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Sea-ice switches and abrupt climate change

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We propose that past abrupt climate changes were probably a result of rapid and extensive variations in sea-ice cover. We explain why this seems a perhaps more likely explanation than a purely thermohaline circulation mechanism. We emphasize that because of the significant influence of sea ice on the climate system, it seems that high priority should be given to developing ways for reconstructing high-resolution (in space and time) sea-ice extent for past climate-change events. If proxy data can confirm that sea ice was indeed the major player in past abrupt climate-change events, it seems less likely that such dramatic abrupt changes will occur due to global warming, when extensive sea-ice cover will not be present.

Keywords: climate; sea ice; thermohaline circulation; abrupt climate change; glacial cycles

1. Introduction

Rapid and dramatic climate changes that occurred within a few decades abound in climate proxies (Taylor *et al.* 1993; Alley *et al.* 1993; Lehman & Keigwin 1992). A number of mechanisms for these abrupt climate changes have been suggested in the past (Adams *et al.* 1999). In particular, a threshold-like behaviour of the thermohaline circulation (THC) and THC instability and variability have been suggested as possible mechanisms for rapid climate instability (Weaver *et al.* 1991; Tziperman *et al.* 1994; Tziperman 1997; Marotzke 2000; Ganopolski & Rahmstorf 2001). It was demonstrated that variations in the THC can result from small, possibly stochastic, perturbations in freshwater flux to the North Atlantic (Griffies & Tziperman 1995) due to the excitation of a damped oscillatory THC mode (see also Ganopolski & Rahmstorf 2001, 2002; Timmermann *et al.* 2002). However, we wish to make the case here that the THC by itself is not likely to be a satisfactory mechanism for observed past rapid climate changes. Rather, it is the dramatic rapid melts and expansions of sea-ice cover—through its strong cooling ice–albedo feedback, its insulating effect on local evaporation, and its role in diverting the storm track—that are a more likely explanation (Gildor & Tziperman 2000, 2001*b*; Adams *et al.* 1999; Broecker 2000; Dansgaard *et al.* 1989).

One contribution of 14 to a Discussion Meeting 'Abrupt climate change: evidence, mechanisms and implications'.

The objective of this paper is to make the following points.

(i) Sea ice is a likely source of large-amplitude abrupt climate changes.

(ii) Sea ice can amplify THC instabilities, which, without sea ice, do not have the ability to explain large-amplitude changes.

(iii) Abrupt climate changes can result from sea-ice variations even without accompanying THC changes.

(iv) Sea ice may also strongly affect temperature–precipitation feedback (explained below), which stabilizes the THC and plays a critical role in glacial dynamics.

We also argue here that abrupt, switch-like, sea-ice melts and expansions have played a role not only in millennial and multi-millennial variability, but also in the dynamics of the 100 000 yr (100 kyr) glacial cycles. In this regard, we briefly review the sea-ice switch mechanism for the 100 kyr cycles proposed by Gildor & Tziperman (2000, 2001*b*). These works have used a simple model that was later also applied to shorter (1–10 kyr) time-scale variability, again demonstrating the dominant role of sea ice (Timmermann *et al*. 2003).

2. Could the THC be responsible for abrupt climate change?

Because of its role in transporting heat to the high latitudes, the THC is often conjectured to play a major role in abrupt climate transitions or in transferring climatic signals between different regions and different components of the climate system (Marotzke 2000; Clark *et al*. 2002; Broecker & Denton 1989; Ganopolski & Rahmstorf 2001). Indeed, the behaviour of the THC in models ranging in complexity from simple box models to general circulation models (GCMs) shows a threshold behaviour, multiple equilibria and hysteresis (Marotzke 1989; Marotzke & Willebrand 1991; Tziperman *et al*. 1994; Tziperman 1997), demonstrating the ability of the THC to jump between different states in a few decades. Moreover, proxy records suggest that variations in the THC correspond to many observed abrupt climate-change events (Boyle 2000). However, these records do not necessarily mean that the THC is the driver of these abrupt climate changes, in particular since significant THC variations seem to occur only during Heinrich events (Elliot *et al*. 2002).

In contrast, recent evidence suggests that variations in the THC cannot, in fact, be the sole mechanism behind abrupt climate changes. First, while abrupt climate changes are seen far away from the North Atlantic (see Stocker (2000) and references cited therein), GCMs (Manabe & Stouffer 1988; Tziperman 1997; Rind *et al*. 1989) and simpler models (DeBlonde & Peltier 1993) show that a THC instability does create a teleconnection to the Southern Ocean, yet does not have a truly global effect. In order for a THC change to cause a significant atmospheric temperature signal even around the North Atlantic, it needs to change by a very significant amplitude that seems somewhat unreasonable and has not been observed for the inter-Heinrich periods (Elliot *et al*. 2002). Second, the overall role of ocean heat transport in the climate of the North Atlantic was questioned recently, with model results suggesting only a minor oceanic role unless accompanied by large variations in sea-ice cover (Seager *et al*. 2002; Kerr 2002). Third, the THC seems to be more stable than previously thought. Recent interpretation of proxy records and recent modelling studies both suggest that during past colder periods (e.g. the Last Glacial Maximum (LGM)) there was a similar rate of water formation in the North Atlantic to that of the present day, possibly at different sinking sites and to a reduced depth (Legrand & Wunsch 1995; Bigg *et al*. 2000; Yu *et al*. 1996; Kitoh *et al*. 2001; Tziperman & Gildor 2002). Part of this observed THC stability may be related to temperature–precipitation feedback (Le Treut & Ghil 1983) and the role of sea ice in it, as explored by Tziperman & Gildor (2002), as follows. The THC tends to become unstable for larger meridional freshwater transport by the atmosphere. Extensive sea-ice cover can reduce the meridional freshwater atmospheric flux through its albedo effect, which cools and therefore dries up the atmosphere, and by shifting the storm track away from the polar ocean, thus stabilizing the THC. Finally, in some models, such as that of Gildor & Tziperman (2001*b*), THC changes and instabilities are not the trigger of rapid climate changes and are not necessary for the existence of rapid climate changes but, rather, are a result of the rapid sea-ice variations.

Regardless of these arguments, one must admit that the proposed proximity of the present-day THC to a stability threshold (Walin 1985; Tziperman *et al*. 1994; Tziperman 1997), as well as the speculated existence of multiple equilibria of the THC (Manabe & Stouffer 1988), relies on two elements that are quite uncertain in today's ocean models (Zhang *et al*. 1999): the amplitude of the freshwater flux, and the formulation of the vertical mixing parametrization, to which the THC is quite sensitive (Bryan 1987). This does not mean that the THC cannot jump between different equilibria states or is not indeed close to such an instability threshold; it does indicate that the uncertainty regarding such proximity to instability threshold is not negligible.

If THC changes cannot solely account for the observed large-amplitude abrupt climate change, then what can? We tend to prefer the hypothesis that sea ice may have been the major player in these observed abrupt change events.

3. Sea ice: the major player?

Sea ice plays a crucial role in the climate system. Since the albedo of sea ice is much higher than that of the open ocean, extensive sea-ice cover has a strong cooling effect. In addition, sea-ice cover effectively insulates the ocean from the atmosphere, thus reducing the heat flux from the ocean to the atmosphere and preventing local evaporation (Jayne & Marotzke 1999; Lohmann & Gerdes 1998). At present, a significant part of the precipitation (20–45%) on the land area which was covered by ice sheets during the LGM comes locally from the Norwegian and Greenland seas and from the North Atlantic (Charles *et al*. 1994). Moreover, the existence of sea-ice cover in the region of deep-water formation in the North Atlantic implies that the temperature of the deep water formed there was close to the freezing temperature, in contrast to being a few degrees higher in the modern ocean (Broecker 2000). This temperature change can be transported by the THC and can affect the temperature of the water upwelling in the Southern Ocean (Gildor & Tziperman 2001*a*) and elsewhere. Finally, by increasing the meridional temperature gradient in the atmosphere, extensive cover of sea ice may be expected to change the atmospheric circulation, and in particular to shift the storm track southward (Hall *et al*. 1996; Kapsner *et al*. 1995).

These climatic effects of sea ice can explain many of the observed characteristics of abrupt climate changes, including the amplitude of temperature changes and variations in dustiness (Mayewski *et al*. 1994). The records of these abrupt changes far away from the North Atlantic might be explained by the effect of sea ice on atmospheric circulation (Mayewski *et al*. 1994) or by the effect of sea ice on the temperature of the water formed in the North Atlantic in the presence of sea ice (Gildor & Tziperman 2001*a*). In particular, these sea-ice effects may account for the so-called 'temperature–precipitation feedback' (Le Treut & Ghil 1983), i.e. the observation that the rate of snow accumulation over land glaciers decreases for colder climates (which were presumably characterized by a larger sea-ice extent (Alley *et al*. 1993)).

Sea ice is a strong candidate for explaining abrupt climate changes because there is an inherent threshold in sea-ice behaviour due to the existence of the freezing point, together with the positive ice–albedo feedback. Thus, once sea ice forms locally somewhere, the sea-ice–albedo feedback tends to enlarge its cover and accelerate its formation, until counteracted by another feedback. Such a counter feedback can be, for example, the insulating effect of sea ice, as in the model of Gildor & Tziperman (2001*b*): the sea ice grows when the ocean temperature is cooled by the air–sea heat flux to below the freezing temperature. When the sea-ice cover increases, it insulates a larger part of the ocean from the colder atmosphere and reduces the oceanic heat loss. The oceanic advection and diffusion of heat from the mid-latitudes then balances the reduced cooling by the atmosphere, resulting in no more net cooling of the ocean, and in stopping sea-ice expansion.

An important point we want to make here is that once the ocean is close to the freezing temperature, the initiation of sea-ice growth may be the result of reduced heat transport by a weakened THC. However, several other mechanisms may lead to sea-ice formation even without variations in the THC. For example, lower insolation (Rind 2002), reduced atmospheric CO_2, volcanic eruption, which may cause temporal cooling, or variations in land-ice sheets through their albedo effect (Adams *et al*. 1999) can all initiate sea-ice growth. We therefore do not necessarily expect a strong correlation between THC variations and abrupt climate changes.

While sea ice is a clear candidate for explaining the large temperature variations observed during Dansgaard–Oeschger oscillations and Heinrich events, it may have also played an important role in the dynamics at longer time-scales. Sea ice, in fact, plays a crucial role in recent theory for the 100 kyr glacial cycles (Gildor & Tziperman 2000), for the glacial–interglacial CO_2 variations (Gildor & Tziperman 2001*a*), and for the Mid-Pleistocene climate transition that led to the initiation of the 100 kyr glacial cycles (Tziperman & Gildor 2003). This 'sea-ice switch' glacial cycle mechanism may be briefly described as follows. Consider an interglacial period as the beginning of a glacial cycle. As the land ice begins to grow from its minimum point, the ocean is ice free and the atmospheric and oceanic temperatures are rather mild. Snow accumulation over glaciers exceeds the ablation, melting and calving term, and therefore the ice sheets gradually grow. The resulting slow increase in land albedo slowly reduces the temperature of the atmosphere and of the ocean. After some 90 000 yr, when the atmosphere is sufficiently cooled and the high-latitude sea surface temperature reaches the freezing temperature, sea ice forms and expands very rapidly. The expansion of sea ice further increases the albedo, induces a further reduction in atmospheric temperature, and results in the creation of more sea ice (a positive feedback). In a few decades, a large sea-ice cover is created in the high latitudes. Sea ice stops growing when it insulates enough of the polar oceans from the cold atmosphere, reducing the air–sea cooling that leads to the sea-ice formation. The sea-ice 'switch' is now turned to 'on'.

At this stage, the average global temperature is lowest, sea-ice and land-ice sheet extents are maximal, and the system is at a glacial maximum. The low atmospheric temperature reduces the poleward atmospheric moisture flux to about half its maximum value. Similarly, the sea-ice cover limits the moisture extraction from the polar ocean and the corresponding snow accumulation over the land ice. As ablation, glacier melting, calving and run-off, being less sensitive to temperature, proceed as before, the glaciers start retreating. The albedo decreases again, and the atmospheric temperature rises slowly. This is the beginning of the termination stage of the glacial period. After some 5000–10 000 yr, the ocean warms sufficiently to allow the sea ice to start melting, again within a few decades, due to the sea ice–albedo feedback working this time to warm the atmosphere and ocean. The sea-ice switch is now turned to 'off', the temperature of both the atmosphere and the ocean increases, and the system has completed a full glacial cycle.

Previously (Gildor & Tziperman 2001*b*) we speculated that the switch-like behaviour of sea ice may account not only for the glacial–interglacial transitions on a 100 kyr time-scale, but perhaps also for the many rapid climate transitions seen in the proxy observations in the context of shorter-term variability. Other model results (Thorndike 1992) also support the idea that reasonable changes in the heat budget of the northern ocean can cause dramatic changes in the sea-ice cover.

Indeed, switch-like behaviour of sea ice on a millennial time-scale was found by Timmermann *et al.* (2003) in a coupled model based on the model of Gildor & Tziperman (2001*b*), while studying the role of mean and stochastic freshwater forcing on the generation of millennial-scale variability in the North Atlantic. Similar results were found in a somewhat different model of Heinrich events by Y. Kaspi & E. Tziperman (2003, unpublished research). In addition to amplifying the atmospheric response to THC variability, sea ice played a crucial role in Timmermann *et al.* (2003) by providing a negative feedback to the abrupt strengthening of the THC. In a set of sensitivity experiments in which the sea-ice fraction and depth were forced to remain constant, millennial-scale oscillations were not observed using the same stochastic and meltwater forcing which generated such oscillation in the presence of an active sea-ice switch.

4. Discussion

Sea ice seems to play a major role in abrupt climate changes. We speculate here that variations in the strength of the THC by themselves may not be able to explain the observed characteristics of past abrupt climate changes as recorded in proxy records. However, THC variability may, possibly in combination with stochastic forcing, serve as a trigger for sea-ice switches which result in an amplified climatic signal (Timmermann *et al.* 2003).

There are still a few obstacles that need to be overcome for sea-ice variations to be able to satisfactorily explain the observed past abrupt climate changes. First, if a given portion of the ocean (e.g. the North Atlantic) is cloud covered, its albedo is already high, and having it covered by sea ice may not cause a sufficient increase in albedo and therefore may not cause sufficient atmospheric cooling. To examine this issue, it is necessary to examine the typical cloud cover and cloud albedo during an interglacial (i.e. present-day climate) over areas that were covered in sea ice during the LGM. Additionally, there have been suggestions that rapid climate change may have been caused by rapid changes in atmospheric circulation, although these suggestions seem to be at a preliminary stage, still lacking a specific physical mechanism. Should such a specific mechanism be suggested, independent of rapid changes in atmospheric circulation induced by sea-ice variations, this could be a serious contender to sea ice as the cause of abrupt climate changes observed in the climate record. Unfortunately, sea-ice proxies have proven to be very difficult to obtain, particularly combined information on the timing and spatial extent of sea-ice variations (de Vernal & Hillaire-Marcel 2000; Sarnthein *et al*. 2003). Perhaps future work on sea-ice proxies will be able to resolve its role in past abrupt climate changes.

Whether sea ice is responsible for past abrupt climate changes or whether they were caused by the THC or by atmospheric dynamics matters quite a bit in the context of global warming. If sea ice is to be implicated for the abrupt past climate transitions, then we cannot expect future global warming to result in as strong or as rapid climate transitions as were observed during the LGM.

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Discussion

P. W. Nienow (*Department of Geography and Topographic Science, University of Glasgow, UK*). Is your sea-ice switch theory supported by the marine geological record? Presumably, if your theory is correct, the foram record off Western Europe should reveal 'cold' foram species during relatively 'high' sea level (as indicated by the $^{18}O/^{16}O$ record) when sea ice is at its maximum extent.

E. Tziperman. One of the advantages of the sea-ice switch mechanism for the glacial cycles is indeed that it makes new falsifiable predictions. What is needed for this purpose are proxies from which information on the relative timing of sea ice and land ice can be extracted, with a good spatial and temporal resolution, such that we can see the relative phases of the two during glaciation and deglaciation. Such records may become available in the near future but are presently not available in sufficient quality or resolution.

J. G. Shepherd (*School of Ocean & Earth Science, University of Southampton, UK*). There is one very curious feature of your model, in that moisture transport depends only on the downstream moisture and temperature. To what extent is this an important factor, and what happens if you use a more normal parametrization?

E. Tziperman. The details of this parametrization (which is based on the meridional temperature gradient and on the downstream humidity) are clearly not easily justifiable on physical grounds. However, this parametrization produces the temperature–precipitation feedback, as it results in less accumulation during colder periods. Any other, more justifiable, parametrization that can produce the temperature–precipitation feedback will result in the same sea-ice switch mechanism as well. For further discussion of different parametrizations, please see Gildor & Tziperman (2001*b*) and Tziperman & Gildor (2002).

R. B. Alley (*Department of Geosciences and EMS Environment Institute, The Pennsylvania State University, USA*). During the mid-Holocene, high-latitude northern temperatures were a little higher than more recently, and available data indicate that Arctic sea ice and glaciers were reduced compared with recently; cooling and sea-ice growth since then occurred with glacier growth. Similarly, warming from the Little Ice Age seems to have reduced sea ice and glaciers. Yet, in your model, warming and sea-ice reduction grow glaciers, contrary to these data. How do you reconcile these?

E. Tziperman. We expect the temperature–precipitation feedback to hold within a certain regime of the climate system. Clearly, a too-warm climate, one in which melting is dramatically enhanced or in which most precipitation falls as rain, will lead to a melting of the ice sheets and the temperature–precipitation feedback will not work any more (Tziperman & Gildor 2003). The favourite conditions for this feedback are warm winters and cold summers (Miller & de Vernal 1992). This feedback may be dominated locally by other effects, but is still likely to be a good representation of the larger-scale glacial climate dynamics. Indeed, there is much evidence of glaciers that have actually grown during the Holocene, such as the Columbia Ice Field in the Canadian Rockies (Luckman 1988) and others. Please see also discussion in Gildor & Tziperman (2001*b*) and Gildor (2003).

J. H. Lawton (*Natural Environment Research Council, UK*). The crucial empirical test of your model is to compare patterns of sea-ice and glacier growth on appropriate time and spatial scales. It would not surprise me at all that at short time-scales there are 'wobbles' around the main trends that do not agree with your predictions. So getting the big-picture empirical test right will be tricky. Do you agree, and do we have the data?

E. Tziperman. We completely agree that short time-scales superimposed on the mean make the observational test more complicated. Moreover, the available data do not enable us to test phase relation with confidence (see the response to the first question above). However, future proxy records might be able to test the prediction made by this theory on the phase relation between sea- and land-ice covers.

A. Mahadevan (*DAMTP, University of Cambridge, UK*). In your model (which proposes that sea ice is the trigger for the climate changes), the 100 kyr time-scale arises from the particular choice of parameters: land-ice accumulation–ablation rate, or the temperature at which the ocean starts to freeze. So rather than say that the model 'explains' the 100 kyr cycle, I would say that with a logical choice of parameters, the model reproduces the 100 kyr cycle. Would you agree?

E. Tziperman. A logical choice of parameters in this mechanism can produce a time-scale ranging from a minimum of 50 kyr (which occurs for a symmetric oscillation) to 140 kyr, which produces a very asymmetric oscillation with long glaciation and short deglaciation (Gildor & Tziperman 2001*b*). For an asymmetric oscillation where the glacier grows during 85% of the period and retreats during the remaining 15%, the time-scale is around 100 kyr. Please note that these time-scale estimates come from a very simple argument, rather than from the detailed model with its many specified parameters (Gildor & Tziperman 2001*b*). The inclusion of Milankovitch forcing leads to a phase locking of the internal 100 kyr climate oscillations for a range of the parameters. As a result, the 100 kyr time-scale is more robust in the presence of Milankovitch forcing. In addition, we have proposed that the asymmetry and time-scale arise from the same mechanism that is responsible for the mid-Pleistocene transition from 41 kyr to 100 kyr glacial oscillations (Tziperman & Gildor 2003).

T. M. Lenton (*Centre for Ecology and Hydrology, Edinburgh, UK*).

(i) You show in Gildor *et al.* (2002) that the period of the glacial–interglacial cycle is sensitive to the inclusion of the carbon cycle. Is it necessary to include the carbon cycle in order to get a reasonable reconstruction of the timing of the ice-volume changes with Milankovitch/orbital forcing?

(ii) What happens when you run your model with future Milankovitch/orbital forcing? (Do you predict that the present interglacial will be anomalously long?)

E. Tziperman.

(i) Given that the model is highly simplified, all we can expect from it is a rough order of magnitude estimate for the time-scale. As explained above, this is indeed obtained, via an intuitive argument for the time-scale based on the sea-ice switch mechanism (Gildor & Tziperman 2000, 2001*b*), which results in an estimate anywhere between 60 and 140 kyr. The model is also capable of predicting that including an active geochemistry may change the time-scale somewhat, but not by how much. Now, the Milankovitch forcing results in phase locking of the glacial cycles, which tends to set the glacial period to be a multiple of the precession period (i.e. about 80 kyr, or 100 kyr, or 120 kyr, etc.), and thus makes it less sensitive to other influences. This means that in the presence of Milankovitch forcing, the sensitivity to active geochemistry may be reduced.

(ii) We have not tried running the model with future Milankovitch forcing, although this is an interesting idea, so perhaps we should.

Additional references

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