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The global response to Younger Dryas boundary conditions in an AGCM simulation

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Abstract. Geological evidence points to a global Younger Dryas (YD) climatic oscillation during the last glacial/present interglacial transition phase. A convincing mechanism to explain this global YD climatic oscillation is not yet available. Nevertheless, a profound understanding of the mechanism behind the YD climate would lead to a better understanding of climate variability.

Therefore, the Hamburg atmospheric circulation model was used to perform four numerical experiments on the YD climate. The objective of this study is to improve the understanding of different forcings influencing climate during the last glacial/interglacial transition and to investigate to what extent the model response agrees with global geological evidence of YD climate change. The following boundary conditions were altered: sea surface conditions, ice sheets, insolation and atmospheric CO_2 concentration. Sea surface temperatures based on foraminiferal assemblages proved to produce insufficient winter cooling in the N Atlantic Ocean in two experiments. It is proposed that this discrepancy is caused by uncertainties in the reconstruction method of sea surface temperatures. Therefore, a model-derived set of Atlantic surface ocean conditions was prescribed in a subsequent simulation. However, the latter set represented an Atlantic Ocean without a thermohaline circulation, which is not in agreement with evidence from ocean cores. The global response to the boundary conditions was analysed using three variables, namely surface temperature, zonal wind speed and precipitation. The statistical significance of the changes was tested with a two-tailed t-test. Moreover, the significant responses to cooled oceans were compared with geological evidence of a YD oscillation. This comparison revealed a good match in Europe, Greenland, Atlantic Canada and the N Pacific region, explaining the YD oscillation in these regions as a response to cooled N Atlantic and N Pacific Oceans.

However, the results leave the YD climate in other regions completely unexplained. This reflects either an insufficient set of boundary conditions or the important role played by feedbacks within the coupled atmosphere-ocean-ice system. These feedbacks are poorly represented in the used atmospheric model, since ice sheets and the ocean surface conditions have to be prescribed.

1 Introduction

The transition from the last glacial to the present interglacial climate was irregular. This phase is known in N Europe as the Weichselian Late Glacial and lasted from about 14.5 to about 11.5 cal ky BP (thousand calendar years before present). A relatively warm stage existed (Late Glacial Interstadial, LGI, ~14.5–12.5 cal ky BP), interrupted by several short cool episodes (Lowe et al. 1994). From ~12.5 to 11.5 cal ky BP conditions were close to glacial levels during the Younger Dryas stadial (YD). After ~11.5 cal ky BP temperatures increased again, indicating the start of the Holocene. In this work the results of numerical simulations on YD climate were analysed to improve the knowledge of the Younger Dryas cooling.

Traditionally, the YD cooling is primarily viewed as a N Atlantic event (Berger 1990), because evidence for the YD is most clearly found in geological records surrounding the N Atlantic Ocean (e.g. Lowe et al. 1994). Moreover, the most influential hypothesis to explain the YD cooling is based on a shut-down of the Atlantic thermohaline circulation under the influence of melt-water influxes (Broecker 1992). This explanation is however contradicted by evidence provided by ocean cores, suggesting that the thermohaline circulation was functioning during YD time (Veum et al. 1992; Sarnthein et al. 1994). Furthermore, various reports point to a climatic oscillation during YD time in areas away from the N Atlantic region (Peteet 1995). These sites include ones in the mid-latitudes of the Northern Hemisphere (NH) such as central USA (Shane and Anderson 1993), the N Pacific area (e.g. Mathewes 1993) and China (Porter and Zhisheng 1995). In addition, reports from the tropics and the high latitudes of the Southern Hemisphere are increasing in number: Africa (e.g. Roberts et al. 1993), S. America (e.g. Thompson et al.

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1995), New Zealand (Denton and Hendy 1994) and Antarctica (Jouzel et al. 1992).

Recently, the discovery of ice-rafted sediments of YD age in N Atlantic Ocean cores gave new insight into the mechanism behind the YD cooling. This finding indicated that the YD was related in origin to preceding cool episodes with intensive ice-rafting in the N Atlantic Ocean, known as Heinrich events (Bond et al. 1993). New reports suggest that these Heinrich events also left their marks in the S Hemisphere (Lowell et al. 1995). A number of suggestions were put forward as mechanisms to explain these supposed global changes, including ocean circulation changes (Broecker 1994) and variations in the atmospheric water-vapour content (Lowell et al. 1995). A convincing mechanism to explain these climate changes is, however, not available. Nevertheless, a profound understanding of the mechanism behind the YD climate is crucial for our understanding of climate sensitivity.

Possible mechanisms responsible for the YD climate can be explored with climate model simulations. Rind et al. (1986, hereafter R86) performed several sensitivity experiments with an atmospheric general circulation model (AGCM) to study the impact of a cold N Atlantic Ocean on climate. They prescribed boundary conditions which were crude estimates of the YD situation. For instance, sea surface temperature estimates of the last glacial maximum (LGM, ~21 cal ky BP) were defined north of 25 °N in the Atlantic Ocean. In addition, the present sea-ice cover was prescribed. Other boundary conditions included 11 ky BP land ice and orbital parameters. R86 found no statistically significant cooling outside the N Atlantic region and vicinity. Consequently, R86 concluded that possible YD effects in Africa, western N America and the S Hemisphere were completely unexplained by their model results.

In this study the Hamburg AGCM was used to carry out four experiments on the YD climate. These experiments were designed to simulate a realistic YD climate in Europe, which can be compared with detailed climate reconstructions based on geological data (e.g. Isarin et al. 1997). The difference in design compared to R86 is summarized as follows. First, a 'second generation' model was used with a higher horizontal resolution (~2.8° × 2.8° versus 8° × 10° latitude by longitude). Second, new studies on the YD situation provided an updated set of boundary conditions, namely cooled N Atlantic and N Pacific oceans (including sea-ice), ice sheets in N America and Scandinavia, changed insolation and lowered atmospheric CO_2 content. It is interesting to investigate if these differences produce significantly different results compared to R86. Therefore it was attempted to identify the impact of the individual boundary conditions.

In this study the response in surface temperature, zonal wind speed and precipitation is analysed and the statistical significance of the deviations is tested. Moreover, the effect of cooled oceans is compared with YD climate reconstructions based on geological evidence. The aim is to improve the understanding of forcings influencing climate during the last glacial/interglacial transition and to investigate to what extent the model response to cooled oceans explains geological evidence of *global* YD climate change. Despite the mentioned differences with R86, the same conclusions are reached, since no significant YD signal is found outside the N Atlantic and N Pacific regions. This could have two causes. First, the set of boundary conditions used could be incomplete. Second, an atmospheric model could be inappropriate to simulate a global YD climate, because unknown feedback processes between atmosphere, oceans and land ice played an important role.

2 Experimental design

The ECHAM3 (European Centre/HAMburg) AGCM was used to perform several simulation experiments on the YD climate. A detailed description of this model can be found in DKRZ report number 6 (DKRZ, Modellbetreuungsgruppe 1993). The ECHAM3 model is capable of simulating most aspects of the observed present time-mean atmospheric circulation and its inter-seasonal variabilities (Roeckner et al. 1992). The T42 version was used, corresponding to a horizontal resolution of approximately 2.8° latitude–longitude. The model has 19 atmospheric layers. A diurnal and annual cycle was simulated. The YD experiments were run for 12 years, of which only the last 10 years were used to account for 'spin-up' time.

A summary of the experiments is given in Table 1. As a control experiment (hereafter called CTRL) a simulation of present climate was used with climatological sea surface temperatures (SSTs) prescribed (see Roeckner et al. 1992). CATL is a sensitivity experiment with a N Atlantic Ocean cooled according to geological evidence. The used SST reconstructions were based on foraminiferal assemblages as presented by Sarnthein et al. (1995). A sine function was fitted through the summer and winter SST estimates to obtain an annual cycle. The N Atlantic cooling at 65 °N amounted 6 to 10 °C in summer and 4 to 6 °C in winter (see Fig. 1a, b). It was assumed that the tropical Atlantic was not substantially cooler than today. This assumption

Table 1. Summary of experiments, with the most important boundary conditions mentioned

Boundary conditions	Experiments			
	CTRL	CATL	YD1	YD2
SSTs and sea ice	Present	YD (Atlantic)	YD (Atlantic)	Summer: Atlantic as YD1 Winter: model output − 2 °C in N Pacific
Land ice	Present	Present	YD	YD
Insolation	Present	Present	YD	YD
CO_2 (ppmv)	345	345	230	230

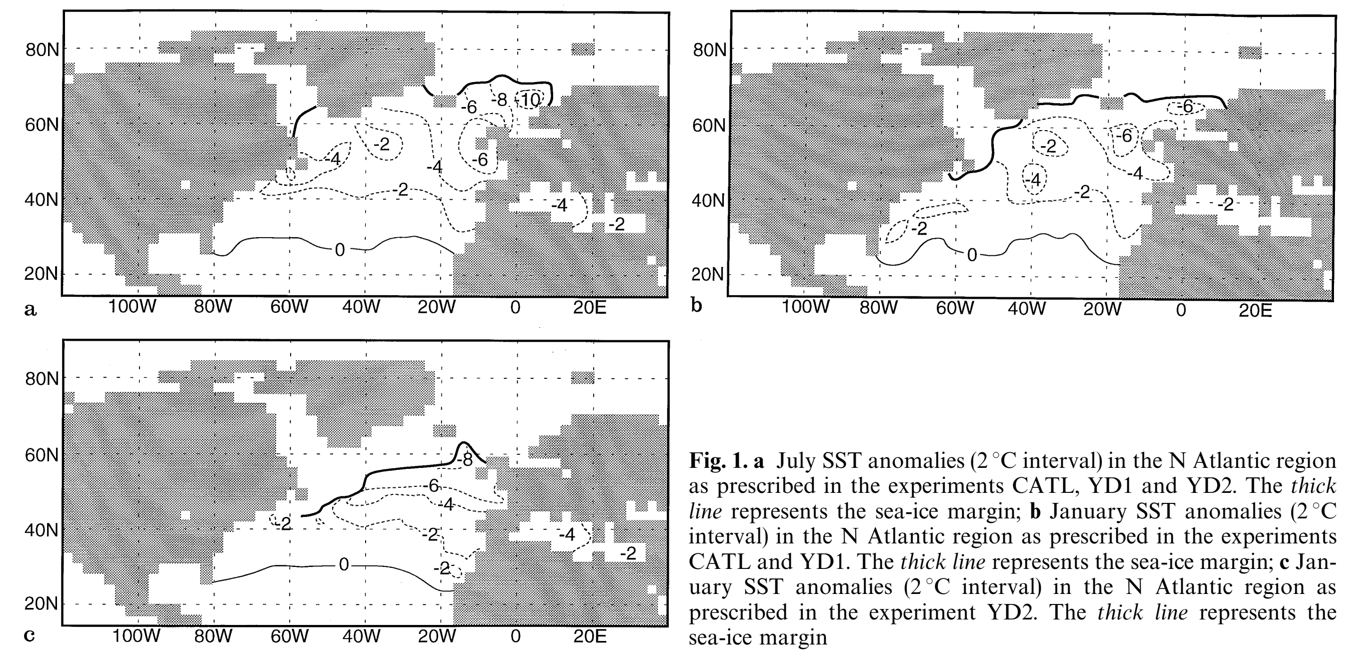


Fig. 1. a July SST anomalies (2 °C interval) in the N Atlantic region as prescribed in the experiments CATL, YD1 and YD2. The *thick line* represents the sea-ice margin; **b** January SST anomalies (2 °C interval) in the N Atlantic region as prescribed in the experiments CATL and YD1. The *thick line* represents the sea-ice margin; **c** January SST anomalies (2 °C interval) in the N Atlantic region as prescribed in the experiment YD2. The *thick line* represents the sea-ice margin

agrees with the SST reconstructions of Schulz (1995). The sea-ice limit in the Greenland–Iceland–Norway seas was defined at 70 °N during summer and at 65–67 °N during winter in agreement with Koç et al. (1993).

In a third experiment (hereafter called YD1) other boundary conditions were altered in addition to the changed N Atlantic prescribed in CATL. Ice sheets were introduced according to Peltier (1994), resulting in Laurentide and Scandinavian ice sheets with maximum elevations of about 1500 m and 1000 m respectively. A surface albedo of 0.8 was prescribed for these ice sheets, whereas a value ranging from 0.6 to 0.8 was used for the ice sheets in Greenland and Antarctica. Furthermore, the insolation was changed to 12 cal ky BP conditions in agreement with Berger (1978), with a reduction during NH winter and an increase during NH summer. In January the largest deviations occurred in the tropics ($\sim 25\ W/m^2$ change), whereas during June the main difference was situated in the middle and high latitudes of the N Hemisphere (~ 40–$50\ W/m^2$ change). Finally, the atmospheric CO_2 concentration was lowered as suggested by Jouzel et al. (1992) and land points were added in the North Sea and Bering Strait areas as suggested by sea level curves.

CATL and YD1 were evaluated by comparing the results with climate reconstructions based on geological evidence (Renssen et al. 1995). This comparison revealed that the simulations produced temperatures in Europe that were considerably higher than in the reconstructions. The difference was especially clear for the winter season, as in CATL and YD1 a cooling of only a few degrees was present in NW Europe, whereas a depression of at least 15 °C was estimated using various types of geological data. In Renssen et al. (1995) it was argued that this deviation between model and data was mainly caused by uncertainties in the set of boundary conditions. Moreover, since the anomalous air temperatures over Europe were similar in CATL (only SSTs changed) and YD1 (total set of YD boundary conditions), it was inferred that the Atlantic winter SST set used was anomalously warm in these simulations for YD conditions (Renssen et al. 1995). It should be noted that the uncertainty of the SST reconstruction is large, typically in the order of 1 °C and increasing in the lower temperature range (e.g. Schulz 1995). An important factor contributing to this uncertainty is that in these lower temperatures the assemblages of foraminifera, used in the SST reconstructions, are dominated by one species. Moreover, the sedimentation rate in the majority of cores used for the SST reconstructions ranged between 2 to 7 cm/ky (Sarnthein et al. 1995), which may have been insufficient to register the maximum YD cold pulse. This cold phase may have lasted only a few hundred years as inferred from Greenland ice cores (e.g. Grootes et al. 1993). Furthermore, sampling density may not have been adequate. With these uncertainties in mind, it could be argued that there is room for several SST sets for the real YD situation and that it is justified to redefine the winter SST set used in CATL and YD1 (Renssen et al. 1995).

Following the discussed evaluation of CATL and YD1, the Atlantic winter SST set was redefined to perform a fourth experiment (hereafter called YD2). For this purpose the output of a coupled ocean-atmosphere model was used (the Hamburg large scale geostrophic ocean model coupled to the atmospheric ECHAM3-T21 model), in which the thermohaline circulation in the N Atlantic Ocean was halted by introducing a large amount of fresh water (Schiller et al. 1996). It is essential to note that the model-output was used to obtain a physically consistent set of SSTs and not because it was believed the YD cooling was caused by a shut-down of the thermohaline circulation. The resulting N Atlantic SSTs are 2 to 4 °C lower than in CATL, causing a southward shift of the mean winter sea-ice margin from 65 °N to 56 °N in YD2 (see Fig. 1c). In the Pacific Ocean north of 40 °N an arbitrary cooling of 2°C was defined in YD2, which is consistent with Kallel et al. (1988). This N Pacific cooling was deduced from the results of an additional AGCM experiment, which is discussed in Renssen et al. (1995).

In this study the simulation results are compared with those of R86 and therefore an overview of their experimental design is given here. They used the AGCM of the NASA/ Goddard Institute for Space Studies (GISS) with a low horizontal resolution of 8° of latitude by 10° of longitude. In addition, nine atmospheric layers are included. The following three experiments were performed together with a control simulation. In a first experiment LGM sea surface temperatures (CLIMAP project members 1981) were prescribed in the N Atlantic north of 25 °N, with a maximum cooling of 12 °C at 60 °N. Therefore, the N Atlantic SSTs were lower than the values prescribed in the present study (4 to 6 °C difference). However, the modern sea-ice cover was defined and the SSTs in the Pacific Ocean were left unchanged. In a second simulation 11 ky BP land ice (Denton and Hughes 1981) and insolation were prescribed together with the present SSTs. The maximum elevation of the Laurentide and European ice sheets was set to around 2000 m and 1500 m respectively, so about 500 m higher than the values used in the present study. In a third experiment the boundary conditions of the first and second simulations were combined. The present CO_2 concentration was prescribed in all experiments.

3 Results and discussion

3.1 Introduction to the results presented

Geological evidence of YD climate change is mostly described as an oscillation or climate reversal, so with reference to the climates of the preceding (LGI) and following (early Holocene) periods. Consequently, to compare the simulation results with geological evidence of the YD climate, one has to take the model response to the boundary conditions that were different during YD time compared to the LGI and early Holocene. Considering the anomalous boundary conditions used in the YD simulations, the most important difference between the YD event and the preceding and following periods was the cooled ocean surface (Wright 1989; Harrison et al. 1992). In other words, if AGCM experiments were to be designed to simulate the LGI or early Holocene climates, approximately the same ice sheets, insolation and CO_2 concentration would be defined as in a YD experiment, but the ocean surface conditions would be changed. SST reconstructions covering the period from the LGI to the present, show clearly a sharp YD cooling in the N Atlantic Ocean (Schulz 1995). Other boundary conditions show changes that would influence climate in a less prominent way. First, the global volume of land ice was decreasing during this deglaciation period (Peltier 1994) and therefore it is unlikely that the changed ice sheet configuration directly caused the YD cooling. Second, the insolation remained approximately at the same level during this period (Berger 1978). Third, since the atmospheric CO_2 concentration was slowly increasing during the time under consideration (Jouzel et al. 1992), it is improbable that this change produced a YD cooling. Therefore, to make a comparison with geological evidence of a YD *oscillation*, first a distinction must be made between the response to cooled oceans on the one hand and the combined response to changed ice sheets, insolation and CO_2 concentration on the other hand.

An evaluation of the experimental results suggested that YD2 provides a better approximation for the situation during the Younger Dryas than YD1 and CATL (Renssen et al. 1995). The YD2-CTRL difference is therefore taken as a representation of the overall response to YD boundary conditions. Moreover, the difference between the results of CATL and YD1 is taken as a measure of the combined response to prescribed ice sheets (including albedo change), insolation changes and CO_2 concentration (see Table 1). Furthermore, this composite effect (YD1-CATL) is subtracted from the YD2-CTRL difference to obtain an estimate of the response to the prescribed cooled oceans alone. In Sect. 3.5 this effect of cooled oceans is compared with geological evidence of the YD cooling.

An important assumption in this context is that the response to the total set of boundary conditions in the AGCM is linear and consists simply of the sum of individual effects. For example, it is assumed that the effect of the prescribed ice sheets is similar in the experiments YD1 and YD2, and that this ice sheet effect is not significantly influenced by the difference in SST sets. It is realized that this assumption represents a rough approximation, since the climate system may react non-linear in reality. Nevertheless, it is suggested that the assumption is valid for the purpose of this study. This is supported by other AGCM studies. A comparison of the experiments of R86 (see Sect. 2), for instance, showed that the responses to a cooled Atlantic, represented by the differences between the first and control simulation on the one hand and between the third and second on the other hand, were very similar. This was true for temperature, sea level pressure, wind fields and precipitation. Thus, in the results of R86 the combined effect of ice sheets and insolation change had no substantial effect on the response to lowered SSTs. Support is also provided by Rind (1987), who performed several sensitivity experiments on the climate of the last glacial maximum. These experiments were designed to study the individual effects of lowered SSTs, 10 m thick ice sheets and full land ice topography. The comparison of the results also suggests that the overall response consists of the sum of the separate effects.

In the Sect. 3.2 to 3.4 the global distributions of the following surface variables are presented: temperature, zonal wind speed and precipitation. These variables were chosen because they may be compared with geological evidence. A detailed analysis of the atmospheric winter circulation was published elsewhere (Renssen et al. 1996). The results are plotted as seasonal averages, namely for December–January–February (DJF) and June–July–August (JJA). The statistical significance of the changes in the YD experiments was determined with regard to the inter-annual variability in the CTRL experiment, using a two-tailed t-test as suggested by Chervin and Schneider

(1976). In these statistical analyses the seasonal means of the last 30 y of the control experiment and the last 10 y of the YD experiments were used. The discussion is focused on the statistically significant changes. The following question is answered: what can be said about the contribution of individual boundary conditions (i.e., cooled oceans and changes in ice sheets, insolation and atmospheric CO_2 concentration) to the overall response?

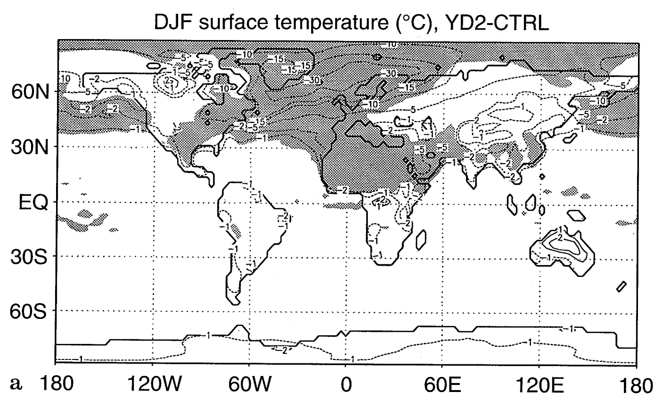
3.2 Surface temperature

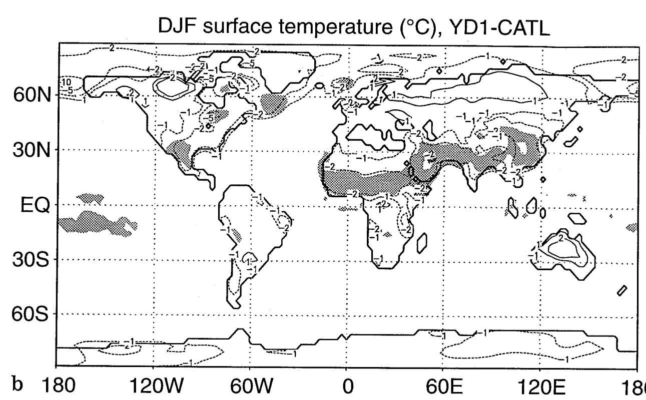
DJF

As expected, the largest cooling in YD2 is present in areas where the surface boundary conditions were altered (see Fig. 2a). Over the N Atlantic Ocean the temperatures are as much as 30 °C lower than in CTRL. Over the adjacent continents cooling is less severe, ranging from 15 °C in NW Europe to 2 °C in eastern USA. Over NH continents away from the N Atlantic Ocean a cooling up to 5 °C is visible. In the S Hemisphere there are only a few patches with a change in temperature. The noted differences with CTRL are significant at a 95% level in Europe, eastern N America, N Africa and some parts of Asia. As shown in Fig. 2b, the combined effect of changes in insolation, land ice and atmospheric CO_2 concentration caused a marked band of significant cooling of 2 °C extending from W Africa through Arabia and N India to China. At the same latitude a patch of significant cooling is present in southern N America. The notion that the DJF changes in insolation were especially large in the tropics (Berger 1978) suggests that this band of cooling was caused by the decreased insolation. Moreover, in other AGCM studies covering the same period a similar cooling in N Africa and E Asia was found. Kutzbach et al. (1993) simulated in their 12 ky BP experiment in these regions a 2 to 4 °C cooling, which was attributed to the prescribed insolation change. Similarly, R86 found a lowering of the temperature ranging from 1 to 4 °C over tropical lands due to changed orbital parameters.

The expected response to the changed CO_2 concentration is very small and it is therefore difficult to estimate this effect. This follows from Kutzbach and Guetter (1986), who performed two LGM simulations with different CO_2 concentrations. In one case the CO_2 content was set to the modern value of 330 ppmv and in a second case to the LGM value of 200 ppmv. Kutzbach and Guetter (1986) found only a small cooling of about 0.2 °C for the NH land surfaces as a result of the lowering of the CO_2 concentration. This small cooling was expected, because the dominant part of the CO_2 effect was already incorporated in the prescribed SSTs and sea-ice. In the experiments presented here, the CO_2 concentration was lowered to a value of 230 ppmv. Consequently, the effect of this lowering is probably even smaller than found by Kutzbach and Guetter (1986).

In Fig. 2b a statistical significant DJF temperature depression of 5 °C is seen over eastern N America at the location where the Laurentide ice sheet was prescribed. Sensitivity experiments with LGM ice sheets performed by Rind (1987) suggest that during winter mainly the topography of ice sheets causes cooling. He noted that the cooling was not simply caused by the increase in altitude, because a temperature inversion is normally present at locations where ice altitude is greatest. Instead, Rind





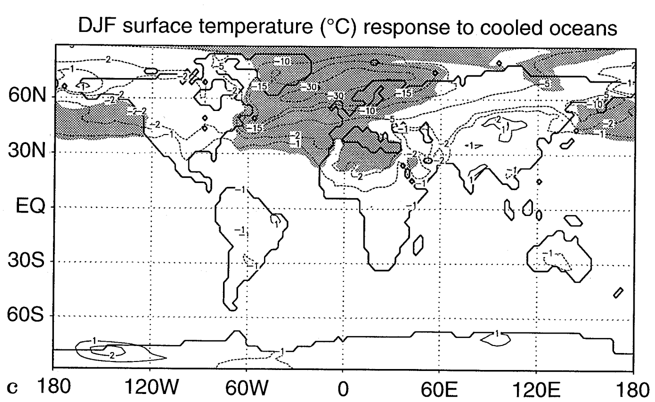


Fig. 2. a Difference in DJF surface temperature (°C) between the experiments YD2 and CTRL, representing the response to the total set of YD boundary conditions: cooled ocean surfaces, insolation, land ice and atmospheric CO_2 concentration. Contours at − 30, − 15, − 10, − 5, − 2, − 1, 1 and 2 °C. *Shading* shows statistically significant changes at 95% level; **b** difference in DJF surface temperature (°C) between the experiments YD1 and CATL, representing the combined response to 12 cal ky BP insolation, land ice and atmospheric CO_2 concentration. Contours at − 30, − 15, − 10, − 5, − 2, − 1, 1 and 2 °C. *Shading* shows statistically significant changes at 95% level; **c** difference in DJF surface temperature (°C) between **a** and **b**, representing the approximate response to the cooled ocean surfaces prescribed in YD2. Contours at 30, − 15, − 10, − 5, − 2, − 1, 1 and 2 °C. *Shading* shows statistically significant changes at 95% level

(1987) suggested that the temperature depression was related to the reduction of the atmospheric mass above the elevated ice sheets, thus reducing the greenhouse capacity of the atmosphere. Presumably, this effect also took place in the YD1 and YD2 experiments, thus producing the cooling of 5 °C. The prescribed Scandinavian ice sheet was apparently too low to cause a significant cooling. It is noteworthy that R86 simulated a much stronger DJF cooling over the Laurentide and Scandinavian ice sheets. Apparently, the more pronounced land ice topography in the experiments of R86 caused the noted difference.

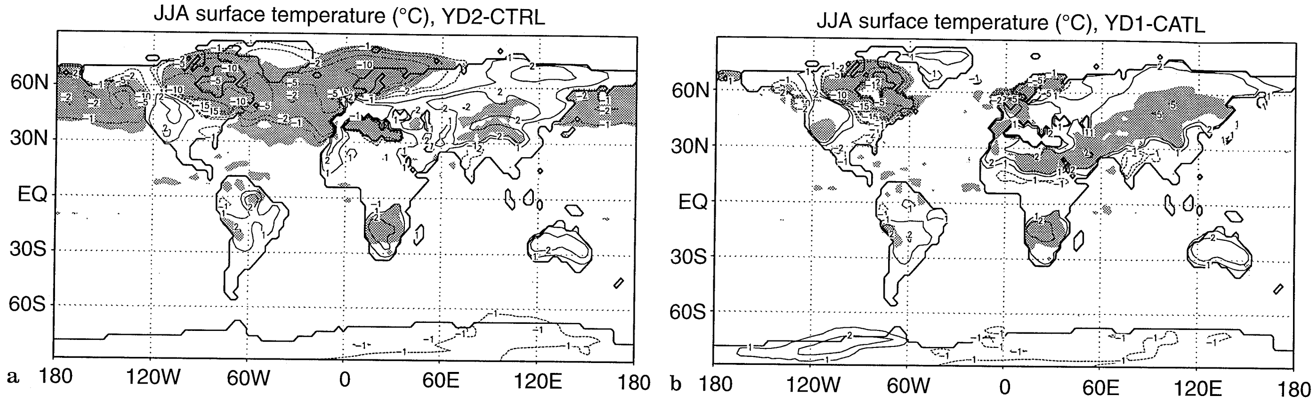
Figure 2c represents the effect of cooled ocean surfaces. It should be noted that climatological SSTs without interannual variability were prescribed in CTRL. Hence, it is expected that lowering of the SSTs in the YD experiments caused statistically significant cooling over the ocean surface. Figure 2c shows also a large significant cooling of 2 to 15 °C in Europe, northernmost Africa, Atlantic Canada and Greenland. A similar pattern of cooling was present in the result of R86, but with less depressed temperatures. Presumably, the difference in temperature was caused by absence of an extended sea-ice cover in the experiments of R86. In the YD2 result DJF temperatures were depressed by more than 30 °C over sea-ice against 10 °C over the Atlantic without sea-ice in R86.

JJA

The JJA results of YD2 show a cooling near the N Atlantic and N Pacific Oceans and over the Laurentide and Scandinavian ice sheets (see Fig. 3a). Away from these regions a warming of 1 to 2 °C is noted over most continental areas. This warming is statistically significant at a 95% level in central Asia, S Africa and parts of S America. The latter effect was produced by the ice sheets and insolation, since the warming is much more extensive in the YD1-CATL result (Fig. 3b). In the 11 ky BP and LGM experiments of R86 and Rind (1987) considerable summer warming was present in SE Europe. Rind (1987) suggested that this warming was produced by the elevated Scandinavian ice sheet through subsidence of air. In the YD1-CATL result no statistically significant warming was apparent in SE Europe, so apparently this effect did not play an important role. Probably the prescribed ice sheet elevation was too low to influence the atmospheric circulation to this extent.

Other influences of ice sheets are however apparent. First, it is likely that part of the JJA warming in western N America and the small cooling in southeast N America was caused by anticyclonic winds developing over the Laurentide ice sheet. This effect was also found in sensitivity experiments with LGM ice sheets (e.g. Rind 1987). Second, it is clear from Fig. 3b that the Laurentide and Scandinavian ice sheets caused a strong cooling (5 to 15 °C) over their surfaces. Part of this cooling may be caused by the increased elevation of these ice sheets compared to CTRL. However, Rind (1987) suggests that cooling during summer results mainly from the high albedo of the ice surfaces.

As the newly prescribed ice sheets apparently did not cause significant warming, the main increase in



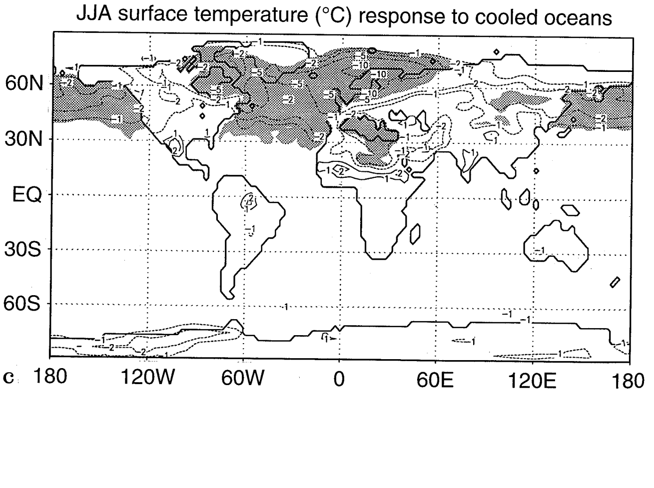


Fig. 3. a Difference in JJA surface temperature (°C) between the experiments YD2 and CTRL, representing the response to the total set of YD boundary conditions: cooled ocean surfaces, insolation, land ice and atmospheric CO_2 concentration. Contours at − 10, − 5, − 2, − 1, 1, 2 and 5 °C. *Shading* shows statistically significant changes at 95% level; **b** difference in JJA surface temperature (°C) between the experiments YD1 and CATL, representing the combined response to 12 cal ky BP insolation, land ice and atmospheric CO_2 concentration. Contours at − 10, − 5, − 2, − 1, 1, 2 and 5 °C. *Shading* shows statistically significant changes at 95% level; **c** difference in JJA surface temperature (°C) between **a** and **b**, representing the approximate response to the cooled ocean surfaces prescribed in YD2. Contours at − 10, − 5, − 2, − 1, 1, 2 and 5 °C. *Shading* shows statistically significant changes at 95% level

temperature seen over most continents (Fig. 3b) was a response to the changed JJA insolation. Moreover, since in YD2 the warming was less pronounced, the insolation effect in this simulation was counteracted by the effect of cooled ocean surfaces. The latter response was quantified in Fig. 3c, showing lower temperatures (depressed by 1 to 2 °C) downwind of the cooled oceans. A maximum cooling of 10 °C is seen over Scandinavia. A similar response to a cooled ocean surface was found in the 11 ky BP experiments of R86. In short, the increased insolation caused widespread warming over land surfaces, which was partly counteracted in YD2 by the cooling effect of ice sheets and lowered SSTs.

3.3 Zonal wind speed

DJF

In Fig. 4a, b statistically significant changes of the zonal wind speed (westerly winds are positive) are found over the African west and east coast and over northeast N America. In the latter region a statistically significant increase of up to 3 m/s was noted in between the Laurentide and Greenland ice sheets. A similar effect was found in several LGM simulation studies (e.g. Kutzbach and Guetter 1986; Rind 1987; Lautenschlager and Herterich 1990) and was attributed to the Laurentide ice sheet. The second important change in zonal wind speed occurred over the tropical coasts of Africa. In W Africa a decrease of more than 1 m/s is present, showing the strengthening of the trade winds. The increase on the African east coast is an expression of a more southerly direction of the surface winds in this region (not shown). These changes were probably related to the marked surface cooling of 2 °C caused by the insolation decrease during DJF (see Sect. 3.2). As suggested by Kutzbach and Webb (1993), the surface cooling caused an intensified winter monsoon circulation with strengthened off-shore winds. It is noteworthy that these noted regional changes near Africa were absent in low resolution AGCM studies on Late Glacial climate (Rind et al. 1986; Kutzbach et al. 1993).

The plot with the DJF difference between Fig. 4a and b, representing the response to cooled oceans, shows decreased zonal wind speeds (by more than 3 m/s) over the N Atlantic Ocean (Fig. 4c). This decrease is the effect of a stabilised atmosphere over a cooled ocean partly covered with sea-ice (Rind 1987; Renssen et al. 1996). It is interesting that this stabilising effect works opposite to the intensifying effect of the Laurentide ice sheet noted above. A similar phenomenon was seen in the results of R86.

JJA

Significant changes in the JJA zonal winds are present in two regions: near the Laurentide ice sheet and at several places in the tropics (Fig. 5a, b). Decreases of 1 m/s to the south and increases of up to 2 m/s to the north of the Laurentide ice sheet signified the existence of a glacial anticyclone. Such anticyclonal winds were present in most

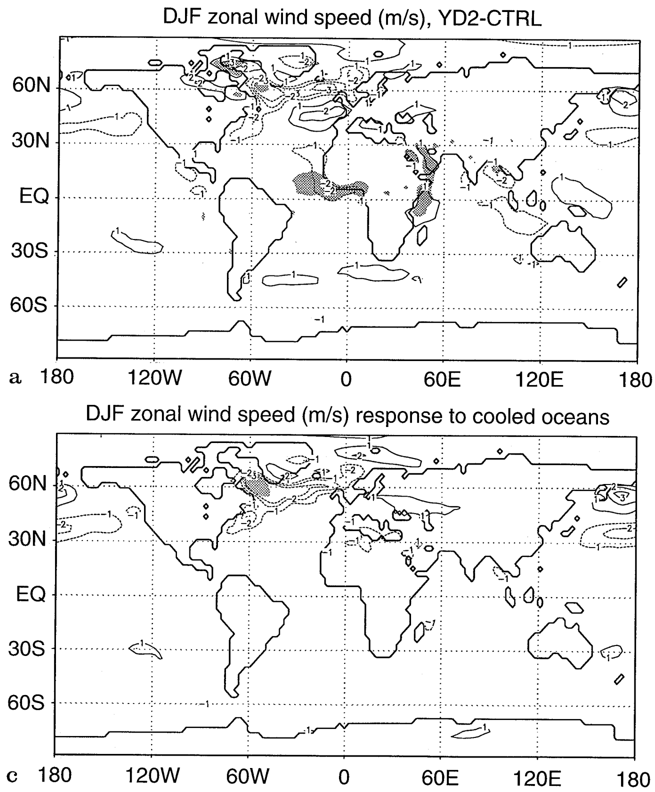
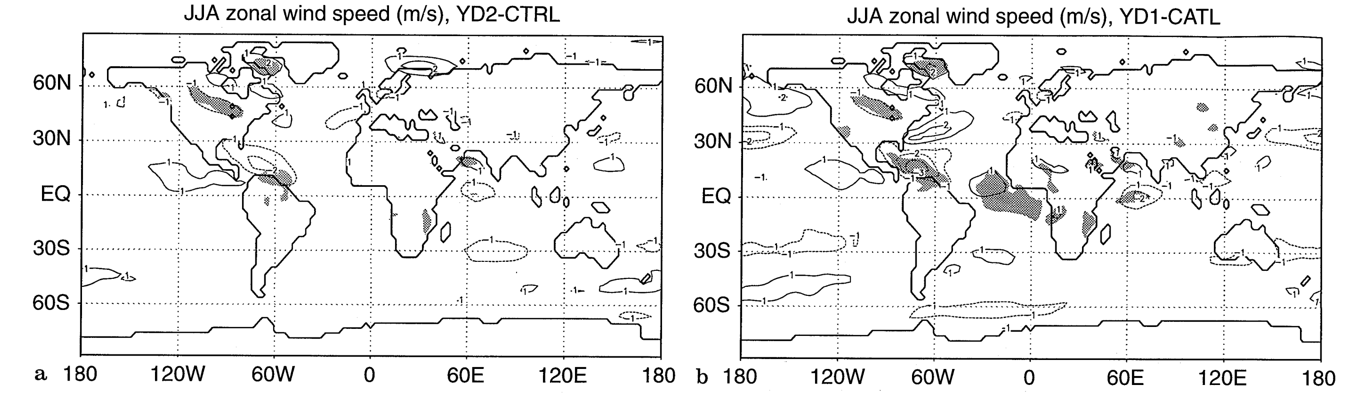


Fig. 4. a Difference in DJF zonal wind speed (m/s) between the experiments YD2 and CTRL, representing the response to the total set of YD boundary conditions: cooled ocean surfaces, insolation, land ice and atmospheric CO_2 concentration. Contours at − 3, − 2, − 1, 1, 2 and 3 m/s. *Shading* shows statistically significant changes at 95% level; **b** difference in DJF zonal wind speed (m/s) between the experiments YD1 and CATL, representing the combined response to 12 cal ky BP insolation, land ice and atmospheric CO_2 concentration. Contours at − 3, − 2, − 1, 1, 2 and 3 m/s. *Shading* shows statistically significant changes at 95% level; **c** difference in DJF zonal wind speed (m/s) between **a** and **b**, representing the approximate response to the cooled ocean surfaces prescribed in YD2. Contours at − 3, − 2, − 1, 1, 2 and 3 m/s. *Shading* shows statistically significant changes at 95% level



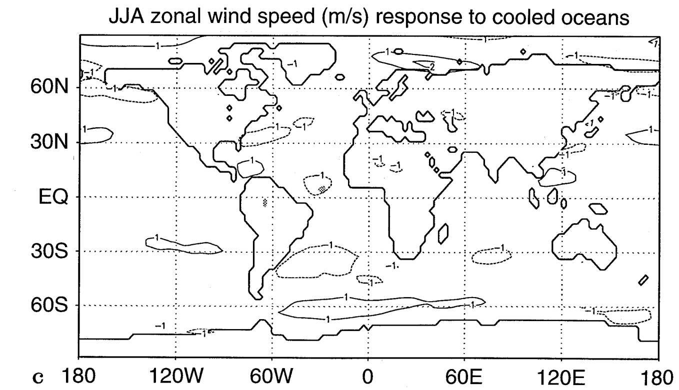


Fig. 5. **a** Difference in JJA zonal wind speed (m/s) between the experiments YD2 and CTRL, representing the response to the total set of YD boundary conditions: cooled ocean surfaces, insolation, land ice and atmospheric CO_2 concentration. Contours at -3, -2, $-1, 1, 2$ and 3 m/s. *Shading* shows statistically significant changes at 95% level; **b** as **a** but for experiments YD1 and CATL, representing the combined response to 12 cal ky BP insolation, land ice and atmospheric CO_2 concentration. Contours at -3, -2, $-1, 1, 2$ and 3 m/s. *Shading* shows statistically significant changes at 95% level; **c** difference in JJA zonal wind speed (m/s) between **a** and **b**, representing the approximate response to the cooled ocean surfaces prescribed in YD2. Contours at -3, -2, $-1, 1, 2$ and 3 m/s. *Shading* shows statistically significant changes at 95% level

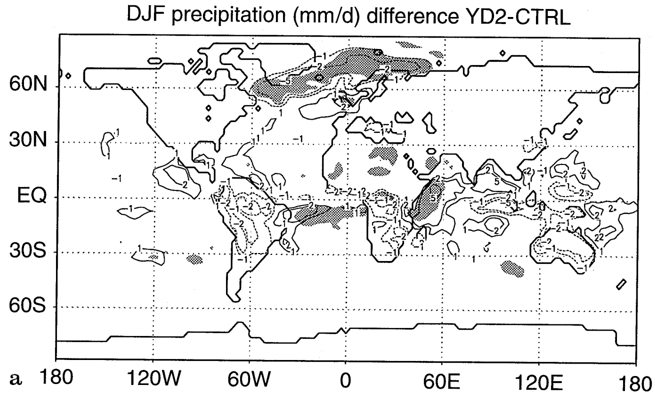
simulation studies on the LGM climate (e.g. Kutzbach and Guetter 1986; Lautenschlager and Herterich 1990). The LGM experiments of Rind (1987) suggested that these anticyclonal winds were mainly produced by the elevation of the Laurentide ice sheet. At tropical latitudes the JJA zonal wind speed changes substantially in the Caribbean Sea (a decrease of 3 m/s), over the Atlantic Ocean near the equator and over the northern Indian Ocean (YD1-CATL result, Fig. 5b). These changes are probably related to a strengthening of the summer monsoon, which was a response to the increased insolation (Kutzbach and Webb 1993). The increase over the tropical Atlantic was not present in the figure with the overall response to YD boundary conditions (Fig. 5a), suggesting an opposite response to the lowered SSTs. Just as in the DJF case, a marked strengthening of winds over tropical oceans was absent in the other Late Glacial AGCM studies (Rind et al. 1986; Kutzbach et al. 1993). According to Fig. 5c, the cooling of the oceans caused no statistically significant variations in the JJA zonal wind speeds.

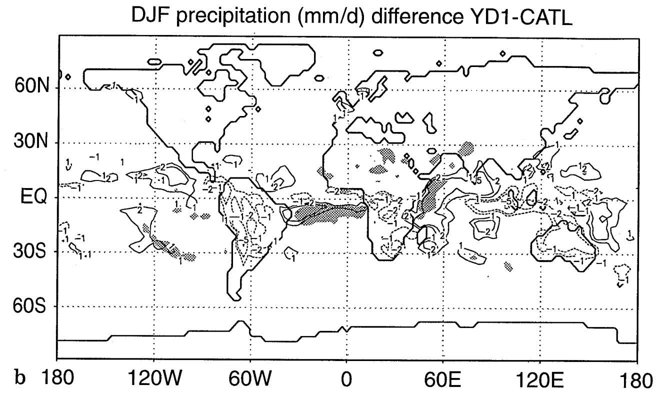
3.4 Precipitation

DJF

The combined YD boundary conditions caused two distinct significant effects in the DJF precipitation result (Fig. 6a). First, a decrease in DJF precipitation of 2 to 5 mm/d is present over the northernmost Atlantic and second, considerable changes (of more than 2 mm/d) with a high variability are present in the tropics. The latter effect returns in Fig. 6b and is therefore presumably a response to the decreased insolation. Generally, in the tropics the precipitation increased over ocean surfaces and decreased over the continents. Furthermore, the centre of these changes is situated at 5 °S, so approximately at the latitude where the ITCZ is expected during NH winter. Although the changes in temperature in this region are not statistically significant, this response may be explained by the changes in monsoonal circulation. According to Kutzbach and Webb (1993) the reduced insolation caused an extra cooling of the continents, producing strengthened winter monsoons. This was characterised by generally more descending air and a decrease of precipitation over land surfaces and rising air masses and increase of precipitation over the oceans. The noted response of precipitation to the decrease insolation is confirmed by the 9 and 12 ky BP simulation results of Kutzbach et al. (1993), showing a similar trend. The other marked change, viz. the one noted over the northernmost Atlantic Ocean, is not present in Fig. 6b, suggesting that this decrease in precipitation was a response to the cooled oceans (see Fig. 6c). As reported earlier (Renssen et al. 1996), the strong cooling over the expanded sea-ice cover caused a stabilisation of the lower atmosphere, leading to a decrease in cyclonic activity and a reduction of precipitation.

In Fig. 6c small insignificant changes are present in the tropics. Presumably, these changes were caused by the insolation and not by the lowered SSTs (compare with Fig. 6b), thus suggesting that the tropical response to the changed insolation is not similar in the experiments YD1 and YD2. Apparently, the basic assumption that the





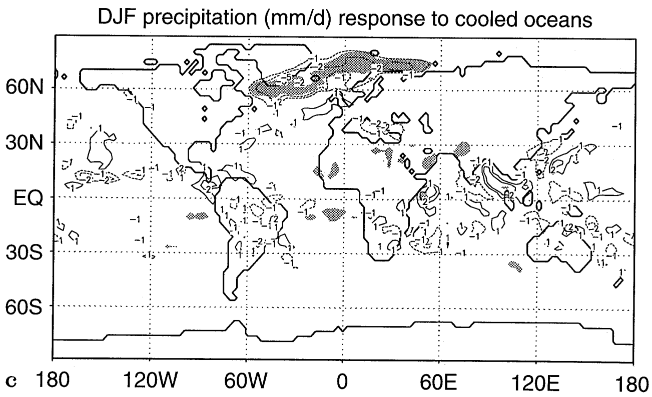


Fig. 6. a Difference in DJF precipitation (mm/d) between the experiments YD2 and CTRL, representing the response to the total set of YD boundary conditions: cooled ocean surfaces, insolation, land ice and atmospheric CO_2 concentration. Contours at -5, -2, -1, 1, 2 and 5 mm/d. *Shading* shows statistically significant changes at 95% level; **b** as **a** but for experiments YD1 and CATL, representing the combined response to 12 cal ky BP insolation, land ice and atmospheric CO_2 concentration. Contours at -5, -2, -1, 1, 2 and 5 mm/d. *Shading* shows statistically significant changes at 95% level; **c** difference in DJF precipitation (mm/d) between **a** and **b**, representing the approximate response to the cooled ocean surfaces prescribed in YD2. Contours at -5, -2, -1, 1, 2 and 5 mm/d. *Shading* shows statistically significant changes at 95% level

overall effect of boundary conditions is the sum of the individual effects, is not valid in this case. This is the result of the high spatial variability of the precipitation.

JJA

The overall response of JJA precipitation is rather similar in Fig. 7a, b. Generally, the reverse picture emerged compared to the DJF situation, with an increase of up to 5 mm/d over tropical continents and a reduction of a similar magnitude over oceans. Again the main belt of changes is at the latitude where the ITCZ is expected during JJA. This fits in also with the concept of Kutzbach and Webb (1993). A slight warming was apparently sufficient to cause a strengthening of the summer monsoons, producing more rainfall over land. This notion is supported by Kutzbach et al. (1993), who found a comparable precipitation pattern in their 12 ky BP experiment due to the increased insolation. These variations in precipitation were consistent with the noted tropical changes in zonal wind speed during JJA. In the mid-latitudes the discussed anticyclonal circulation over the Laurentide ice sheet also had a small effect. A patch with reduced precipitation (1 mm/d decrease) is present in northeast N America and is presumably caused by a cold and dry northerly flow. As shown in Fig. 7c, the changed SSTs had no significant effect on the precipitation. However, analogous to the DJF case, in Fig. 7c several changes were present in the tropics that can supposedly be attributed to the prescribed insolation.

3.5 Comparison of the response to cooled oceans with geological evidence of YD climate change

In this section the following question is answered: to what extent does the model response to the cooled oceans explain global geological evidence of climate change during YD time? It is recalled from Sect. 3.1 that in comparing the simulation results with geological evidence, one has to take the response to the cooled ocean surfaces into consideration. The other YD boundary conditions used in the experiments probably had only a minor influence on the YD *oscillation*. The temperature response to cooled ocean surfaces was characterised by a down wind cooling over the adjacent continents, which was statistically significant over W Europe, Greenland, Atlantic Canada, N Africa and the Pacific coasts. The zonal winds showed significant changes over the northernmost Atlantic Ocean. In this region also a significant reduction of the precipitation occurred during NH winter. These model results are compared with YD climate reconstructions based on geological data, while referring to the LGI and start of the Holocene.

In **Europe** extensive evidence for a YD oscillation was found in the geological records. According to climate reconstructions based on these geological data, the average winter temperatures during YD time were as low as -15 to $-20\,^{\circ}$C in NW Europe (Atkinson et al. 1987; Walker et al. 1994). As a winter temperature of about $0\,^{\circ}$C was reconstructed for the start of the LGI in England, this implies a cooling of at least $15\,^{\circ}$C within the Late Glacial period. Geological evidence was also used to reconstruct

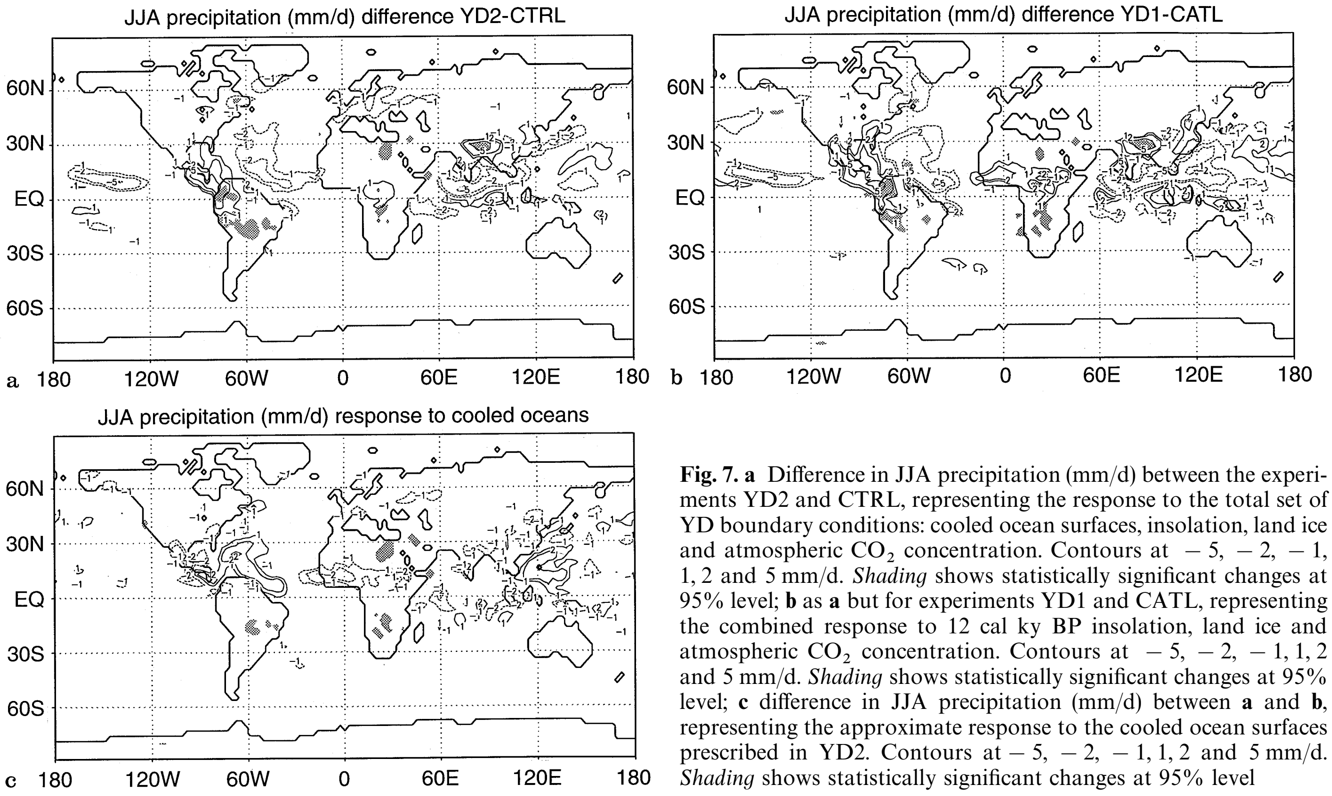


Fig. 7. a Difference in JJA precipitation (mm/d) between the experiments YD2 and CTRL, representing the response to the total set of YD boundary conditions: cooled ocean surfaces, insolation, land ice and atmospheric CO_2 concentration. Contours at -5, -2, -1, 1, 2 and 5 mm/d. *Shading* shows statistically significant changes at 95% level; **b** as **a** but for experiments YD1 and CATL, representing the combined response to 12 cal ky BP insolation, land ice and atmospheric CO_2 concentration. Contours at -5, -2, -1, 1, 2 and 5 mm/d. *Shading* shows statistically significant changes at 95% level; **c** difference in JJA precipitation (mm/d) between **a** and **b**, representing the approximate response to the cooled ocean surfaces prescribed in YD2. Contours at -5, -2, -1, 1, 2 and 5 mm/d. *Shading* shows statistically significant changes at 95% level

summer conditions during the YD. July temperature estimates varied from 4 °C in N Norway, to 7 °C in Ireland, 10 °C in England and The Netherlands and 14 °C in Switzerland (Lowe et al. 1994). These temperatures translate in coastal regions to a YD summer cooling of about 5 to 8 °C compared to the warmest phase of the LGI. Further inland (Switzerland) cooling was less pronounced, with only a few degrees lowering of the temperature. Therefore, these reconstructions suggest a strong temperature gradient inland from the coast.

The simulated response to cooled ocean surfaces agrees with the above estimates for Europe. During winter the model response to lowered SSTs produced a cooling ranging from 30 °C in northernmost Europe, to 15 °C in Scotland and S Scandinavia and 10 °C in England and The Netherlands. These figures correspond with the reconstructed winter cooling of at least 15 °C. During summer the effect of lowered SSTs in the model ranged from a 10 °C depression in N Scandinavia to a 5 °C cooling in W Europe. Again these values are similar to the reconstructed July estimates of a 8 to 5 °C cooling. Even the noted temperature gradient going inland was reproduced, as in central Europe the simulated response to a cooled ocean was a 2 °C cooling. In conclusion, the agreement between the estimates of a YD oscillation and the model response to lowered SSTs strongly suggests that the YD signal embedded in European geological records was produced by a cooling of the Atlantic Ocean, including an extension of sea-ice cover.

Summer temperature reconstructions in **Atlantic Canada** point to a cooling of a few degrees compared to preceding and following periods (Lowe et al. 1994). This cooling was present in the simulation results as a response to lowered SSTs. Consequently, the simulations explain the existence of a YD oscillation in Atlantic Canada as an effect of a cooled N Atlantic Ocean.

The **northernmost Atlantic Ocean** was the third region with statistically significant response to lowered SSTs. The simulation results are compared with the Greenland ice core records, containing ample evidence of YD climate change. Dansgaard et al. (1989) reconstructed for the Dye 3 core a 7 °C rise in temperature for the YD to Holocene transition. Additionally, Grootes et al. (1993) estimated a temperature drop of 10 °C during the YD in the GISP2 ice core. Moreover, a sharp decrease in YD snow accumulation was found by Alley et al. (1993). Furthermore, Mayewski et al. (1994) concluded on the basis changes in the dust record that the atmospheric circulation pattern was very different during the YD compared to the LGI and Holocene. These examples of a different climate during the YD were fully explained by the simulated response to a cooled ocean surface. The reconstructed annual temperature depression of 7 to 10 °C agrees with the simulations for S Greenland, producing a winter cooling of 15 °C and a summer cooling of 5 °C. Similarly, the reduction in snow fall is consistent with the simulated decrease in winter precipitation. Finally, the reconstructed change in atmospheric circulation pattern may be explained by the simulated southward shift of the surface westerlies (see also Renssen et al. 1996).

Most geological evidence from **Northern Africa** point to the YD event as an arid period in between relatively wet episodes (e.g. Rossignol-Strick et al. 1982; Roberts et al. 1993). This observation was not explained by the simulation results, since the response to a cooled ocean surface showed no significant decrease in precipitation in N Africa. The lowered SSTs caused however a small cooling. One could argue that this cooling could weaken the strengthening of the summer monsoons, consequently causing less precipitation.

Several authors have reconstructed a YD event in the **Pacific region**. For instance, Mathewes (1993) reconstructed a cool wet phase during YD time at the N American west coast. Similarly, Engstrom et al. (1990) found a climatic oscillation in Alaska. In the simulation experiments a cooling of 1 to 2 °C was prescribed in the N Pacific Ocean and as a result the surrounding areas experience a comparable temperature reduction. Away from the coast this cooling is, however, not statistically significant. Still, the simulation results suggest that a YD oscillation found in geological records in the Pacific region may be explained by a cooling of the ocean surface.

Reports of a YD oscillation in other places (i.e. the tropics and the middle and high latitudes of the S Hemisphere) are not explained by the presented model results. Although a strengthened monsoon circulation was present in the YD2 result in agreement with various reports (e.g. Sirocko et al. 1993; Beveridge 1994; Porter and Zhisheng 1995), this was a model response to the prescribed insolation. Therefore, this is not a representation of a YD signal in the simulation results, since the insolation was approximately similar during preceding (LGI) and following (early Holocene) periods.

This comparison of YD simulation results with climate reconstructions based on geological evidence leaves one important question open: what mechanism caused the YD cooling? The response of experiment YD2 to a N Atlantic Ocean without a thermohaline circulation during winter explained the YD signals recorded in geological data in the middle and high latitudes of the N Hemisphere. This result is consistent with the theory of Broecker (1992), explaining the YD by a shut-down of the thermohaline circulation under influence of meltwater fluxes. Still, analyses from ocean cores show that the thermohaline circulation was functioning during YD time, suggesting that the YD cooling of the N Atlantic was caused by another mechanism (e.g. Veum et al. 1992; Sarnthein et al. 1994). This mechanism is still controversial and it may be speculated that the internal variability of the coupled atmosphere-ocean-ice system or additional forcings played a role.

As discussed in Sect. 2, the degree of winter cooling in the N Atlantic derived from geologically derived SST estimates was insufficient. It is proposed that the latter inconsistency was due the uncertainties in the SST reconstruction method. This may imply that in future improved estimates of N Atlantic SSTs may show a YD ocean cooling similar in magnitude to conditions without a thermohaline circulation, as used in YD2 for the winter season.

In the N Pacific Ocean an arbitrary cooling of 2 °C was prescribed in YD2 in accordance with inferences from Kallel et al. (1988) and Kennett and Ingram (1995). This lowering of the SSTs produced a simulation result in YD2 that explains the terrestrial cooling reconstructed for the Pacific region. However, it has to be noted that the amount of cooling (i.e., 2 °C) is speculative. Kennett and Ingram (1995) suggest that the most likely mechanism to explain a synchronous cooling of both the N Atlantic and N Pacific Oceans is through transmittance of the climate change signal through the atmosphere.

Additional factors that may have played a role in forcing the YD oscillation are the atmospheric concentrations of dust and methane. Indeed, these boundary conditions were not considered in the described YD experiments. The Greenland ice core dust record shows that the YD aerosol content was higher than during the Holocene and LGI in the N Atlantic region (e.g. Mayewski et al. 1994). Dust records from ice cores from the tropics (e.g. Thompson et al. 1995) and Antarctica (e.g. Jouzel et al. 1995) show no significant increase during YD time. Consequently, the higher aerosol possibly contributed to the YD signal in the N Atlantic region. Also, according to ice core analyses (e.g. Brook et al. 1996), the atmospheric methane content was lower during YD time (550 ppb) than during the Holocene (700 ppb). Since methane is a powerful greenhouse gas, it is likely that this contributed to the YD oscillation on a global scale.

The lack of a YD signal in regions away from the influence of the cooled oceans, as evident from the presented YD simulation, was also found by R86. However, this lack contrasts with reports from high latitudes of the S Hemisphere, e.g. New Zealand (Denton and Hendy 1994), suggesting a global YD oscillation. The fact that the presented results and the experiments of R86 do not explain such a global response, could have two causes. First, it could imply that the boundary conditions prescribed in these simulations are insufficient. Indeed, little is known about the YD sea surface conditions in the S Hemisphere. Moreover the effect of omitted boundary conditions (methane, dust) is unknown. Second, an atmospheric model with fixed land ice and SSTs could be unsuitable to simulate a global YD climate. The recent discovery of ice rafted sediments in N Atlantic cores of YD age and the subsequent linkage of the YD to the Heinrich events, suggests that feedback processes between atmosphere, ocean and ice played an important role. Therefore, a future experiment on the YD with a coupled ocean-atmosphere-ice model would be very useful.

4 Conclusions

1. The results of an AGCM experiment with Younger Dryas boundary conditions (cooled oceans, land ice, insolation and atmospheric CO_2 content) were evaluated. It is inferred that the prescribed 12 ky BP insolation produced the following significant results: widespread summer warming of 2 °C, winter cooling of 2 °C in the tropics, intensified monsoonal circulation with associated variations in winds and precipitation in the tropics. Similarly, it is inferred that the defined 12 ky BP ice sheets produced primarily summer cooling of 5 to 15 °C in their vicinity. Moreover, the Laurentide ice sheet influenced the surface circulation substantially, with the development of a glacial anticyclone during summer and a strengthening of the winds to the northeast during winter.

2. The cooled N Atlantic and N Pacific Oceans caused a significant cooling of 2 to 15 °C of the down-wind continents. The extended N Atlantic sea-ice cover proved especially important. Over its surface the air cooled as much as 30 °C, causing a stabilisation of the lower atmosphere, a weakening of the westerlies and a decrease of precipitation.

3. The YD climatic oscillation evidenced by geological records in Europe, Atlantic Canada, Greenland and along Pacific shores is completely explained by the model response to cooled N Atlantic and N Pacific Oceans. However, during winter, the prescribed N Atlantic cooling represented model-derived conditions without a thermohaline circulation which is not in accordance with evidence from ocean cores. A model-derived SST set was used because estimates based on foraminifera proved to show insufficient winter cooling. It is proposed that the latter discrepancy is caused by uncertainties in the SST reconstruction method. Moreover, the prescribed cooling of the N Pacific Ocean was speculative.

4. Reports of a YD oscillation in areas away from the influence of the N Atlantic and N Pacific Oceans were not explained by the presented AGCM results. This conclusion is similar to the one reached by Rind et al. (1986), although a second generation AGCM and an updated set of boundary conditions were used in the present study. This suggests that either the set of boundary conditions used was insufficient or that an atmospheric model is unsuitable for a simulation of a global YD climate. The latter would be related to the importance of feedback processes with oceans and land ice, which are poorly represented in the used model. The ultimate cause of the Younger Dryas oscillations is still uncertain and needs to be studied in future.

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