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The atmospheric winter circulation during the Younger Dryas stadial in the Atlantic/European sector

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Abstract. The Younger Dryas (YD) stadial signified an interruption of the warming during the transition from the last glacial to the present interglacial. The mechanism responsible for this cooling is still uncertain, so valuable information concerning climate variability can be obtained by numerical simulation of the YD climate. We performed four experiments on the Younger Dryas climate with the Hamburg atmospheric general circulation model. Here we use the results of these experiments, which differed in prescribed boundary conditions, to characterize the atmospheric winter circulation during the YD stadial in the North Atlantic/European sector. The 10 year means of the following variables are presented: sea level pressure, 500 hPa geopotential heights and 200 hPa winds. In addition, we used daily values to calculate an index to assess the occurrence of blocking and strong zonal flow and to compute storm tracks. Our results show that the YD cooling in Europe was present with a strong and stable westerly circulation without blocking. This is in conflict with an earlier study suggesting frequent easterly winds over NW-Europe. In our experiments the sea-ice cover in the North Atlantic Ocean was the crucial factor forcing this specific YD circulation. Moreover, the jet stream over the North Atlantic was strengthened considerably, causing an enhanced cyclonic activity over the Eurasian continent. The YD winter circulation was different from the circulation found in most simulation studies on the Last Glacial Maximum, since no glacial anticyclones were present and no split of the jet stream occurred.

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1 Introduction

Geological records from the circum-North Atlantic region give evidence of an irregular transition from the last glacial to the present interglacial climate. This transition phase, known as the Weichselian Late Glacial, lasted from approximately 14.5 to 11.5 cal ky BP (thousand calendar years before present). A relatively warm stage existed (Bølling-Allerød interstadial complex, BA) from 14.5 to 12.5 cal ky BP, which was interrupted by several short cool episodes. From 12.5 to 11.5 cal ky BP temperature dropped to almost glacial levels (Younger Dryas stadial, YD), before the warming at the start of the Holocene. Here we focus on the numerical simulation of the YD climate, realizing that the ultimate cause of this severe cooling is still uncertain (Berger 1990).

During the Younger Dryas stadial, boundary conditions forcing climate were mostly transitional between glacial and interglacial situations. The main difference in surface conditions between the BA and YD stages was the temperature of the North Atlantic Ocean. During the YD event the Atlantic was certainly cooler than today, but not as cold as during the Last Glacial Maximum (LGM, 21 cal ky BP) (Schulz 1995; Sarnthein et al. 1995). Furthermore, the Late Glacial Laurentide- and Scandinavian ice sheets were considerably reduced in height and extent. According to the COHMAP members (1988), the land-ice volume around 12 cal ky BP was about 50% of the LGM value. Another important boundary condition, the atmospheric CO_2 content, was also at a level (230 ppmv) falling between the full glacial (200 ppmv) and the pre-industrial concentration (265 ppmv) (Barnola et al. 1987). One factor does not fit in this picture: the solar radiation. During the LGM the received solar radiation in the Northern Hemisphere was nearly the same as at present. In contrast, solar variations around the time of the Late Glacial increased the difference between the summer- and winter seasons, since insolation in the Northern Hemisphere increased by 7% in summer and decreased by 7% in winter (COHMAP members 1988).

Detailed observational information concerning the effect of these boundary conditions on the atmosphere during Late Glacial is provided by geological records. In NW Europe, fossil dunes of Younger Dryas age indicate westerly winds during formation (e.g., Böse 1991). Today, a dominantly westerly circulation brings maritime air from the North Atlantic to Europe and it is therefore associated with a relatively mild climate. Hence, assuming that these YD dunes imply a dominantly westerly circulation, a conflict exists with the permafrost conditions during YD, evidenced by numerous periglacial data in Northern Europe (e.g., Bohncke 1993; Berglund et al. 1994). The dust record in Greenland ice-cores also contains information on the atmospheric circulation. The influx of dust was relatively low during warmer periods such as BA and Holocene, and high during cold episodes such as YD and LGM (Mayewski et al. 1993). The high dust content during cold periods has been attributed to an enlarged polar atmospheric cell and to expanded source regions. Thus, geological records suggest on the one hand a westerly circulation as presently found in Europe and on the other hand an atmospheric circulation considerably different from the modern one.

Although geological records provide important information about the atmosphere during the Late Glacial, the overall picture remains incomplete, as the relationship between the findings at different sites is ambiguous. Nevertheless, understanding the general functioning of the atmospheric circulation during a transitional phase such as the Late Glacial is essential for a better insight into climate variability. This insight can be improved by means of simulations with atmospheric general circulation models (AGCMs), as these models provide the opportunity to analyze the atmospheric circulation in detail. Most AGCM simulation studies on former climates are focused on the LGM (e.g., Kutzbach and Wright 1985; Broccoli and Manabe 1987; Lautenschlager and Herterich 1990). In these studies the reconstructions of the earth's surface as composed by CLIMAP project members (1981) were used. The Late Glacial is the subject of a few simulation studies. The most comprehensive one was performed by Rind et al. (1986), who carried out several AGCM experiments with boundary conditions of Late Glacial time. However, as the prescribed ice sheets and surface ocean conditions are crude estimates, these experiments must be regarded as sensitivity studies rather than actual simulations of the Late Glacial climate. An AGCM study aimed at the simulation of this climate is therefore very useful, since it would improve our understanding of climate variability considerably.

We performed four experiments on the Younger Dryas climate with the Harmburg atmospheric general circulation model. The aim of the present study is to characterize the atmospheric winter circulation in the North Atlantic/European sector. This region is chosen because the Younger Dryas signal is most clearly found here. A detailed comparison with paleoclimatic reconstructions is to be published elsewhere. The first three simulations are sensitivity experiments that are only used in this work to infer the impact of the boundary conditions. These experiments are fully described in Renssen et al. (1995). Here we focus on a fourth experiment that is most likely to resemble the YD climate. We look at the circulation at three levels: surface, middle troposphere and jet stream level.

2 Experimental design

Four experiments were carried out with the ECHAM3 (European Centre-HAMburg) AGCM in T42 configuration, which corresponds to a Gaussian grid of approximately 2.8° latitude-longitude. This spectral model adequately simulates most aspects of the observed large-scale time-mean circulation and its intraseasonal variabilities (Roeckner et al. 1992). For a detailed model description the reader is referred to DKRZ-report number 6 (1993). The experimental design is summarized in Table 1.

In a control experiment (hereafter called CTRL) present boundary conditions were prescribed. The results of CTRL are close to observations of present climate (Roeckner et al. 1992). The control experiment is used as a standard with which the results of the other three experiments will be compared.

In a second experiment we cooled the North Atlantic Ocean according to geological evidence (hereafter called CATL). The aim of this sensitivity experiment was to study the effect of a North Atlantic with Younger Dryas sea surface temperatures (SSTs) on present climate. The set of SSTs used was derived from Sarnthein et al. (1995), who reconstructed YD sea surface temperatures from several marine cores in the North Atlantic Ocean. These reconstructions are based on the relation between the present distribution of planktonic foraminifera species and the surface conditions in the ocean. We fitted a sine function through the provided winter and summer SST estimates to obtain an annual cycle. This procedure was also used in LGM simulations using CLIMAP SSTs (e.g., Rind 1987; Lautenschlager and Herterich 1990). In all our experiments a climatological annual cycle of SSTs was used without an interannual variability. We assumed that the surface conditions in the tropical Atlantic were not significantly different from today. This assumption agrees with SST reconstructions of Schulz (1995). The sea-ice limits in the Greenland-Iceland-Norway (GIN) seas were based on Koç et al. (1993). From the western North Atlantic Ocean fewer data were available and the chosen sea-ice limit is therefore more speculative.

The aim of the third experiment was to investigate the response of the model to a more complete set of YD boundary conditions (hereafter called YD_A). We used the same SST set as in CATL, but with an additional cooling north of 40°N in the Pacific Ocean. Although there are indications that the North Pacific Ocean was cooled during the Younger Dryas (e.g., Kallel et al. 1988), no set of reconstructed SSTs exists for this region. We, therefore, specified an arbitrary cooling of 2 °C here. Additionally, we altered other im-

Table 1. Overview of performed AGCM experiments with the most important boundary conditions mentioned

Boundary conditions	Experiments			
	CTRL	CATL	YD_A	YD_B
SSTs and sea-ice	Present	YD (Atlantic)	YD (Atlantic + Pacific)	Summer: as YD_A Winter: model output (conveyor off)
Land ice	Present	Present	YD	YD
Insolation	Present	Present	YD	YD
CO_2 (ppmv)	345	345	230	230

CTRL is a control simulation of present climate, CATL is a sensitivity experiment with a cooled N-Atlantic and YD_A and YD_B are simulations of YD climate. YD_A and YD_B differ only in prescribed N-Atlantic surface conditions in winter

portant boundary conditions. Ice sheets were introduced according to Peltier (1994), with the most important being the Laurentide and Scandinavian ice sheets. The surface albedo of these newly defined ice sheets was set to 0.8. A value varying between 0.6 and 0.8 was used for the ice sheets of Greenland and Antarctica. We introduced additional land points in the Bering Strait and North Sea areas, which agrees with sea level curves. To these new land points the surface albedo of adjacent land points was assigned. The modern surface albedo was used for the remaining points. Furthermore, the atmospheric CO_2 concentration was set to 230 ppmv, as suggested by Barnola et al. (1987) on the basis of measurements in an Antarctic ice-core. The insolation was calculated according to Berger (1978). A comparison with reconstructions based on geological records revealed that experiment YD_A produced too high winter temperatures in Europe (Renssen et al. 1995).

In the present study we will concentrate on a fourth experiment (YD_B). In YD_B the N Atlantic surface conditions in winter were redefined, because the too high air temperatures in experiment YD_A were caused by the prescribed SST set (Renssen et al. 1995). We left the summer conditions the same as in YD_A. For the new winter SST set we used the output of a coupled ocean-atmosphere model (the Hamburg large-scale geostrophic ocean model coupled to the atmospheric ECHAM3-T21 model) in which the conveyor belt was turned off by introducing a large amount of fresh water in the North Atlantic (Schiller et al. 1996). We used the ocean-model output to obtain a physically consistent set of SSTs and not because we believe that the Younger Dryas cooling was caused by a shut-down of the conveyor belt. In short, the difference between the experimental design of YD_A and YD_B is an additional cooling of 2 to 4 °C in the N Atlantic during winter and the resulting southward displacement of the sea-ice margin from an average position at 65 °N to 55 °N. A summary of the boundary conditions in experiment YD_B is given in Fig. 1.

The total simulation time per experiment was set to 12 model years of 360 days each. The first two years were taken as spin-up time and the last ten years were used in the analyses. We averaged the results of these ten years to obtain a yearly average climate. Additionally, we used daily results to calculate an index to measure the occurrence of blocking and strong zonal flow as defined by Liu (1994) and high frequency variability (storm tracks) with a time-filter defined by Blackmon (1976).

3 Results and discussion

Paleoclimate reconstructions indicate that the mean annual temperature in NW Europe was below 0 °C (permafrost conditions) during the coldest spike of the Younger Dryas (e.g., Bohncke 1993). In NW Europe summer temperatures appear to have been around

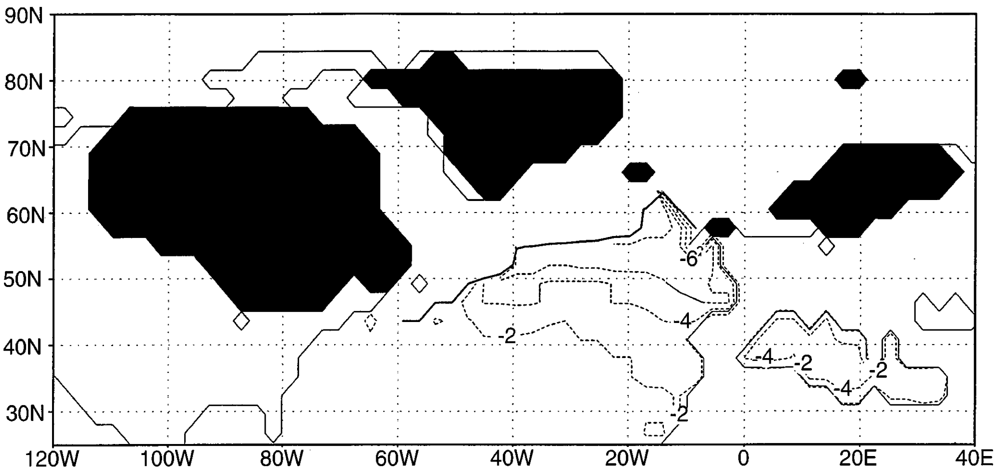

Fig. 1. The most important surface boundary conditions in the North Atlantic/European sector in experiment YD_B plotted on the used model grid. The area covered with ice sheets is *hatched*. The introduced January SST anomalies in the North Atlantic Ocean are given by the *broken lines* with a 2 °C interval. The resulting sea-ice margin is represented by the *thick continuous line*

10 °C, whereas the winters must have been very cold with temperatures below −10 °C. These values imply a summer cooling of −6 °C and a winter cooling of −15 °C compared with present conditions. In experiment YD_A the simulated differences with CTRL were about 0 °C in summer and −5 °C in winter (Renssen et al. 1995). This means that the temperatures in experiment YD_A are closer to BA interstadial conditions than to those during the Younger Dryas. Analyses of the results of experiment YD_A (Renssen et al. 1995) showed that the SST-set and the ice sheets were the main factors controlling the cooling in the North Atlantic/European sector and that the changed insolation and CO_2 concentration contributed little to this.

The results of experiment YD_B were much closer to the conditions suggested by geological evidence, as winter temperatures were around 12 °C lower in NW Europe compared to CTRL, i.e., about 7 °C cooler than experiment YD_A (not shown). Since the value of −12 °C is reasonably close to the reconstructed cooling (−15 °C), some confidence can be placed in the boundary conditions prescribed in YD_B. However, summer temperatures in Europe are still too high in YD_B, as these figures are comparable with YD_A. Yet, we believe that the calculated winter climate in experiment YD_B is a reasonable approximation of the conditions during the Younger Dryas, and we therefore assume that the simulated winter atmospheric circulation is a realistic description of the YD situation.

In the following sections we discuss the changes of the atmospheric winter circulation at the surface and at the 500 and 200 hPa levels. Emphasis will be on the causes of these changes and their climatic implications. In these sections the presented data are averages of 10 winters (December, January, February, DJF), unless otherwise stated.

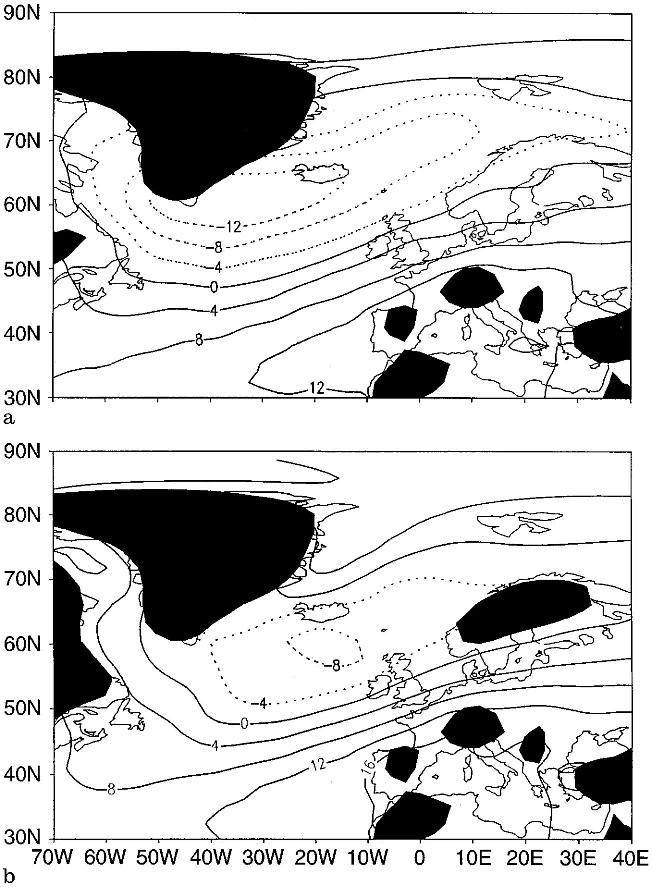

Fig. 2a–b. DJF mean sea level pressures (hPa) for experiments CTRL (*top*) and YD_B (*bottom*), given as deviations from the global mean to account for the introduction of ice sheets in YD_B. Contour interval is 4 hPa. Elevated areas have been *blacked out*

3.1 Surface circulation

A good indication of the atmospheric circulation at the earth surface is provided by the mean sea level pressure. In Fig. 2a, b the winter mean sea level pressures (MSLP) are plotted for experiments CTRL and YD_B. Although the overall pattern looks quite similar, indicating a westerly circulation in the North Atlantic region, a few important differences are evident. First, the Icelandic low is displaced in experiment YD_B from a position between Greenland and Iceland to a location south of Iceland, so by 15° of longitude to the east. Second, the minimum pressure in the Icelandic low is substantially higher (6 hPa) in YD_B. Nevertheless, the pressure gradient over the eastern N Atlantic increased considerably, which is caused by the eastward shift of the Icelandic low in conjunction with a small increase of pressure in the subtropics.

These differences are probably mainly caused by the southward expansion of sea-ice. In the model only minor heat transport from the ocean to the atmosphere is possible after sea-ice formation. Together with the high albedo of the sea-ice, this results in an extreme cooling at the surface. In experiment CTRL the temperature is 5 °C over open sea, whereas in YD_B the DJF surface temperature in the region of Iceland is as low as −32 °C (not shown). Consequently, in this region a cooling of 37 °C is apparent in YD_B. This severe cooling of the lower air layer in YD_B causes contraction and consequently a relatively high MSLP. However, ice sheets affect the surface circulation in the opposite way. The comparison of experiments CATL and YD_A in Renssen et al. (1995) made clear that the introduction of ice sheets in experiment YD_A caused an intensification of the Icelandic low, since the pressure was 5 hPa lower in this experiment than in CTRL and CATL. Evidently, in experiment YD_B the extended sea-ice has a strong stabilizing effect on the lower atmosphere, which is more powerful than the opposite effect of ice sheets.

Important changes in surface winds may be deduced from the variations in sea level pressure. A strengthened northerly flow is expected between Greenland and Iceland as a result of the eastward displacement of the Icelandic low. Moreover, the increased pressure gradient over the eastern N Atlantic results in stronger surface westerlies over NW Europe. The net result of these anomalous winds is transport of very cold arctic air masses (cooled over sea-ice and ice sheets) with westerlies to Europe. In experiment YD_A, with a less extensive sea-ice cover, the winter temperature is about 7 °C higher in Northern Europe and the conditions are therefore closer to those of the BA interstadial complex than of the Younger Dryas. Therefore, the formation of sea-ice in the North Atlantic Ocean turns out to be an efficient mechanism to cool Europe to YD temperatures.

Geological data give conflicting evidence on the extension of sea-ice in the North Atlantic Ocean. Analyses on the dust record from the GISP2 Greenland ice-core support the inference that changes in sea-ice cover were important during the Late Glacial. Mayewski et al. (1994) suggested that the sea-salt input into the ice-core dust record was limited by the extended Atlantic sea-ice cover during the YD stadial compared to BA conditions. However, according to Ruddiman and McIntyre (1981) no significant increase in sea-ice occurred in the North Atlantic during the YD stadial. This conclusion was based on a reconstruction of the foraminiferal and coccolith productivity in the N Atlantic, which was comparable to the productivity during the BA interstadial. In conclusion, the extension of YD sea-ice in the North Atlantic Ocean is still uncertain and needs to be studied further.

Our finding that the atmospheric surface circulation in the Atlantic region is very sensitive to sea-ice cover agrees with Kutzbach and Ruddiman (1993). They used the climate community model (CCM) to study the sensitivity of the glacial climate to the extent of North Atlantic sea-ice. Two January LGM simulations were performed which only differed in winter sea-ice cover. In the standard LGM simulation the sea-ice limit was specified at 45 °N, whereas in the sensitivity experiment this limit was located at 60 °N. Kutzbach and Ruddiman (1993) found a 30 °C lower surface temperature over the sea-ice covered area in the extended-ice experiment. Furthermore, in the latter simulation the minimum pressure of the Icelandic low was 10 hPa higher and the pressure system was moved to the east. Moreover, the core of surface westerlies was displaced 15° of latitude to the south in the case with extended sea-ice. The strong westerlies over NW Europe in the reduced-ice case caused advection of relatively warm air. Consequently, the January temperatures in this region were 5–10 °C higher in the reduced-ice experiment. The differences between the reduced-ice and extended-ice simulations found by Kutzbach and Ruddiman (1993) are comparable to the differences between our YD_A and YD_B experiments. Therefore, it confirms our conclusion that sea-ice cover in the North Atlantic is the main factor controlling the changes in surface flow.

The inferred westerly winds over Europe are different from those found in most glacial experiments. In many LGM simulations with CLIMAP ice sheets a strong glacial anticyclone is present over the Scandinavian ice sheet, resulting in easterly-northeasterly surface winds in Europe (e.g., Kutzbach and Wright 1985; Rind 1987). Such a surface circulation is also present in the Younger Dryas simulation of Rind et al. (1986). However, the response to LGM ice sheets may be model dependant. In LGM experiments with the AGCM of the Laboratoire Météorologie Dynamique no clear anticyclone is apparent over Europe (Joussaume 1993). Moreover, recent studies suggest that the ice sheets were lower than the CLIMAP reconstructions used in the mentioned LGM experiments (Peltier 1994). In our YD_B experiment there is little evidence for a glacial anticyclone over Northern Europe. Probably the Scandinavian ice sheet was too small for the development of a glacial anticyclone. Nevertheless, the strong westerly circulation in our results is in excellent agreement with the dunes of Younger Dryas age, which indicate westerly winds during deposition (Böse 1991). We tentatively conclude therefore that if these dunes can give an indication of mean surface flow during winter, our simulation of YD surface flow is realistic.

3.2 Circulation at 500 hPa

In this section we discuss both the mean state and the high frequency variability at 500hPa.

Mean state. In Fig. 3a, b the mean hemispheric DJF fields of the 500 hPa geopotential heights are plotted. From a comparison of these figures it is clear that the trough over the North Atlantic Ocean has deepened in YD_B and that this has enlarged the height gradient considerably in this region. Hence, a strengthened westerly circulation is present in the YD_B case. To study the effect of different forcings, we plotted cross sections through the 500 hPa geopotential height field at 60 °N of all four experiments mentioned in Table 1 (see Fig. 4). This has the advantage that the location of ridges and troughs is readily identified. By comparing the experiments CTRL and CATL, it can be seen that the main effect of a cold North Atlantic Ocean is a reduced ridge over Europe. The major differences between CATL and YD_A are a deepening of the Atlantic trough and a relocation to the East of the minimum height of this trough. Since the same set of Atlantic SSTs is used in CATL and YD_A, these variations can be attributed to the introduction of the Laurentide ice sheet in North America. Comparing the 500 hPa heights of experiments YD_A and YD_B, we see a further reduction of the European ridge in YD_B. This lower ridge was caused by the lowering of the Atlantic SSTs and the extended sea-ice cover, because this is the only difference between the design of the experiments YD_A and YD_B. A similar effect is seen in the result by Rind (1987), who carried out a LGM sensitivity experiment with CLIMAP SSTs. In summary, lowering of the N Atlantic sea surface temperatures results in a reduction in height of the European ridge, whereas the presence of the Laurentide ice sheet causes a

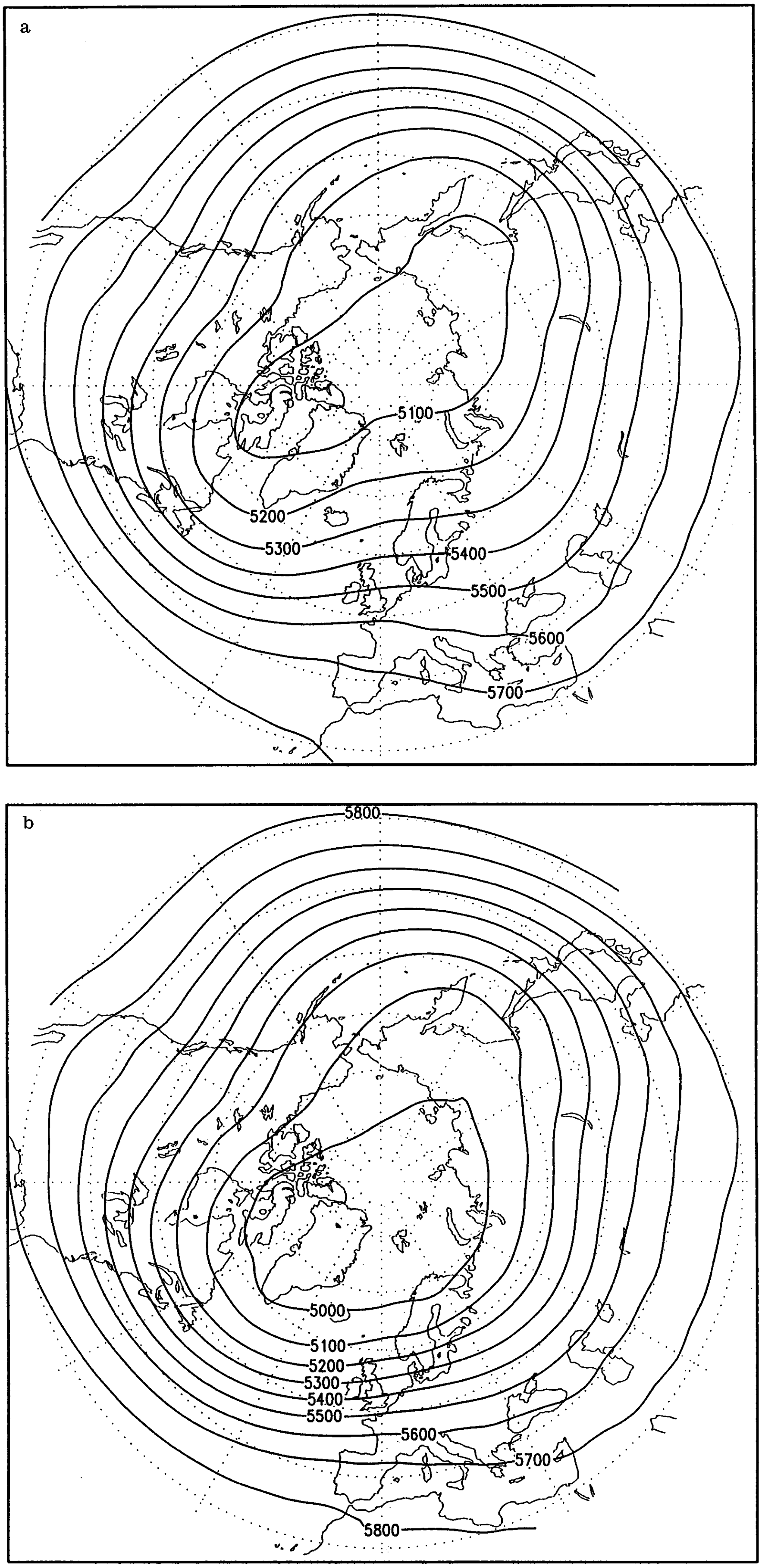

Fig. 3a, b. Hemispheric projection of the 500 hPa DJF geopotential height fields (m) for experiments CTRL (*top*) and YD_B (*bottom*) with an interval of 100 m

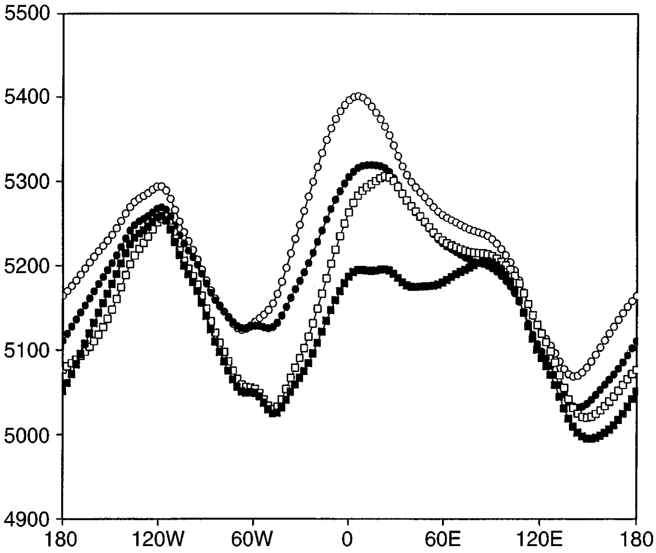

Fig. 4. Cross section of DJF 500 hPa geopotential heights (m) at 60°N for the experiments CTRL (*open circles*), CATL (*solid circles*), YD_A (*open boxes*) and YD_B (*solid boxes*)

deepening of the west Atlantic trough and a displacement to the east. The noted intensifying of the west Atlantic trough under influence of the Laurentide ice sheet is also found in LGM simulations (e.g., Broccoli and Manabe 1987; Rind 1987). In these full glacial experiments a ridge is present over western North America under influence of a southerly flow, induced by the glacial anticyclone over the Laurentide Ice sheet. The enhancement of this ridge is not seen in our results, presumably because the Late Glacial Laurentide ice sheet was too small.

The occurrence of blocking and strong zonal flow. Although it is obvious that a strengthened westerly circulation exists in experiment YD_B, it remains uncertain from the mean state presented in Figs. 3 and 4 what the day-to-day variability is. Was the mean state of Fig. 3b stable? To analyze this, we calculated an index based on daily winter 500 hPa geopotential height fields, as proposed by Liu (1994). With this index (named *BINX*) one can identify both the occurrences of strong zonal flow regimes and situations with blocking. In accordance to Liu (1994) 31 grid-cells were selected at 60°N covering the North Atlantic/European region between 55°W and 30°E. Subsequently, we calculated the average difference with the climatological mean of Fig. 3a for each day of experiment CTRL. As a next step the 200 days with the largest positive anomalies were selected and the matching anomaly fields were averaged. We defined this average as the 'blocking anomaly field' (Fig. 5). Figure 5 shows that this blocking anomaly field is a large positive anomaly of the 500 hPa geopotential height field of the control climate, centred around the point 60°N, 15°W. Liu (1994) has shown that both blocking and strong zonal flow (SZF) regimes over the North Atlantic and Western Europe are associated with approximately the same anomaly pattern but with opposite sign. Therefore, by projecting each of the 900 daily 500 hPa geopotential height fields onto the blocking anomaly field of Fig. 5, we could measure the resemblance of a particular circulation pattern with the blocking regime or the strong zonal flow regime. As a projection area the Atlantic/European area of 30°N to 85°N and from 90°W to 60°E was chosen. The blocking index *BINX* is defined by Liu (1994) as

$$BINX = \langle z_b, z_d \rangle / \langle z_b, z_b \rangle$$

where z_b is the blocking anomaly field in geopotential height and z_d the daily geopotential anomaly field. The angle brackets denote a squared norm inner product. A blocking day is defined as a day with $BINX \geq 0.5$ and a blocking event is defined as a period of consecutive days for which $BINX \geq 0.5$. A criterion of $BINX \leq -0.5$ is used to define a day with strong zonal flow.

We calculated *BINX* for the experiments CTRL and YD_B, both with the blocking anomaly pattern from Fig. 5. Liu (1994) has calculated *BINX* for 10 winters (1982/83 to 1991/92) of ECMWF analyses and found 233 blocking days and 242 days with strong zonal flow (see Table 2). In the observations the average duration of both blocking and SZF was 9.3 days. The number of blocking and SZF days in CTRL is in agreement with the observations. However, the model underestimates the duration of both types of circulation, as the average blocking event in the control experiment lasts 5.9 days and the average duration of an SZF event is 5.3 days. It has to be noted, however, that the observations and results of CTRL are not fully comparable. First, the ECMWF analyses data have a higher spatial resolution than the used model. Second, interannual SST variability may have influenced the observations considerably. In experiment YD_B the situation is dramatically different compared to control climate. From Table 2 it is evident that blocking, as it took place in CTRL, only occurred on 7 of the 900 days. On 84.7% of the calculated days (namely 763 out of 900) there was a circulation with strong zonal flow. The average duration of SZF events in experiment YD_B was 25.4 days. So, from these calculations we can conclude that the atmospheric circulation in experiment YD_B is characterized by a stable strong zonal flow over the North Atlantic/European sector.

In the present climate the occurrence of blocking during winter is associated with colder than normal winters in Europe (Moses et al. 1987). Consequently, Lamb and Woodroffe (1970) have suggested that a circulation type with a high frequency of blocking anticyclones and the accompanying easterly flow over NW Europe was present during the Younger Dryas. However, our results show that a Younger Dryas winter cooling in Europe could be present with a strong westerly flow regime without the occurrence of significant blocking. As noted earlier, of prime importance for the YD cooling in Europe seems to be the existence of extensive sea-ice in the North Atlantic. In our experiment YD_B a large reservoir of arctic air builds up

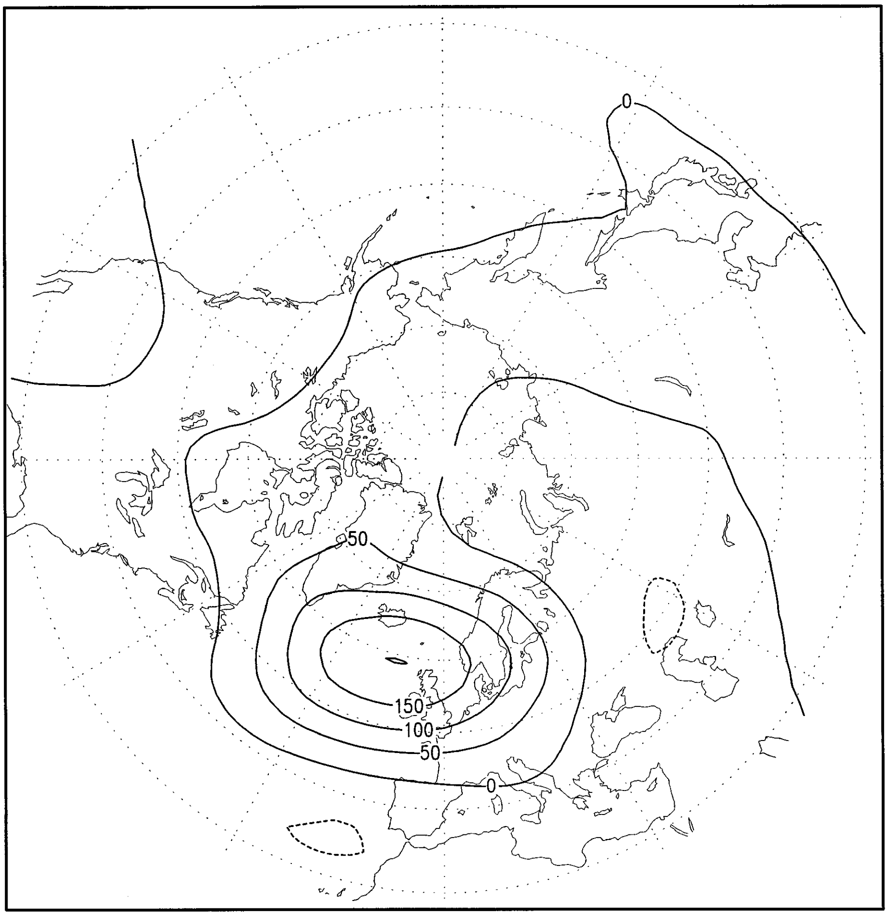

Fig. 5. 'Blocking anomaly' of the 500 hPa geopotential height field in experiment CTRL. Difference between the mean of 200 days (out of 900) with the largest positive anomaly at 60 °N and the overall DJF mean. Intervals at 50 m

Table 2. Blocking and strong zonal flow in 10 winters in observations (ECMWF analyses of 1982/83–1991/92, Liu, 1994) and experiments CTRL and YD_B. The total number of days used in each analysis is 900 days

	Observations	CTRL	YD_B
Number of blocking days	233	234	7
Number of blocking events	25	40	3
Average duration events	9.3	5.9	2.3
Number of SZF days	242	233	763
Number of SZF events	26	44	30
Average duration events	9.3	5.3	25.4

over this sea-ice, which is transported by westerly storms into Europe.

Storm tracks. Situations with strong zonal flow are associated with the frequent occurrence of travelling cyclones in Europe. It is interesting to see how the activity of travelling cyclones changed in experiment YD_B when compared with CTRL. To analyze this, the high-frequency transient variance of the DJF 500 hPa geopotential height was calculated, using the time-filter of 2.5 to 6 days as introduced by Blackmon (1976) (Fig. 6a, b). Areas with high values of this high frequency variance are usually associated with storm tracks. Figure 6a is taken from Roeckner et al. (1992), who calculated the storm tracks with the above method for the same control experiment as elsewhere described in this study, but for another 10 year sample. In Fig. 6a the storm track over the North Atlantic starts at 50 °N with a maximum variability in the 500 hPa geopotential height field of 60 m near the North American east coast. Subsequently, the track follows a southwesterly route towards Iceland with values around 40 m. In experiment YD_B the storm track begins at the same location, but with a higher variability of 70 m. Moreover, the axis of high transient variability in YD_B lies in a more west-east direction. Furthermore, the zone of high cyclonic activity extends far over the Eurasian continent. The maximum value seen over the western North Atlantic in the control climate (60 m) can be found in YD_B as far east as 60 °E in western Siberia. From these analyses we conclude that the activity of travelling cyclones is very different in experiment YD_B compared with CTRL. First, the cyclonic activity is enhanced substantially and second, the orientation of the storm tracks is different, causing the cyclones to travel far inland.

The enhanced cyclonic activity was presumably caused by the increased baroclinicity due to strong thermal gradients at the surface and its concomitant strong increase of wind speeds with altitude. In experiment YD_B the North Atlantic Ocean was considera-

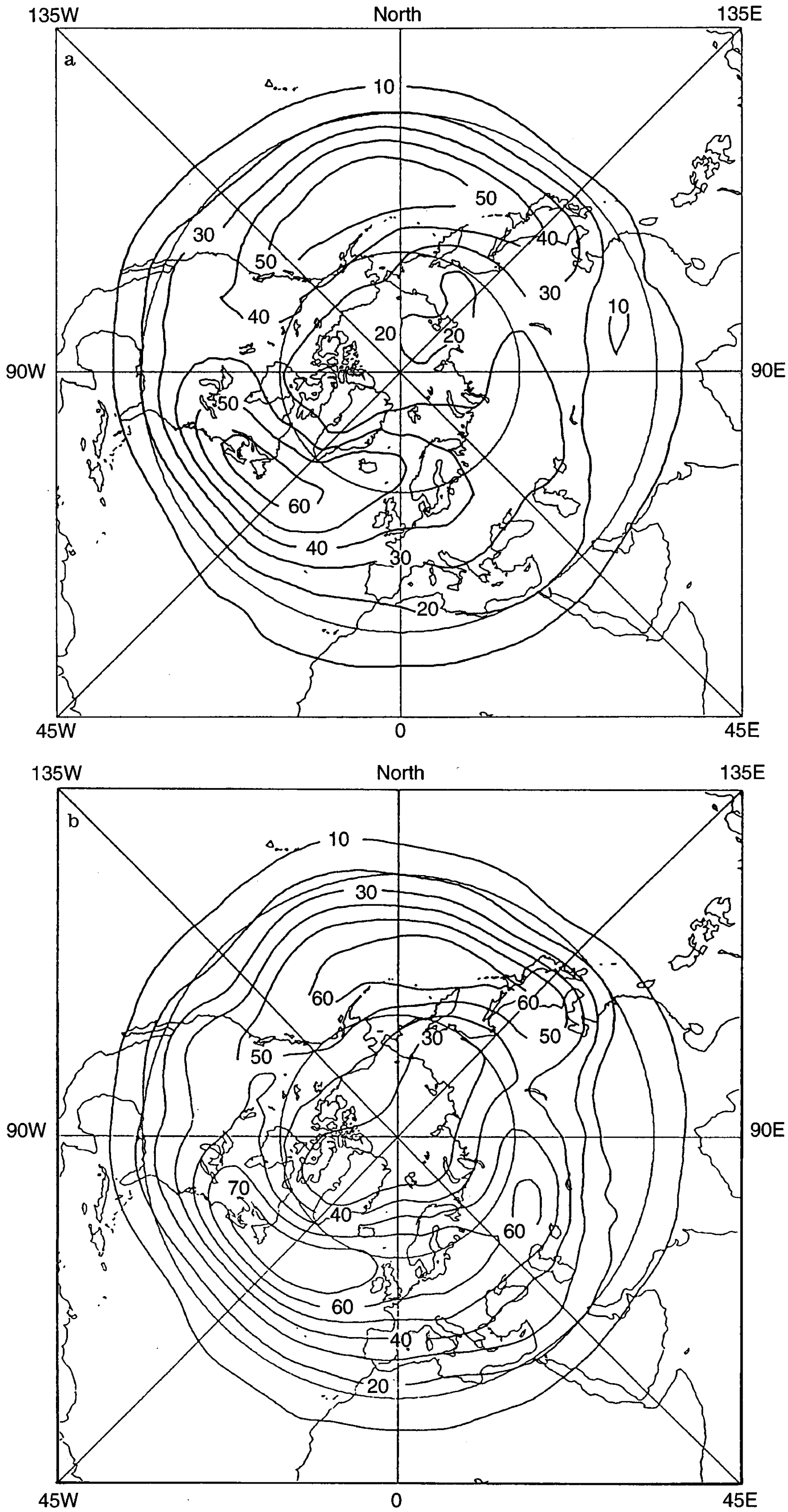

Fig. 6a, b. Climatological mean of the bandpass-filtered root mean square DJF 500 hPa geopotential heights in metres (storm tracks) for experiments CTRL (*top*, Roeckner et al. 1992) and experiment YD_B (*bottom*) bly cooled, whereas the tropical Atlantic was left unchanged. As a consequence, the surface temperature gradient increased substantially. Moreover, the introduction of ice sheets in North America and Europe caused an enlarged temperature gradient over land (Renssen et al. 1995). These suggestions are supported by Valdes and Hall (1994), who have found in a perpetual February LGM simulation with the ECMWF model that mid-latitude depressions closely follow the edge of Atlantic sea-ice due to the presence of a strong temperature gradient. Moreover, they noted a considerable increase in eddy kinetic energy over the European continent as far east as 90°E.

3.3 Circulation at 200 hPa

The winds at 200 hPa are plotted for all four experiments in Fig. 7a–d. In experiment CTRL (Fig. 7a) an area with high wind speeds over 40 m/s is located over the American east coast between 100°W and 45°W with a centre at 70°W. In Fig. 7b (experiment CATL) a strengthening of the jet stream is seen with wind speeds exceeding 50 m/s. This proves that the enlarged temperature gradient influences the whole troposphere. However, the maximum is at the same location. Comparison of YD_A (see Fig. 7c) with CATL reveals that the effect of the Laurentide ice sheet is a strengthening and relocation of the jet-core to the east with 10° of longitude, as the maximum wind speeds (>50 m/s) are now centred at 60°W. The eastward shift of the jet-core agrees with the discussed deepening of the trough in the 500 hPa geopotential height field over the western N Atlantic. A further cooling of the surface in experiment YD_B (Fig. 7d) causes an additional intensifying of the jet stream, with the area of high wind speeds (>40 m/s) reaching the European west coast. In summary, a strengthened jet stream is present over the zone with the largest surface temperature gradient and with enhanced baroclinicity.

The effect of ice sheets on the jet stream was studied by Shinn and Barron (1989), who performed two LGM perpetual January experiments with the CCM AGCM, one with small ice sheets and a second with huge ice sheets. They distinguished a thermal forcing (albedo and IR opacity) and an orographic effect (ice sheet as a physical barrier). Shinn and Barron (1989) found that an increase in height of the Laurentide ice

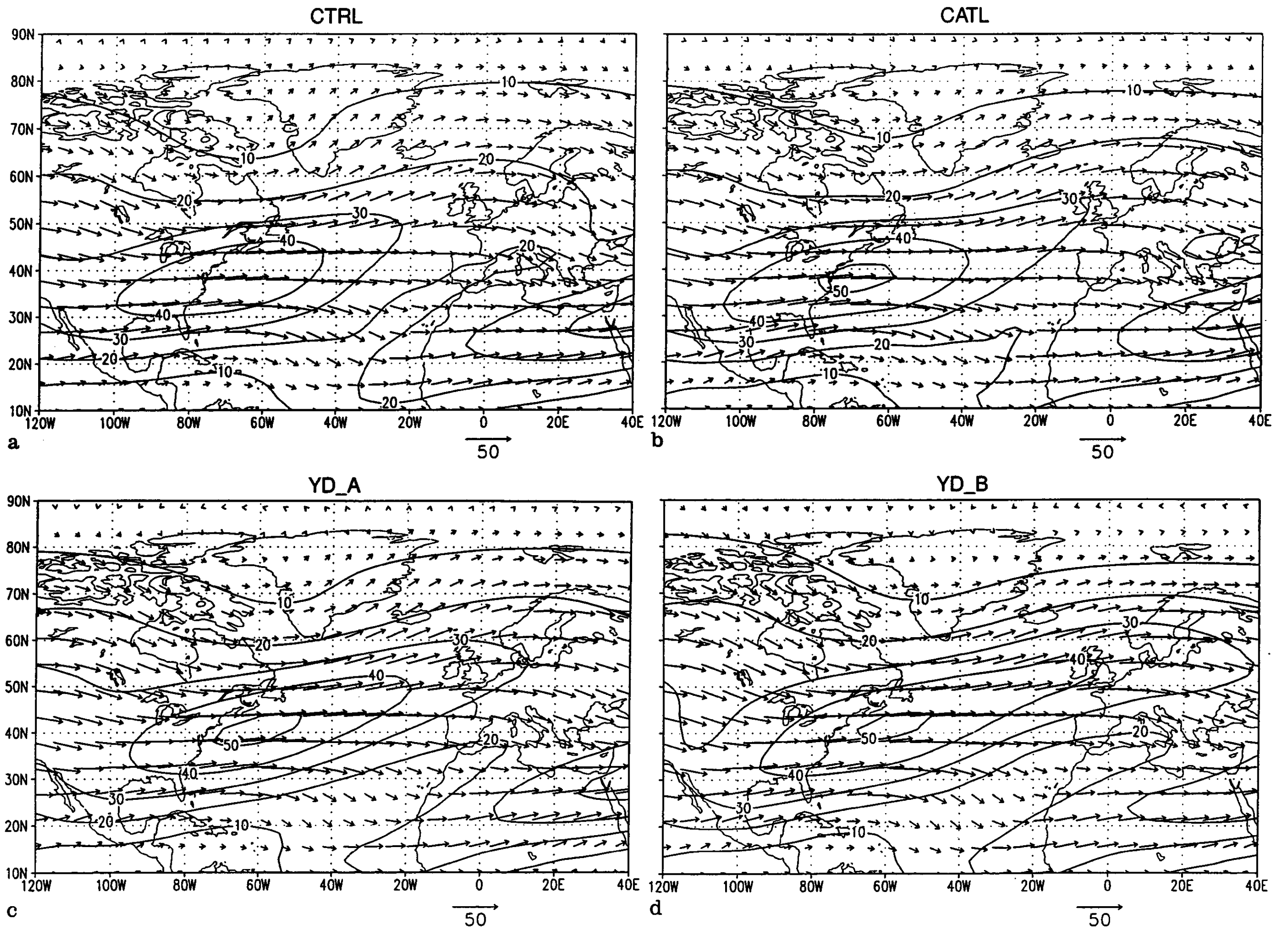

Fig. 7a–d. DJF winds at 200 hPa for the experiments *CTRL, CATL, YD_A* and *YD_B*. Note the *arrow* for scale. Wind speeds are given at 10 m/s intervals

sheet intensified the jet stream over the western Atlantic. Another orographic effect was splitting of the jet. The thermal effect, on the other hand, had more control on the latitudinal position of the jet. The ice sheets increased the land-sea thermal contrast, thereby reducing the areal extent of oceanic polar lows. The result was a poleward shift of the jet over the North Atlantic in the case with the larger ice sheets. The latter effect acted in opposition to the thermal effects of sea-ice and the oceanic polar fronts, since the jet transported relatively warm air northwards. Although in our results the introduction of the Laurentide ice sheet causes indeed a small relocation of the jet-core, no poleward shift is present. Probably the Late Glacial ice sheets have a too small areal extent to cause such an effect.

The orographic effect noted by Shinn and Barron (1989) is partly seen as a strengthening of the jet stream over the Atlantic Ocean in YD_A compared to CATL. However, no split of the jet is visible in our experiments, as might be expected with the much smaller ice sheets. This is consistent with the absence of the enhanced ridge in the 500 hPa geopotential height field over western N America as discussed in Sect. 3.2. Besides a stronger jet as a result of the Laurentide ice sheet, a further intensifying as a result of surface cooling in the N-Atlantic region is apparent (compare CTRL with CATL and YD_A with YD_B). In experiment YD_B the jet is especially stronger over Europe. This is in agreement with the discussed increase in cyclonic activity in this region, as storm track changes can be expected to follow changes in jet stream patterns (Kutzbach and Wright 1985; Shinn and Barron 1989). A similar effect can be seen in the Younger Dryas simulation of Rind et al. (1986), where the colder Atlantic Ocean shifted the upper air trough eastward from its position over eastern North America in the Allerød simulation. The result was a strengthening of the jet stream across the Atlantic Ocean.

4 Conclusions

1. According to our AGCM simulations, the low winter temperatures during YD time in Europe were present with westerly winds and not with frequent easterly winds as suggested by Lamb and Woodroffe (1970). Furthermore, the results show that blockings, which are necessary for easterly and northeasterly winds in Europe, are almost absent. This conclusion agrees well with reconstructed wind directions based on dunes of YD age.
2. The successive experiments make clear that the extended sea-ice cover in the North Atlantic Ocean is the crucial factor forcing the specific YD circulation and associated low temperatures.
3. The simulations show that the YD winter circulation is characterized by a strengthened jet stream over the North Atlantic, which caused a considerable increase in cyclonic activity over an area spanning from North America to far into Eurasia.
4. Our results reveal a YD winter circulation that is different from the LGM circulation found in most simulations using CLIMAP ice sheets. No clear glacial anticyclones are present and no split of the jet stream occurs as a result of the Laurentide ice sheet. It is likely that this difference is caused by the reduced size of the prescribed ice sheets.
5. The decisive role of sea-ice cover in the North Atlantic Ocean revealed by our experiments needs to be confirmed by independent studies.

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