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# Surface temperature in NW Europe during the Younger Dryas: AGCM simulation compared with temperature reconstructions

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**Abstract** During the Younger Dryas (YD) the climate in NW Europe returned to near-glacial conditions. To improve our understanding of climate variability during this cold interval, we compare an AGCM simulation of this climate, performed with the ECHAM model, with temperature reconstructions for NW Europe based on geological and paleoecological records. Maps for the mean winter, summer and annual temperature are presented. The simulated winters are consistent with reconstructions in the northern part of the study area. A strong deviation is noted in Ireland and England, where the simulation is too warm by at least 10 °C. It appears that the N Atlantic was cooler than prescribed in the YD simulation, including a southward expansion of the sea-ice margin. The comparison for the summer shows a too warm continental Europe in the simulation. Supposedly, these anomalously warm conditions are caused by the AGCM's response to the prescribed increased summer insolation. The region of maximum summer cooling is similar in both the simulation and reconstruction, i.e., S Sweden. We suggest that this is due to the local cooling effect of the Scandinavian ice sheet. Compared to the present climate a considerable increase of the annual temperature range is inferred, especially for regions close to the Atlantic Ocean.

## 1 Introduction

The Younger Dryas (YD, 11.0–10.0 $^{14}$C ky BP, Mangerud et al. 1974, or ~12.5–11.5 calendar ky BP) represents a marked cooling during the last glacial/interglacial transition in Europe (e.g., Walker 1995). New information concerning climate variability in this interval may be obtained by means of simulations with atmospheric general circulation models (AGCMs). The validity of these simulations may be judged by comparing the AGCM results with independent climate reconstructions based on geological and paleoecological, (so-called, proxy) data (COHMAP members 1988; Crowley and North 1991).

We compare new simulation results on the YD climate (Renssen et al. 1996) with detailed temperature reconstructions for NW Europe (Isarin 1997a, b). The aims of this comparison are 1) to evaluate the AGCM experiments (including the boundary conditions) and the climate reconstructions, and 2) to improve our understanding of the nature of the YD climate conditions in Europe.

AGCM simulations of paleoclimates have been mainly performed for the climate of the last glacial maximum (LGM, 18 $^{14}$C ky BP or ~21 cal ky BP). In these studies the set of boundary conditions compiled by the CLIMAP project members (1981) was used (e.g., Kutzbach and Wright 1985; Rind 1987; Lautenschlager and Herterich 1990; Joussaume 1993). Such a detailed set of boundary conditions was not available for the YD, thus hampering efforts to simulate the YD climate. AGCM studies focusing on the YD therefore used crude estimates of the surface conditions at that time (Rind et al. 1986; Overpeck et al. 1989). Moreover, the latter simulations were all performed with a low resolution AGCM. Recently, however Renssen et al. (1995, 1996) have carried out simulations on the YD climate using a second-generation AGCM and an updated set of boundary conditions.

Although reconstructions of the YD climate have been published for sites around the globe, most studies focus on the North Atlantic region (e.g., Lowe and NASP members 1995). A wealth of data is available from Europe, where the YD cooling is clearly registered

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in geological records (e.g., Walker 1995) and temperature reconstruction studies date back to the 1950s (e.g., Iversen 1954). Recently, Isarin (1997a,b) made a comprehensive and detailed analysis of high quality data providing information on the YD climate in NW Europe, thereby using different types of well-dated proxy records.

Comparison of simulations with climate reconstructions provides understanding of the relevance of the AGCM results. For instance, Kutzbach and Wright (1985) comprehensively compared LGM simulation results with data from locations in North America. Similarly, the COHMAP group performed a global comparison for the 18, 15, 12, 9, 6 and 3 ky BP time slices (COHMAP members 1988; Wright et al. 1993). Rind et al. (1986) compared their experimental results on the YD climate with temperature reconstructions on numerous sites. However, a detailed comparison of YD simulation results with a coherent spatial overview of reconstructed temperatures, in the form of isothermal maps, has not been carried out.

## 2 Methods

Surface air temperatures are presented as winter (December–January–February), summer (June–July–August) and annual means. We converted all values to sea level to facilitate the model-reconstruction comparison, using a lapse rate of 6 °C/km. The reconstructed temperature fields are compared with observed temperature averages for the period 1961–1990 (Climate Research Unit 1992). Similarly, we compared the AGCM results on the YD climate with a simulation of the present climate. The simulation results are derived from 10 calculated annual cycles. We focus on NW Europe, as visualized in Fig. 1.

### 2.1 AGCM experiments

Simulation experiments on the YD climate were performed with the ECHAM3 (European Centre/HAMburg) AGCM. This model is described in detail in DKRZ report number 6 (DKRZ Modellbetreuungsgruppe 1993). We used the T42 version with a spatial resolution of ~2.8 degrees of latitude-longitude. With this model an experiment of present climate was performed (control experiment, hereafter CTRL, Roeckner et al. 1992). The model satisfactorily simulates most aspects of modern climate. However, some regional temperature errors have been recognized that may be related to the inadequate representation of surface inhomogeneities such as orography, surface roughness and surface albedo (Roeckner et al. 1992). For instance, these authors showed that in a simulation of modern climate an anomalous southerly air flow caused the simulated winter temperatures to be too high over Europe (max. 2 °C at 850 hPa). Similarly, they demonstrated that during summer the ECHAM3/T42 model simulates somewhat too high temperatures over Europe when compared with observations.

Initially, we performed three experiments with different versions of sets of YD boundary conditions. In a first simulation we only cooled the N Atlantic Ocean according to sea surface temperature (SST) reconstructions based on foraminifera as published by Sarnthein et al. (1995). The sea-ice cover was prescribed according to Koç et al. (1993). In a second experiment we introduced the Laurentide and Scandinavian ice sheets of 12 calendar ky BP in agreement with Peltier (1994) and lowered the atmospheric $CO_2$ concentration to

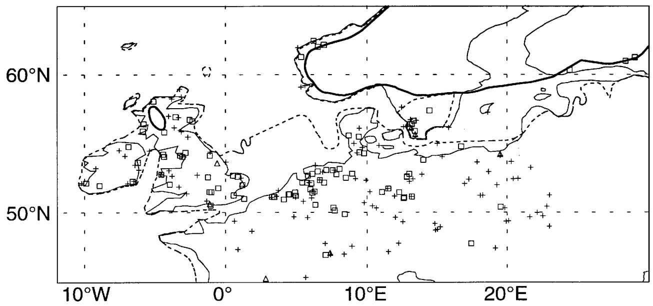



**Fig. 1** Study area with sites used in the temperature reconstructions: botanical sites (+), periglacial sites (△), beetle sites (□). See Isarin (1997a,b) for references. Younger Dryas coast line (*dotted line*) and ice sheet margins (*bold line*) after Gerasimov and Velichko (1982) and Rainio et al. (1995). Some symbols indicate more than one site

230 ppmv as inferred from Jouzel et al. (1992). Also, we changed the insolation to 12 cal ky BP conditions according to Berger (1978), resulting in a pronounced increase during summer and a decrease during winter. In addition, we added land points in the North Sea and Bering Strait areas in agreement with sea-level curves (e.g. Fairbanks 1989). The design of the third simulation was identical to that of the second, except for an additional cooling of 2 °C in the N Pacific, which is in agreement with Kallel et al. (1988). To find the set of boundary conditions with the best resemblance to the YD situation, these experiments were evaluated by crudely comparing the results with climate reconstructions (Renssen et al. 1995).

This comparison revealed that the simulations produced temperatures in Europe that were considerably higher than suggested by geological data. The difference was especially clear for the winter season, as a cooling of only a few degrees was simulated in NW Europe, whereas a depression of at least 15 °C was estimated by 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 the first experiment (only SSTs changed) and the second and third (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). As discussed in Renssen (1997), it is important to realize 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). 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, one could argue 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 the three initial experiments (Renssen et al. 1995). Therefore, we performed a fourth experiment (hereafter YDSIM), identical in design to the third simulation, except for the set of winter SSTs. We used in YDSIM the SSTs from the output of an experiment with a coupled ocean-atmosphere model (Hamburg LSG ocean model coupled to T21 ECHAM3) in which the thermohaline circulation was halted by introduction of a large amount of fresh water in the North Atlantic Ocean (Schiller et al. 1996). It is important 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 winter SSTs were 2 to 4 °C lower than the set of Sarnthein et al. (1995). We assumed that the tropical Atlantic Ocean was not significantly cooler than today. This assumption agrees with SST reconstructions presented by Schulz (1995). In short, the N Atlantic SSTs in YDSIM were maximally 8 °C lower than in CTRL during winter and 10 °C during summer (see Renssen 1997). Sea-ice was defined when SSTs reached values below $-1.8\,^{\circ}$C. As a result of the cooling the sea-ice margin was shifted southwards, with a mean position in YDSIM at 56 °N in winter. During summer the sea-ice margin was prescribed in the Norwegian Sea at 70 °N following to Koç et al. (1993).

We realize that it would be convenient to compare this fourth YD simulation directly with the reconstructed temperatures. However, since the simulation of present climate differs from the observed temperatures, it is more accurate to compare on the one hand the difference between a control experiment and the simulation of YD climate, with on the other hand the difference between observations of modern climate and reconstructed YD temperatures.

## 2.2 Temperature reconstructions

Temperature conditions during a specific interval may be reconstructed from various sources of proxy climate evidence (see Table 1). By combining the individual lines of evidence, i.e., a multi-proxy approach, a wide range of climate parameters is reconstructed. Furthermore, the conjunction of individual, independently derived estimates enhances the credibility of the reconstructed temperatures and narrows the temperature ranges. However, differences exist in the quality of the records and thus in their contribution to the climate reconstructions. This quality is reflected in the way data are documented and, most importantly, in the geochronological control of the records (Huijzer and Isarin 1997). We reconstructed YD temperatures using periglacial features (e.g., ice-wedge casts) and paleoecological records, including pollen and plant macro fossils. Entomologic (beetles) and glaciological studies, the latter describing changes of equilibrium line altitudes (ELAs) of glaciers, were used in conjunction. We focused on the first part of the YD for it is assumed that in this part (approximately between 10.9 and 10.5 $^{14}$C ky BP) the phase of maximum cold occurred (compare Bohncke et al. 1993; Berglund et al. 1994). The phase of maximum cold is followed by a phase of warming, ultimately resulting in the start of the Holocene at $\sim$10.0 $^{14}$C ky BP.

Climate records were stored in a relational 'Multi-proxy database' (Huijzer and Isarin 1997). This relational database contains tables that are related to the biotic and abiotic evidence, geochronological control, site information and inferred climate parameter values. In the climate conversion tables proxy data are translated into climate parameter values. A direct link with a geographical information system facilitates data management and spatial analysis.

### *2.2.1 The reconstruction of mean annual temperatures*

Estimates of YD mean annual temperatures are based on periglacial records. A comparison of specific relict phenomena and their counterparts from the modern periglacial domain enables an estimation of maximum mean annual temperatures during a specific cold interval to be made (see, e.g., Karte 1983; Huijzer 1993; Vandenberghe and Pissart 1993; Ballantyne and Harris 1994). Whereas some features originate exclusively in the continuous permafrost zone (e.g., ice-wedge polygons and closed-system pingos), others are restricted to the zone of discontinuous permafrost (e.g., open-system pingos). The reconstructed $-8\,^{\circ}$C isotherm is based on ice-wedge polygons in coarse sediments and represents the southern limit of the continuous permafrost zone (compare Péwé 1966; Romanovskij 1985). The distribution of open-system pingos and casts of immature ice-wedges constitutes the $-4\,^{\circ}$C isotherm. The $-1\,^{\circ}$C isotherm is based on small-scale cryoturbations and seasonal frost cracks, and consequently represents the southern boundary of the discontinuous permafrost zone, see Isarin (1997a) for further details concerning this reconstruction. The inferred isotherms may be illustrated with the modern discontinuous permafrost zone in Canada, which is situated between the $-1.1$ and $-6.7\,^{\circ}$C mean annual air temperature isotherms (Brown 1969). It should be realized that temperature estimates based on periglacial evidence are threshold values, indicating maximum temperatures. This implies that no precise annual temperatures can be inferred for the areas north of the $-8\,^{\circ}$C isotherm and south of the $-1\,^{\circ}$C isotherm.

A total of 109 sites with YD periglacial climate evidence has been analyzed (see Fig. 1). We distinguished between high quality and low quality data, based on the documentation quality and geochronological control of the data. Emphasis was on those records in which age determination is based on radiometric (conventional $^{14}$C, AMS) measurements, the presence of the Laacher See tephra (dated to $\sim$11.0 $^{14}$C ky BP, Van de Bogaard and Schminke 1985) or well-dated soils. Interstadial soils and organic layers have been frequently observed and dated (e.g., Preece 1994) providing a maximum age of the YD periglacial activity at a specific site. Unfortunately, the age of a number of features is based on their lithostratigraphical position only, decreasing their value for the reconstruction.

### *2.2.2 The reconstruction of summer temperatures*

Paleoecological records constitute the primary source for the reconstruction of summer temperatures. A wealth of Weichselian Late Glacial pollen and plant macro fossil diagrams exists, in which the YD is a pronounced event (Walker et al. 1994). Marked changes in the vegetation from relatively closed to relatively open plant communities characterize this cool interval. As many as 140 pollen diagrams were analyzed (see Fig. 1), whereby emphasis was on diagrams that have: (1) a good geochronological control, (2) high pollen counts ($>$300 pollen grains per sample, frequently $>$500), and (3) a high sampling density.

We employed plant climate indicator species (CIS) to reconstruct minimum mean July temperatures. The reader is referred to Isarin

**Table 1** Overview of reconstruction methods discussed in this study (after Huijzer and Isarin 1997)

| Reconstructed temperature | Proxy data | Translation method | Format of parameter |
|---|---|---|---|
| 1. Maximum mean annual temperature | Periglacial | Modern analogues | Threshold |
| 2. Minimum mean temperature of the warmest month | Botanic | Climate indicator species | Threshold |
| 3. Maximum mean temperature of the coldest month | Periglacial | Modern analogues | Threshold |
| 4. Maximum mean temperature of the coldest month | Botanic and periglacial | Calculated from parameters 1 and 2 | Threshold |
| 5. Mean temperature of the warmest month | Entomologic | Mutual climatic range | Range |
| 6. Mean temperature of the coldest month | Entomologic | Mutual climatic range | Range |
| 7. Mean annual temperature | Entomologic | Calculated from parameters 5 and 6 | Range |
| 8. Mean temperature of the ablation season (May to September) | Glacial | Equilibrium line Altitudes | Range |

(1997b) for a detailed discussion of the CIS based temperature reconstruction. Iversen (1954) greatly contributed to this method that was adopted by numerous paleobotanists (e.g., Kolstrup 1980; Bohncke 1993; Zagwijn 1994). The CIS method is based on the specific temperature requirements of some plant species to grow and reproduce. Attention is focused on littoral and aquatic species, such as *Typha latifolia* and *Nymphaea alba*. It is assumed that these plants are the most reliable indicators since they migrate relatively quickly in response to climatic changes (Iversen 1973). Indicator values are based on a comparison of the present-day distribution of species (e.g., Hultén 1964) and July isotherms.

It should be noted that although climate indicator species indicate minimum mean July temperatures, we assume that these inferences are valid as minimum mean values for the summer months. Furthermore, we assume that the reconstructed minimum mean summer temperatures approximate true mean summer temperatures. These assumptions are based on a comparison of our CIS based temperature inferences with beetle data that have been processed with the mutual climatic range (MCR) method. This method allows for accurate estimates of the absolute mean temperatures of the warmest and coldest month (Atkinson et al. 1987). Unfortunately, compared to paleobotanical data, only a limited number of analyses have been published yet (compare Coope and Lemdahl 1995).

### *2.2.3 The reconstruction of winter temperatures*

We reconstructed mean winter temperatures from periglacial data, beetle evidence and a calculation based on the discussed minimum mean summer temperatures and maximum mean annual temperatures. In the latter method, it is assumed that the annual cycle approximates a sine function. The continuous permafrost zone, indicated by the ice-wedge polygons, indicates temperatures of the coldest month of $<-20\,^{\circ}$C (Lachenbruch 1962; Péwé 1966). This value agrees with Atkinson et al. (1987), who reconstructed mean winter temperatures from beetle data of $-15$ to $-20\,^{\circ}$C for the British Isles. Similar values have been produced for southern Sweden by Lemdahl (1991) and Hammarlund and Lemdahl (1994). The calculated estimates are largely in line with the above values.

## 3 Results

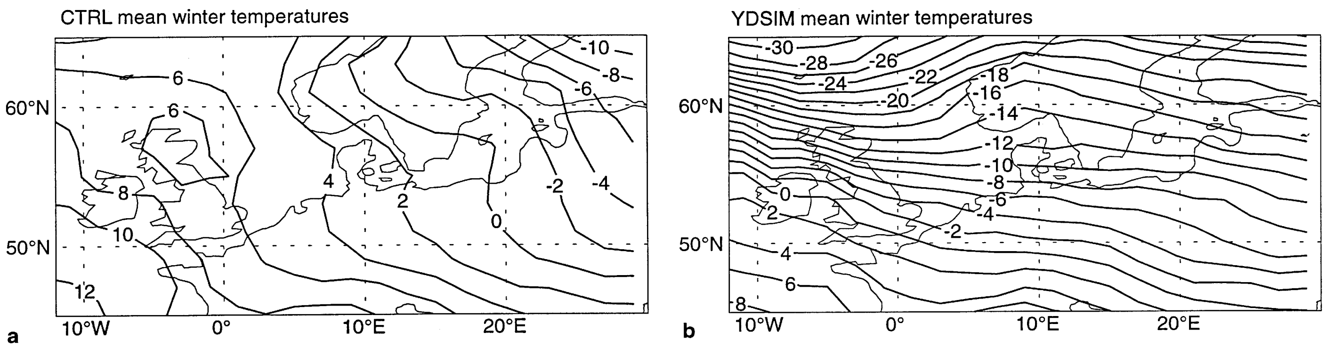
### 3.1 Simulation results

### *3.1.1 Winter*

The isotherms in experiment CTRL show a west-to-east gradient (Fig. 2a), as the temperature ranges from 10 °C in Ireland to below 0 °C in Eastern Europe. The results in YDSIM are strikingly different (Fig. 2b). Although a similar west-to-east gradient of 10 °C is present, ranging from 2 °C in Ireland to $\leq -8\,^{\circ}$C in Poland, the main temperature gradient is now situated in a south-to-north direction. Moreover, the gradient is steeper and the values are lower, since the temperatures range from 4 °C in central France to below $-20\,^{\circ}$C in northern Scandinavia. The noted variations lead to cooling everywhere in the study area in the YDSIM-CTRL difference plot (Fig. 2c). The deviations range from $-4\,^{\circ}$C in France to more than $-20\,^{\circ}$C north of Scotland and Norway. At 55 °N the difference varies between $-10$ and $-14\,^{\circ}$C.

### *3.1.2 Summer*

During summer the temperature distribution in CTRL shows a 10 °C south-to-north gradient as the value varies from 22 °C in France to 12 °C in Norway (Fig. 3a). In YDSIM this thermal gradient is doubled to 20 °C, with temperatures between 24 °C south of 48 °N to 2 to 4 °C in northern Europe (Fig. 3b). This difference between the two simulation experiments is



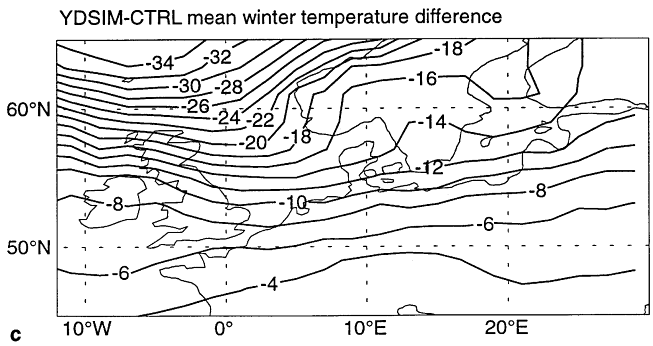




**Fig. 2a–c** Simulated mean winter temperatures (in °C) at sea level: **a** experiment CTRL, **b** experiment YDSIM and **c** YDSIM-CTRL difference. Contour interval: 2 °C

depicted in the YDSIM-CTRL plot (Fig. 3c). Close to the coast and near the Scandinavian ice sheet a cooling is evident, with differences ranging from close to $-6\,^{\circ}$C in Ireland to more than $-10\,^{\circ}$C in Scotland and W Norway. In contrast, this cooling changes to a warming of up to 4 °C in the interior of the European continent. Most striking in the YDSIM plot is the extreme thermal gradient along the coast and the southern margin of the Scandinavian ice sheet, where the temperature rises by 12–15 °C within two grid cells. This main front between cold air and relatively mild air lies at 56 °N.

### *3.1.3 Annual*

As expected, the pattern in the mean annual temperature plot is intermediate between the winter and summer results (see Fig. 4a–c). A strong south-to-north gradient of 20 °C is present in YDSIM. The mean annual temperatures in southern Europe are only slightly lower ($-2\,^{\circ}$C deviation in Fig. 4c), whereas an intense cooling is introduced over northern Europe (more than $-10\,^{\circ}$C difference). The main thermal gradient is situated over the intermediate region with the main axis at 55 to 56 °N. North of this axis annual average temperatures are below 0 °C in YDSIM (Fig. 4b).

## 3.2 Reconstruction results and comparison with the model simulation

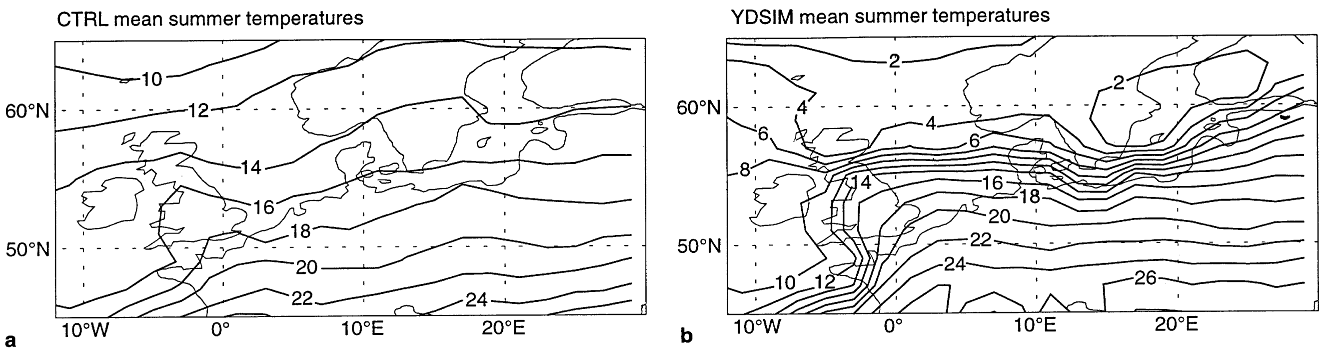
### *3.2.1 Winter*

The observed meteorological data show a southwestern-to-northeastern temperature gradient indicating mild temperature conditions in the Atlantic regions (Fig. 5a). The results of CTRL (Fig. 2a) are largely in accordance with observations. Winter temperatures range from 7 °C in Ireland to $-1\,^{\circ}$C in S Sweden and E Poland (Fig. 5a). The reconstructions suggest strong south-to-north gradient with mean winter temperatures of $-15\,^{\circ}$C at about 50 °N and $-20\,^{\circ}$C at 55 °N (Fig. 5b). The magnitude and south-to-north orientation of this gradient agree with the simulated gradient (YDSIM; Fig. 2b).

The difference plot (reconstructions minus observation; Fig. 5c) shows a large cooling in the Atlantic areas of $>25\,^{\circ}$C in Ireland and Scotland, and between 20 and 25 °C in England, Denmark and the Norwegian coast. The more continental and southern regions of the study area experienced a cooling of 15 to 20 °C when compared to the present-day situation. A comparison of Figs. 2c and 5c shows that the reconstructed and the simulated gradients are comparable west of 5 °E (i.e., $\geq 5\,^{\circ}$C in 5° of latitude). However, the reconstructed iso therms seem to be displaced southward by $\sim 5^{\circ}$ of latitude, or approximately two AGCM gridcells. Also, the orientation of the isotherms does not completely match. Compatibility with minor deviations ($\leq 5\,^{\circ}$C) are observed in Finland, along the Norwegian coast and Poland. The difference between the two datasets is largest in Ireland ($\geq 15\,^{\circ}$C) and S England ($\geq 10\,^{\circ}$C).

### *3.2.2 Summer*

The observed meteorological data show a south-east to north-west temperature gradient (Fig. 6a). Comparing



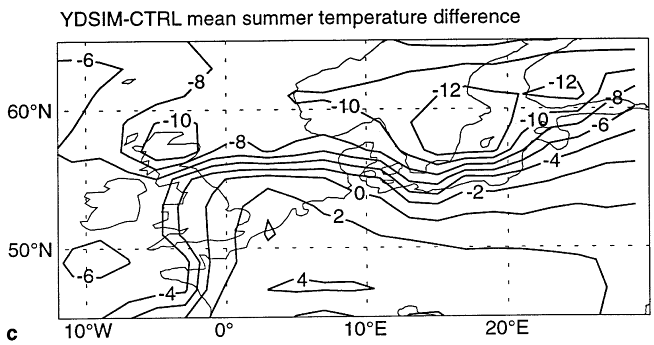




**Fig. 3a–c** Simulated mean summer temperatures (in °C) at sea level: **a** experiment CTRL, **b** experiment YDSIM and **c** YDSIM-CTRL difference. Contour interval: 2 °C

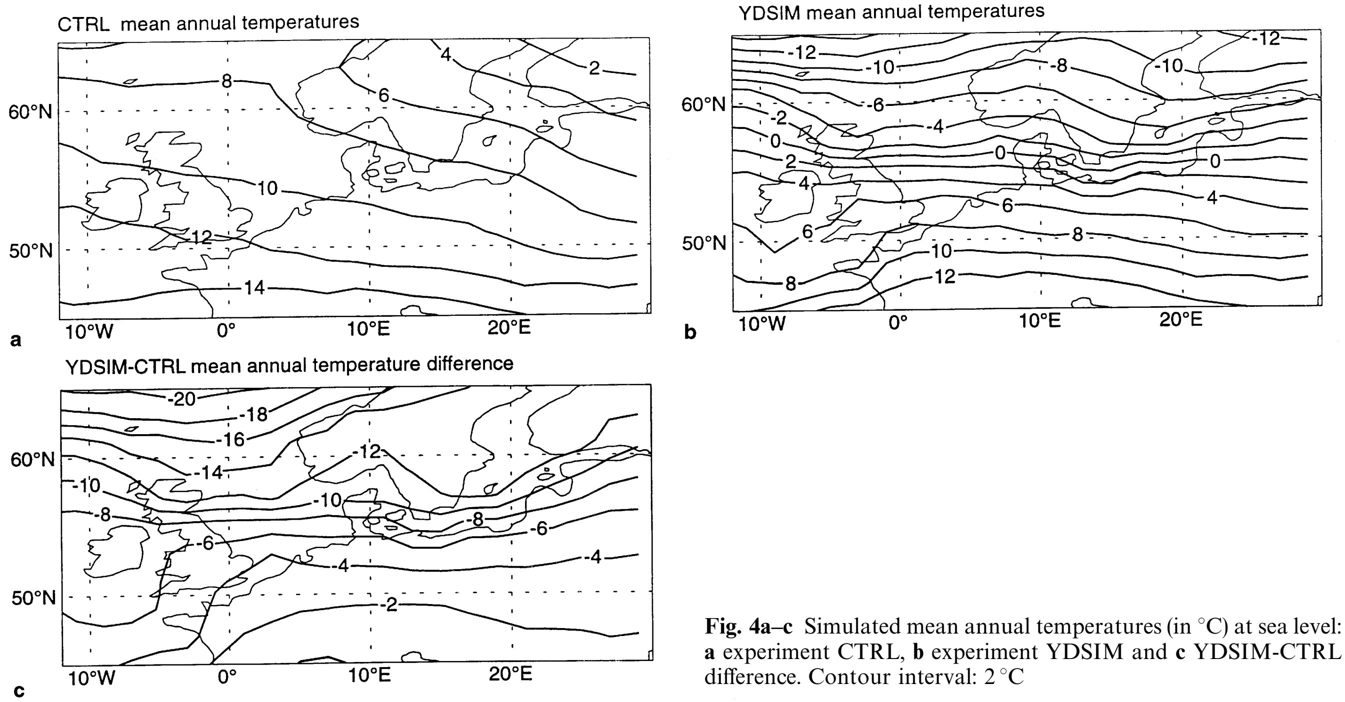



**Fig. 4a–c** Simulated mean annual temperatures (in °C) at sea level: **a** experiment CTRL, **b** experiment YDSIM and **c** YDSIM-CTRL difference. Contour interval: 2 °C

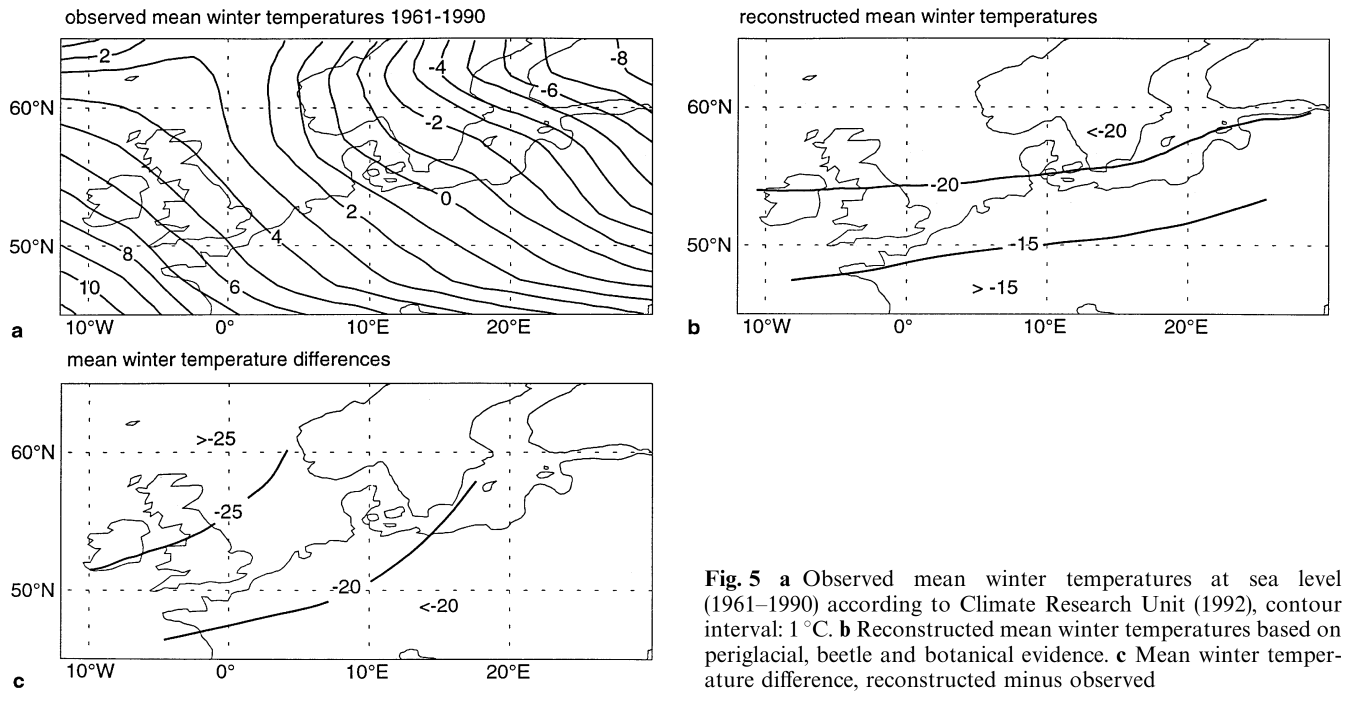
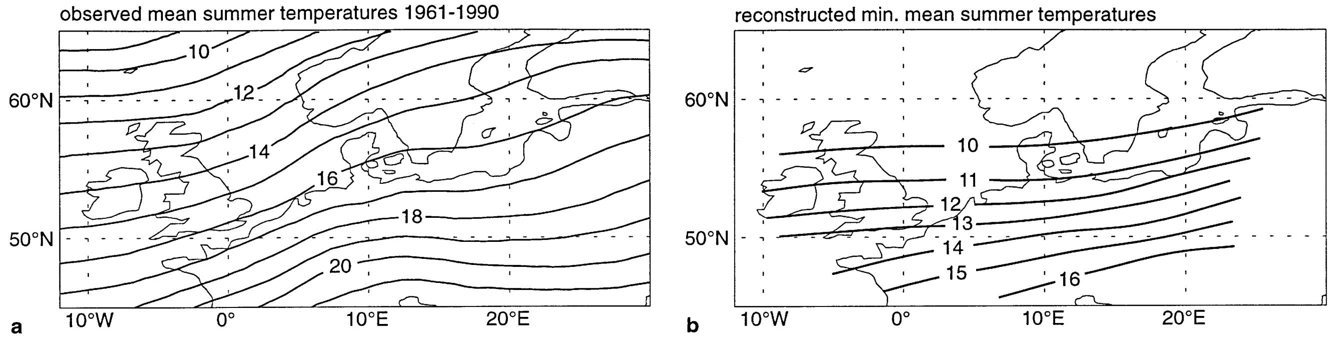



**Fig. 5** **a** Observed mean winter temperatures at sea level (1961–1990) according to Climate Research Unit (1992), contour interval: 1 °C. **b** Reconstructed mean winter temperatures based on periglacial, beetle and botanical evidence. **c** Mean winter temperature difference, reconstructed minus observed

Fig. 6a with the simulated data (Fig. 3a) shows that the AGCM satisfactorily simulates the modern climate. Over the continent, however, the simulated temperatures are a few degrees too high. In general, the reconstructed YD summer temperature gradient (Fig. 6b) has an orientation similar to the present one. However, the reconstructed YD summer temperatures are lower. The difference plot between the observations and reconstructions (Fig. 6c) suggests a slight increase in cooling from west to east at 55 °N. The minimum summer difference ranges from $-3\,^{\circ}$C in Ireland to $-6\,^{\circ}$C in S Sweden. A comparison of this map with the YDSIM-CTRL difference plot (Fig. 3c) shows that maximum cooling takes place in S Sweden in both the AGCM experiment and the climate reconstruction. North of 56 °N the simulated cooling appears too strong, as differences of $\geq 6\,^{\circ}$C may be inferred (Fig. 3c minus Fig. 6c). In continental Europe the opposite is evident, since the simulation implies a warming of up to 4 °C, whereas the reconstruction suggests a cooling of about 4 °C. Thus, an anomaly of 6 to 8 °C is inferred. Both Figs. 3c and 6c show that values are comparable along the Atlantic coast (i.e., Ireland and the French and British west coasts), as well as in areas



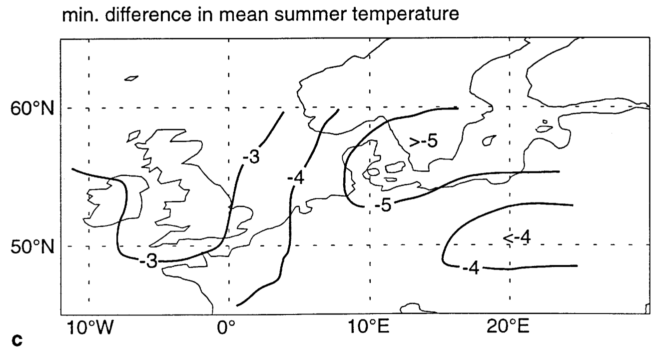
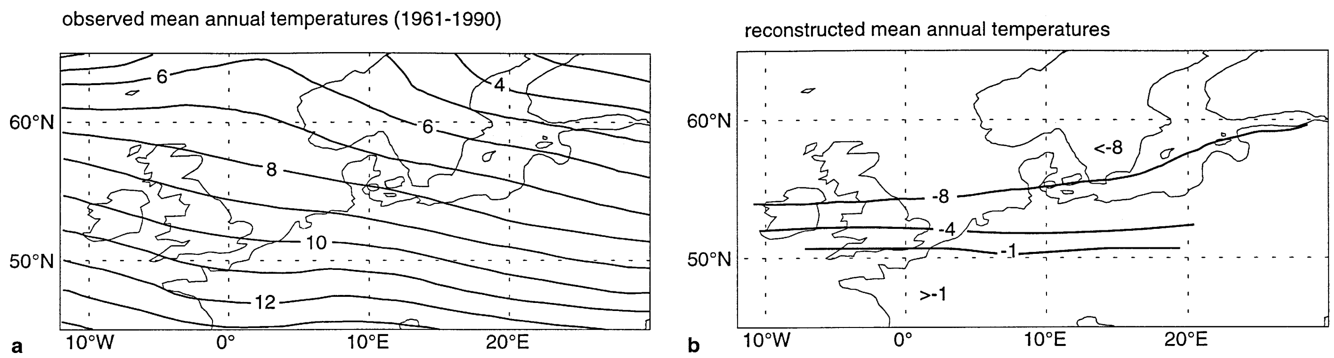




**Fig. 6** **a** Observed mean summer temperatures at sea level (1961–1990) according to Climate Research Unit (1992), contour interval: 1 °C. **b** Reconstructed minimum mean summer temperatures based on botanical evidence. **c** Minimum mean summer temperature difference, reconstructed minus observed



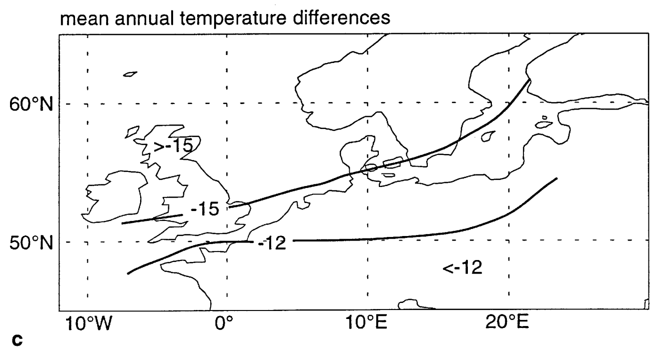




**Fig. 7** **a** Observed mean annual temperatures at sea level (1961–1990) according to Climate Research Unit (1992), contour interval: 1 °C. **b** Reconstructed mean annual temperatures based on periglacial evidence. **c** Mean annual temperature difference, reconstructed minus observed

south of the Scandinavian ice sheet (i.e., Denmark and N Poland).

### *3.2.3 Annual*

The observed mean annual temperature gradient (Fig. 7a) is consistent with that of the CTRL experiment (Fig. 4a). However, simulated annual temperatures are up to 2 °C higher. The reconstructed mean annual temperatures show a south-to-north gradient of 7 °C between ~50 °N and 55 °N (Fig. 7b). The −1 and −8 °C isotherms follow the southern limits of the discontinuous and continuous permafrost zones. The difference plot (Fig. 7c; reconstruction minus observations) suggests a progressive cooling from south-east to north-west with deviations of −12 °C in Poland to ≤−15 °C in Ireland and England. A comparison of the reconstructed and the simulated difference plots (Figs. 4c and 7c) suggests comparable coolings (deviations of ≤5 °C) in Norway, S Sweden, Finland and along the Baltic coast, i.e., near the margins of the Scandinavian ice sheet. Large discrepancies, up to 10 °C, are observed south of 57 °N and west of 10 °E.

## 4 Discussion

When comparing AGCM results with climate reconstructions it is important to realize that three factors of uncertainty are of influence: (1) the AGCM performance, (2) the set of boundary conditions used to drive the simulation, and (3) the climate reconstruction. Since the CTRL results approach the observational data, the AGCM performance may be considered of secondary importance compared to the other two factors.

### 4.1 Winter

A deviation between the simulated and reconstructed winter temperatures was found in the central and western part of the study area, with a maximum near the Atlantic coast. What caused this deviation: the reconstructions, the boundary conditions or model error?

The reconstructions suggest very low mean winter temperatures in Europe, with the −20 °C isotherm located at 54–55 °N (Fig. 5b). The −20 °C isotherm was based on the southern limit of the continuous permafrost. It should be noted that a YD age of some ice-wedge polygons from the British Isles (see Ballantyne and Harris 1994) and Denmark (see Kolstrup 1991) is ambiguous. Nevertheless, beetle records from the British Isles and S Sweden indicate mean temperatures of the coldest month as low as −20 °C (Atkinson et al. 1987; Lemdahl 1991). Calculations assuming an annual cycle following a sine function suggest slightly lower winter temperatures for this area. On the basis of the combination of evidence, we conclude that the reconstructed winter temperatures are reliable.

Alternatively, the set of boundary conditions used in YDSIM may have caused the deviation in the simulation-reconstruction comparison. The fact that the maximum disagreement was found in Ireland and Britain, i.e., close to the Atlantic Ocean, may indicate that the N Atlantic was even cooler during the YD than prescribed in YDSIM, particularly in a latitudinal zone ranging from 55 °N to 50 °N. As noted in Sect. 2.1, the low SSTs in YDSIM caused the mean position of the winter sea-ice margin to be situated at 56 °N, except near the W European coast (at 15°W) where the sea-ice boundary was prescribed at 62 °N (Renssen et al. 1996). A lowering of SSTs in the latter region by a few degrees would result in a southward expansion of the N Atlantic sea-ice. Such a southward sea-ice expansion, possibly as far south as 52 °N, would indeed explain the very low winter temperatures reconstructed for Ireland, Britain and The Low Countries. The simulated temperature gradient was similar to the reconstructed one, but with the isotherms shifted by 5° of latitude to the north in the simulation. Therefore, it is likely that a simulation with the proposed sea-ice expansion would result in a cooling that is comparable to the reconstruction.

However, it should be noted that model performance may have contributed to the difference with the reconstructed temperatures. We assume that the model error is of similar magnitude in CTRL and YDSIM, and expect that the influence of this error is eliminated by comparing the YDSIM-CTRL difference with the reconstructions. As discussed in Sect. 2.1, the model shows a tendency to generate a southerly flow too easily during winter (see Sect. 2.1). Thus, if the magnitude of this erroneous southerly flow was stronger in YDSIM than in CTRL, this may have produced too high simulated temperatures.

The simulation and reconstruction were comparable in the northern part of the study area, i.e., along the Norwegian coast and Finland. This gives us confidence in the prescribed winter surface ocean conditions in this region. The large cooling north of 55°N in Fig. 2c is a result of the low SSTs and the resulting sea-ice limit, reflecting the absence of the warm Gulf Stream. As explained in Sect. 2.1, the winter SST set was derived from an atmosphere-ocean model experiment in which the thermohaline circulation was halted (Schiller et al. 1996). The inferred influence of a cold N Atlantic on N European climate agrees with the theory of Broecker (1992), who suggested that the YD climate was caused by a shut-down of the thermohaline circulation in the Atlantic Ocean under influence of meltwater influxes. On the other hand, paleo-evidence from ocean cores suggests that the thermohaline circulation was functioning during YD time (Veum et al. 1992; Sarnthein et al. 1994), implying that another mechanism caused the cooling of the N Atlantic Ocean.

Renssen et al. (1996) proposed that an expanded sea-ice cover in the N Atlantic Ocean was the driving force behind the YD winter cold in Europe. They showed that a strong zonal atmospheric circulation was present in YDSIM in the N Atlantic/European sector and that associated strong westerlies transported cold polar air, cooled over sea-ice and ice sheets, to NW Europe. Isarin et al. (1997) showed that such a westerly circulation is in full agreement with dunes of YD age in NW Europe, indicating westerly depositional winds.

The mid-latitudinal position of the winter sea-ice margin is supported by several independent paleorecords. Bond et al. (1993) observed ice-rafted sediments of YD age in an Atlantic ocean core located as far south as 50°N. Moreover, Mayewski et al. (1994) concluded on the basis of low sea-salt concentrations in the GISP 2 Greenland ice core that N Atlantic sea-ice expansion was considerably larger during the YD, limiting the sea-salt transport to Greenland, than during the preceding Late Glacial Interstadial and following Preboreal periods. However, this remains a subject of controversy since Ruddiman and McIntyre (1981) suggested, on the basis of foraminiferal and coccolith productivity estimates, that the N Atlantic sea-ice was not greatly expanded during the YD event compared to the preceding warm interval.

### 4.2 Summer

The comparison between the simulated and reconstructed summer temperatures shows that north of 56°N the simulated values were lower than suggested by the reconstruction. Concerning the reliability of the reconstruction, we note that the estimated mean summer temperatures, based on climate indicator species, largely agree plant with beetle data. However, the reconstructed values do not agree with temperature reconstructions based on variations in the equilibrium-line altitudes of glaciers in Scotland and W Norway (e.g., Sissons 1979; Larsen et al. 1984; Sutherland 1984; Ballantyne 1988). ELA based temperature estimates, 6 to 8 °C at sea level, would increase the summer temperature anomaly between reconstruction and the present situation in these areas to between −7 and −5 °C. The latter values would in fact be in better agreement with the simulated temperature differences (−7 °C) than the CIS based anomaly (−3 °C). However, it must be noted that ELA based temperature calculations require the assessment of former precipitation values (Ballantyne 1988). Accurate YD precipitation estimates are difficult to obtain and therefore modern values are used, which in turn may very well produce ambiguous YD temperature estimates.

The simulation-reconstruction comparison shows a large deviation for continental Europe. The simulations suggest a warming of up to 4 °C compared to CTRL, whereas the reconstruction indicates a temperature depression of about 4 °C. It should be recalled that the CIS based temperature are minimum values, implying that the actual cooling may have been smaller. The climate reconstruction of Fig. 6b is based on a large number of sites that show a consistent pattern. Our estimates are largely comparable with the reconstructions of Lowe and NASP Members (1995), who suggested an overall cooling of 5 to 7 °C compared to the present-day situation. In any case, the reconstructed temperatures suggest that the simulated summer warming in Europe is unrealistic. Renssen (1997) showed that in YDSIM the 12 ky calendar BP insolation caused a summer warming in continental Europe of more than 5 °C compared to CTRL. In YDSIM this warming was partly counteracted by the response to other boundary conditions, in particular the cooled ocean surfaces and ice sheets, causing a cooling ranging from 5 °C in coastal regions to 2 °C in inland areas. The net result was a warming of up to 4 °C as visible in Fig. 3c. If we assume that the three effects mentioned (cold oceans, ice sheets and insolation) were the most important ones, the reconstructed continental cooling of a few degrees suggests that the combined cooling effect of cooled oceans and ice sheets was larger than the warming caused by the increased summer insolation. This is not shown in the model results.

It should be realized that another factor may have been important in forcing the summer temperatures: the atmospheric dust content. In YDSIM the present amount of aerosols was prescribed. However, the Greenland ice core dust record suggests that the aerosol loading was larger during YD time than today (Dansgaard et al. 1989; Mayewski et al. 1994). The effect of an increased atmospheric dust content could be cooling of the atmosphere due to an increase of planetary albedo. In the mid-latitudes this cooling effect may have been especially strong during summer, as during this season a relatively large part of the incoming energy depends on the incoming solar radiation as opposed to winter when the transport of energy from lower latitudes is more important (Peixoto and Oort 1992). On the other hand, a modelling study by Overpeck et al. (1996) suggests that the opposite may occur, as in their results dust actually warmed the climate in some locations. Clearly, it would be interesting to perform a YD simulation with an AGCM that includes a model for the dust cycle (e.g., Joussaume 1993).

As discussed in the section on winter temperatures (Sect. 4.1), we cannot exclude that model errors influence the simulation result. Possibly the model reacted incorrectly to the YD boundary conditions. For instance, the model may have produced an incorrect flow in response to the Laurentide and Scandinavian ice sheets. Alternatively, the simulated winds could be too weak during summer, causing an underestimation of the advection of cool maritime air from the Atlantic. The model could also have produced an unrealistic cloud cover in relation to the increased summer insolation.

A summer warming in Europe was also found in other AGCM studies on Late Glacial climates. For instance, Kutzbach et al. (1993) simulated in a 12 ky BP experiment a warming in continental Europe, which was attributed to the increased insolation due to changes in orbital parameters. Rind et al. (1986) performed AGCM sensitivity experiments on the YD climate and found a 6 °C warming in E Europe. They suggested that this warming was caused by subsidence of air over the Scandinavian ice sheet. In YDSIM the maximum warming is found between 5 °E and 10 °E (see Fig. 3c). Since this location is situated southwest of the Scandinavian ice sheet, we exclude subsidence of air as the main cause of the anomalously high temperatures.

There is agreement on the location of the maximum cooling during summer, i.e., S Sweden (compare Figs. 3c and 6c). This points to the importance of the cooling effect of the Scandinavian ice sheet on its surroundings during summer. Isarin et al. (1997) showed that the mean wind direction during summer changed in a region from S Sweden to Poland from W in CTRL to WNW–NW in YDSIM. Such a change in wind direction could imply transport of cold air from the Scandinavian ice sheet to S Sweden and N Poland, thus explaining the cooling in this region.

The simulation results are consistent with the reconstruction for regions adjacent to the Atlantic Ocean (W France, SW England and Ireland). This agreement suggests that the SST set prescribed in YDSIM, which was based on the SST estimates of Sarnthein et al. (1995), is realistic, at least for a zone stretching from 45 °N to 55 °N.

### 4.3 Annual

The difference in the mean annual temperature estimates reflects the discussed winter and summer results. The YDSIM-CTRL result is too warm compared to the reconstructed cooling. In England and Ireland this deviation is mainly caused by the anomalously high simulated winter temperatures in this region. Similarly, the too high temperatures in inland Europe are a result of the simulated summer conditions.

What can be said about the reliability of the reconstructed mean annual temperatures? The reconstruction of the $-8$, $-4$ and $-1$ °C isotherms in Fig. 7b is based on a large number of well-dated periglacial features (Isarin 1997a) and therefore considered reliable. Moreover, calculations based on beetle data support our reconstructed annual temperatures (compare Atkinson et al. 1987).

The results clearly show an increase of the annual temperature ranges compared to today, especially near the Atlantic Ocean. The difference between the reconstructed summer and winter temperatures is around 30 °C throughout the study area. In contrast, in the present climate the continentality clearly increases inland, with annual temperature ranges varying from 7 °C in Ireland to 17 °C in Poland. The noted change in the continentality can only be explained by extremely low surface temperatures during the YD winters, presumably caused by the southward expansion of N Atlantic sea-ice. The presence of land in the southern North Sea region during YD time (see Fig. 1) may have increased the continentality further, particularly in E England, N France, The Low Countries, N Germany and Denmark. It is noteworthy that YD temperatures in NW Europe may have shown similarities with the present-day temperature regime of sub-polar oceanic W Alaska (compare Karte 1979).

## 5 Conclusions

1. The simulation result for the winter months agrees with reconstructions for the northern part of the study area, with corresponding temperature depressions of approximately 25 °C along the Norwegian coast and 15 °C in Finland. Moreover, the simulation and reconstruction results show a comparable south-to-north temperature gradient west of 5 °E, i.e., 5 °C in 5° of latitude. These notions give confidence in the prescribed sea surface conditions north of 60 °N.

2. We inferred a maximum deviation between the simulated and reconstructed winter temperatures of more than 15 °C in Ireland. We suggest that the mean winter N Atlantic sea-ice margin may have been positioned further south during the YD than prescribed in the simulation, possibly at 52 °N.

3. The simulation results agree with the reconstructed summer temperatures in W France, Ireland and SW England. This may indicate that the prescribed summer SSTs were realistic at 50 °N. However, the simulated summer warming of 4 °C in continental Europe seems unrealistic for the YD situation as the reconstructions suggest a cooling of about 4 °C. We propose that the high simulated summer temperatures are caused by the model's response to the prescribed summer insolation. However, it must be noted that we prescribed a lower aerosol content (i.e., the modern value) than suggested by ice cores. This may have contributed to the noted high temperatures in our YD simulation. Also, the model could have had an incorrect response on other boundary conditions (e.g., ice sheets, cooled oceans).

4. Both the reconstruction and the simulation suggest that the Scandinavian ice sheet had a cooling effect on S Sweden and surroundings, possibly involving an increase of northwesterly winds during summer.

5. Our results suggest that during the YD the yearly temperature range increased throughout the study area to values of about 30 °C. Continentality increased most strongly near the Atlantic Ocean where, for example, in Ireland the annual temperature range today is 7 °C. This increase primarily reflects the extremely low winter temperatures.

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