day climate, also shows, when subjected to orbital forcing from 115 kya, increased areas of perennial snow in exactly the areas where glacial reconstructions put the origin of the glacial ice sheets. Furthermore, the difference in snow deposition between OP115 and CONT is of the same order of magnitude as the reconstructions based on sea level data. CCSM4 achieves this by the most basic mechanism, which was already postulated by Milankovitch (1941): reduced northern hemisphere summer insolation reduces snow and sea-ice melt, which leads to larger areas with perennial snow and sea-ice, and increased albedo. This, in turn, further reduces melting, leading to the positive snow/ice-albedo feedback.

This positive feedback is opposed and almost compensated for by increased meridional heat transport in the atmosphere, and by reduced low cloud cover mainly over the Arctic ocean. The former is on solid theoretical grounds (e.g.; Stone 1978) and has support from other GCM studies (e.g.; Shaffrey and Sutton 2006), but the physics of low Arctic clouds is one of the weak points of current GCMs (e.g.; Kay et al. 2011). This does not mean that the CCSM4 response of Arctic clouds to insolation is necessarily wrong. The paucity of observations relevant for sea-ice/cloud feedbacks means, however, that there have been only few opportunities to
test this aspect of the model.

An equally large source of uncertainty is the stability of the AMOC. To our knowledge the only other inception study with a full, freely evolving GCM is JPLM, and they find a 30% reduction of the AMOC strength under 115 kya solar forcing. The present GCM has better physics and higher resolution, so it is tempting to claim that its results are more realistic. However, the fact that CCSM4 is able to recreate a reasonable inception scenario without an active ocean removes the 115 kya scenario as a testbed for the ocean model. As discussed in Section 3, the observations are not conclusive either, nor is there a comprehensive theory for the AMOC strength (see Kuhlbrodt et al. 2007 for a recent review). Thus, from the present set of experiments we can only conclude that ocean feedbacks are not necessary for glacial inception, but we cannot rule out that the ocean did play a role.

While the increase in perennial snow cover and snow deposition is encouraging, the present study is only a first step toward reproducing the glacial inception in a GCM. It still remains to be shown that the OP115 climate allows for the growth of the Scandinavian, Siberian and Laurentide ice sheets. More importantly, though, the lack of any significant southern hemisphere polar response needs explaining (Figure 2). While Petit et al. (1999) suggests that Antarctica cooled by about $10^\circ$C during the last inception, some studies point to the substantial uncertainties of these ice-core based temperature measurements (e.g.; Stenni et al. 2010). There is, however, clear evidence that during the LGM the Southern Ocean sea-ice cover was much more extensive (Gersonde et al. 2005), so that the GCMs should at least show some increased sea-ice concentrations during glacial inception. Unfortunately, the
present day simulation of CCSM4 does already have a significant excess of sea ice 
(Landrum et al., to be submitted to J. Climate), due at least in part to its excessively strong Southern Ocean winds (Holland and Raphael 2006). Arguably this leads to a reduced sensitivity to reduced springtime insolation, so that one focus of future model development will have to be improvement of the overly strong Southern Ocean winds - a problem which, strangely enough, has been unresolved since Boville (1991).

Thus, while the present study finally provides numerical support for the Milankovitch hypothesis of glacial inception, it also identifies two foci of future research: Firstly, the sensitivity of Arctic clouds to climate fluctuation needs to be validated and possibly improved upon; and secondly, the Southern Ocean sea-ice concentration and the strength of the zonal winds need to become more realistic.

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References


Table 1: Summary of the annual mean heat flux feedbacks north of 60°N.

<table>
<thead>
<tr>
<th>Process</th>
<th>Strength [W/m²]</th>
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<tbody>
<tr>
<td>insolation forcing</td>
<td>+ 4.3</td>
</tr>
<tr>
<td>× present day albedo</td>
<td>- 2.4</td>
</tr>
<tr>
<td>snow/ice albedo feedback</td>
<td>+ 6.7</td>
</tr>
<tr>
<td>low-cloud/sea-ice feedback</td>
<td>- 3.1</td>
</tr>
<tr>
<td>merid. heat transp. feedback</td>
<td>- 3.1</td>
</tr>
<tr>
<td>total forcing</td>
<td>+ 2.4</td>
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</tbody>
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Figure 1: Seasonal cycle of insolation at the top of the atmosphere for CONT (black) and OP115 (red) at 70°N (maximum in June) and 70°S. In OP115 the June insolation is 38 W/m² weaker than in CONT.

Figure 2: Difference in annual mean surface temperature between OP115 and CONT (a), and between their respective slab-ocean counterparts (b). c: Difference in annual mean snow depth between OP115 and CONT, d: area where the summer (minimum) snow depth is larger than 1 cm and the summer sea-ice concentration is larger than 40% for CONT (red), and its increase in OP115 (blue),

Figure 3: a: Strength of the AMOC (in Sv) at 50°N for CONT (black) and OP115 (red). The straight black lines denote the mean and mean ± one standard deviation of the annual mean of CONT. Here, and in the next 2 panels, only years 501-700 are shown, because CONT ends at year 700; OP115 slowly approaches the values cited in the text. b: Like (a), but for the meridional heat transport (in PW). c: Maximum strength of the subpolar gyre (Sv, typically the maximum is located between Greenland and Labrador). Again, the black line is based on CONT, the red on OP115. Note that all the timeseries have been smoothed with an 11-year running mean. d: Mean temperature (color) and salinity (contour interval: 0.02 psu, minimum: -0.4 psu, maximum: 0.1 psu) difference between Labrador and Norway.

Figure 4: a: Mean maximum annual boundary layer depth (m, in color) and annual mean sea ice concentration (contour lines: 10%) for CONT. b: Like (a), but for years 551-600 of OP115. c: Mean temperature profile south of Iceland. d: Depth integrated transport in CONT (color), and difference between OP115 and CONT (OP115-CONT, contour interval: 1 Sv).
Figure 5: Sea Surface Salinity north of 50°N. Also shown are the locations of several seas and passages that are mentioned in the text.

Figure 6: Change in the total freshwater (FW) import from the Arctic Ocean to the subpolar Atlantic (thick black line) in OP115 compared to the mean of years 501-700 from CONT. This anomaly is also shown split up into the contribution of the FW import east (red line) and west (blue line) of Greenland, as well as split up into the solid FW import (mainly east of Greenland; solid thin black line) and the liquid FW import (dashed line). The fluxes are computed relative to a salinity of 34.8 (Aagaard and Carmack 1989).

Figure 7: a: Difference of zonally averaged flux at the top of the atmosphere (OP115-CONT). Black: insolation. Blue: insolation times albedo of CONT. Red: clear-sky net shortwave radiation. Green: net shortwave radiation. b: Spatial pattern of the difference in net shortwave radiation [in W/m²]. c: Difference in maximum (summer) low level cloud concentration (color, in %) and minimum (summer) sea-ice concentration (contour interval 5 %). For the sake of clarity only the cloud cover over the ocean is shown; with the exception of the Canadian Archipelago, the cloud cover changes over land are less than 5%. d: Difference in snow melt (color, in mm/day) and snow fall (contour lines: 0.1 mm/day).
Figure 1: Seasonal cycle of insolation at the top of the atmosphere for CONT (black) and OP115 (red) at 70°N (maximum in June) and 70°S. In OP115 the June insolation is 38 W/m² weaker than in CONT.
Figure 2: Difference in annual mean surface temperature between OP115 and CONT (a), and between their respective slab-ocean counterparts (b). c: Difference in annual mean snow depth between OP115 and CONT, d: area where the summer (minimum) snow depth is larger than 1 cm and the summer sea-ice concentration is larger than 40% for CONT (red), and its increase in OP115 (blue),
Figure 3: 

a: Strength of the AMOC (in Sv) at 50°N for CONT (black) and OP115 (red). The straight black lines denote the mean and mean ± one standard deviation of the annual mean of CONT. Here, and in the next 2 panels, only years 501-700 are shown, because CONT ends at year 700; OP115 slowly approaches the values cited in the text.

b: Like (a), but for the meridional heat transport (in PW).

c: Maximum strength of the subpolar gyre (Sv, typically the maximum is located between Greenland and Labrador). Again, the black line is based on CONT, the red on OP115. Note that all the timeseries have been smoothed with an 11-year running mean.

d: Mean temperature (color) and salinity (contour interval: 0.02 psu, minimum: -0.4 psu, maximum: 0.1 psu) difference between Labrador and Norway.
Figure 4: **a:** Mean maximum annual boundary layer depth ($m$, in color) and annual mean sea ice concentration (contour lines: 10%) for CONT. **b:** Like (a), but for years 551-600 of OP115. **c:** Mean temperature profile south of Iceland. Black: CONT; red: OP115 (yrs 551-600); green: OP115 (yrs 751-800). **d:** Barotropic streamfunction of control (color) and difference between OP115 and control (contour line: 1 Sv).
Figure 5: Sea Surface Salinity north of 50°N. Also shown are the locations of several seas and passages that are mentioned in the text.
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