Tellus (1997), 49A, 347–361
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TELLUS
ISSN 0280–6495

An analysis of extra-tropical storms in the North Atlantic region as simulated in a control and 2 × CO_2 time-slice experiment with a high-resolution atmospheric model

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(Manuscript received 17 July 1996; in final form 24 January 1997)

ABSTRACT

Climate simulations with an atmospheric model at T106 resolution are analysed to study the effect of greenhouse warming on the North Atlantic wind climatology. The analysis is based on winter data from 2 high-resolution numerical (time-slice) experiments: a control run and a $2 \times CO_2$ run. The storm track (500 hPa height variability), mean sea level pressure and average surface winds in the control experiment compare well with 5 years of analyses from the European Centre for Medium-Range Weather Forecasts (ECMWF) for the period 1991–1995. The mean sea level pressure field however differs markedly from a pressure climatology of 39 years (1955–1993) of analyses prepared by the Norwegian Meteorological Institute (DNMI) and from the ECMWF analysis for the period 1986–1990. It is argued that these differences are a manifestation of the low-frequency variability of the climate system in the North Atlantic region. The $2 \times CO_2$ experiment differs slightly from the control experiment in several aspects: (1) it has, relative to the control experiment, a negative pressure anomaly of 5 hPa over Scandinavia and a weaker positive anomaly over the central North Atlantic; (2) associated with this pressure anomaly pattern there is a wind direction anomaly over the eastern North Atlantic; (3) there is a small intensification of the variability of the 500 hPa height, which is most pronounced over the North Sea, the Bay of Biscay and central Europe; (4) the number of mid-latitude storm events decreases somewhat in most of the North Atlantic, particularly north of 55°N, but increases in the North Sea and the Bay of Biscay; (5) the overall frequency of deep depressions decreases slightly while the frequency of weak depressions increases. Considering the high level of observed interannual variability, it is not possible to say whether the anomalies in the $2 \times CO_2$ experiment are caused by the external greenhouse forcing or are just the result of natural climate variability. The model results suggest that the effect of a greenhouse-induced decrease in the meridional temperature gradient near the surface and the increase of this gradient in the upper troposphere balance approximately, in such a way that the net effect of CO_2 doubling on mid-latitude storminess is small.

1. Introduction

Off-shore industries, fisheries, shipping companies and the insurance business, are very sensitive to extra-tropical gales and the associated ocean waves and surges. From time to time it is postulated in the public press that storms or waves have increased in strength or frequency over the last century and that this increase is associated with an enhanced greenhouse effect. Bacon and Carter (1991) presented evidence of positive trends in significant wave

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heights at various North Atlantic sites, between the early 1960s and the late 1980s. A similar trend was obtained by Bouws et al. (1996) from an analysis of shiprouting charts. Some evidence of recent increased storminess around the North Atlantic was found, e.g., by Dronia (1991), Schinke (1993) and Stein and Hense (1994), but Nicholls et al. (1996) have doubts about the consistency of the meteorological analyses from which these increases were derived. According to Nicholls et al. there is no clear evidence of any uniform increase in extra-tropical storminess. Schmidt and von Storch (1993) and Kaas et al. (1996) have also argued that these apparent increases are manifestations of either observational inhomogeneities in the data on which the statements are based or simply a consequence of natural variability in the climate system on decadal time scales since many of the postulated increases are based on very short time series.

Nevertheless, there is a considerable interest in possible future effects of increased greenhouse gas concentrations on the storm climate. A suitable tool for the study of this question is analysis of numerical model simulations. The GCM's that are normally used for global change studies are not suitable for this purpose, because, for lack of computer capacity, their resolution is too coarse to simulate extreme aspects of developments in the tropics and the extra-tropics. It is for this reason that the so-called time-slice simulations have been made (Bengtsson et al., 1995, 1996). These simulations are made at T106 resolution with the global ECHAM3 model. This is the highest horizontal resolution to date in a climate change experiment. As such these simulations are well suited for the study of extreme events. A disadvantage is their limited duration (a little over 5 years), which makes it impossible to distinguish effects of natural variability on decadal time scales.

Bengtsson et al. (1995) were the first to analyse these simulations. They concentrated on tropical cyclones and concluded that the model was capable of simulating the strength, the spatial structure, the geographical distribution and seasonal variability of these extreme events with a degree of realism unattainable with lower resolution models. In a second paper, Bengtsson et al. (1996) studied the tropical cyclones in an atmosphere with doubled CO_2 concentration. They found a substantial reduction in the number of tropical cyclones particularly in the Southern Hemisphere in the $2 \times CO_2$ experiment relative to both the control experiment and the "observed" data. They argued that this decrease is consistent with the overall increased static stability of the tropical troposphere in the $2 \times CO_2$ experiment relative to the control, due to a warming in the upper troposphere through deep convection. Furthermore they found a weaker Hadley circulation and somewhat reduced evaporation.

This paper studies the occurrence of extra-tropical storms in the numerical simulations of Bengtsson et al. (1995, 1996), in particular in the North Atlantic region. The discussion will be given in terms of near surface (10 m) winds, mean sea level pressure (MSLP), the 500 hPa height variability and the frequency of storms. The following questions are addressed. (1) To what extent do high resolution climate simulations reproduce the observed storms in the North Atlantic region? (2) What is the effect of CO_2 doubling? The physical mechanisms driving tropical and extra-tropical storms are very different. Therefore, it is obvious that one can not extrapolate the tropical cyclone results of Bengtsson et al. (1996) to the extra-tropics. Models of the equilibrium response of climate to an increase in atmospheric CO_2 concentration show the largest surface warming at the winter pole. The resulting decrease in the pole-to-equator temperature gradient could weaken the midlatitude baroclinic eddies (cyclones) during northern winter. Most models however also predict an important competing change, i.e., an increase in the pole-to-equator temperature gradient in the upper troposphere. Since the dominant wintertime baroclinic eddies are coherent through the depth of the troposphere, this could lead to a strengthening of baroclinic eddies. It is unclear how these effects add up (Held, 1993). So far, there is little agreement between models on the changes in mid-latitude storminess that might occur in a warmer world. Conclusions about extreme storm events are obviously even more uncertain (Kattenberg et al., 1996; see also Beersma, 1994).

The paper is organized as follows: The time-slice simulations as well as the observations used for verification of the control simulation are described in Section 2. Section 3 addresses the first of two main questions raised in this paper: How well does the ECHAM3 model at T106 resolution reproduce the observed wind climate over the North Atlantic? In Section 4, the model is used to answer the second question: Does the strength or frequency of storms in the region change as a result of enhanced greenhouse gas concentrations? Finally, a discussion and the conclusions are given in Section 5. We have organized the figures in such a way that the $2 \times CO_2$ results can be easily compared with the reference runs.

2. Description of the climate experiments and the observations

The ECHAM atmospheric model has evolved from the numerical weather forecasting model of the European Centre for Medium-Range Weather Forecasts (ECMWF) and has been modified at the Max-Planck Institute for Meteorology in Hamburg for global climate simulations. The main characteristics of the third generation model (ECHAM3) and the employed physical parameterizations related to storms are given below. The prognostic variables in ECHAM3 include vorticity, divergence, temperature, surface pressure, water vapour and cloud water. The reference resolution of the spectral model is T42 with 19 hybrid vertical levels, but the model is set up to use resolutions in the range T21 to T106. A semi-implicit, leap frog time integration scheme with a resolution dependent time step is used. The vertical turbulent transfer of momentum, heat, water vapour and cloud water is based on the Monin-Obukhov similarity theory for the surface layer and the eddy diffusivity approach above the surface layer (Louis, 1979). The drag and heat transfer coefficients depend on roughness length and Richardson number, and the eddy diffusion coefficients depend on wind stress, mixing length and Richardson number (Brinkop, 1991, 1992). The roughness length over sea is based on the modified Charnock formula (Miller et al., 1992). Over sea-ice the roughness length is constant (0.01 m) and over land it depends on vegetation and orography. The effect of orographically excited gravity waves on the momentum budget is based on linear theory and dimensional considerations (Palmer et al., 1986; Miller et al., 1989). The vertical structure of the momentum flux induced by the gravity waves is calculated from a local Richardson number. A more detailed description of the ECHAM3 model can be found in DKRZ (1992).

The climate experiments that are used here are the so-called time-slice simulations with the ECHAM3 atmospheric model at T106 resolution (Bengtsson et al., 1995, 1996). The only change with respect to the reference T42 resolution version is an adjustment of the horizontal diffusion for wavenumbers smaller than 30 (Bengtsson et al., 1995). For the control experiment the high resolution ECHAM3/T106 model was integrated for a little over 5 years using the climatological sea surface temperatures (SST) and sea ice limits from the period 1979–1988 and 1980 CO_2 concentrations as boundary conditions.

The $2 \times CO_2$ ECHAM3/T106 experiment makes use of earlier simulations with the ECHAM1 model at T21 resolution coupled to the Large Scale Geostrophic (LSG) ocean model (Cubasch et al., 1992). These earlier simulations included a control experiment and an experiment in which the coupled model was integrated under the assumption of a monotonous increase in CO_2 concentration (about 1% per year), leading to a doubling of CO_2, with respect to the level in the early 1980s, after some 60 years of integration. The averaged SST anomaly pattern obtained from the coupled ECHAM1/T21 experiment at the time of CO_2 doubling was added to the climatological SSTs (1979–1988) and this was used as a boundary condition for the 5-year $2 \times CO_2$ integration with the ECHAM3/T106 atmospheric model. In the $2 \times CO_2$ experiment the only change to the atmospheric model compared to the control experiment was a doubling of the CO_2 concentration. In both time-slice experiments the interannual fluctuations of SST and sea ice are therefore artificially suppressed. One may anticipate that this also reduces the atmospheric interannual variability.

The ECHAM3/T106 control experiment is compared with ECMWF analyses from the period 1986–1995. The ECMWF analyses were retrieved on the same grid as the ECHAM3/T106 model. Model and analysed data are given twice daily (00 and 12 UTC). Because the most and the heaviest storms occur during winter the analysis is restricted to the winter season (December, January, February). The study area covers the North Atlantic Ocean and most of the European continent (14–86°N, 83°W–54°E). Changes in the ECMWF assimilation model (e.g., horizontal and vertical resolution, and modifications to the boundary layer scheme) have led to inhomogeneities in the analysed data. It is, however, not known how large these inhomogeneities are. It is for this reason that reanalysis projects have been started at ECMWF and NMC (National Meteorological Center, USA).

In addition, mean sea level pressure data for the time period 1955–1993 prepared by the Norwegian Meteorological Institute (DNMI) are used for evaluating the model results. The DNMI pressure fields were prepared manually until 1981, while the 1982–1993 data are taken from analyses made as a part of operational numerical forecasting systems. The data are available four times daily and cover the northern part of the North Atlantic on a grid with a resolution of 75 km. In view of high spatial and temporal resolution the DNMI data are particularly suitable to study the high frequency storm activity in the northern North Atlantic.

3. The climate of the control experiment

In this section various aspects of the climate of the ECHAM3/T106 control experiment are com-

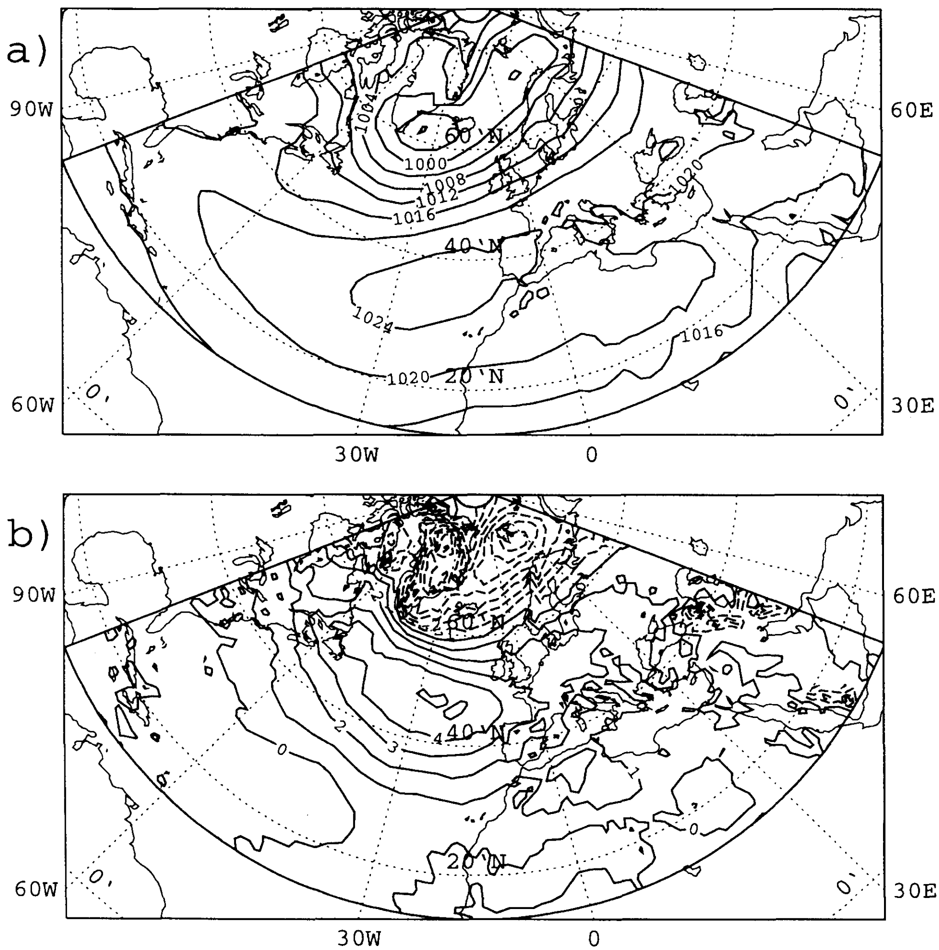

Fig. 1. Mean sea level pressure (DJF). (a) ECMWF analyses: 1991–1995; (b) ECMWF analyses: 1991–1995 less 1986–1990; (c) ECHAM3/T106: control experiment (6 winters); (d) ECHAM3/T106: $2 \times CO_2$ — control (6 winters). Negative contours are dashed.

pared with those of the ECMWF analyses as well as the DNMI data. In particular, mean sea level pressure (MSLP), surface wind fields and 500 hPa fields are examined.

3.1. *Mean sea level pressure*

The atmospheric circulation over the North Atlantic area is characterised by a large variability. This variability remains a dominant feature up to decadal time scales. A recurrent observed anomaly pattern on a decadal time scale in the MSLP is associated with the North Atlantic Oscillation (NAO), see e.g., Wallace and Gutzler (1981). The NAO is characterised by variations in the winter north-south pressure gradient over a large area of the central North Atlantic between Iceland and the Azores (Hurrell, 1995). During winters with a large pressure gradient (high NAO index) the mean zonal flow over Europe is much stronger than during winters with a small pressure gradient (low NAO index). The NAO accounts for about one third of the regional variance of the winter MSLP field (Deser and Blackmon, 1993; Hurrell, 1995). There are peaks of variability in the NAO on biennial and quasi-decadal time scales (Deser and Blackmon, 1993; Hurrell and van Loon, 1995).

The NAO shows up prominently in the ECMWF analysis. The difference in average

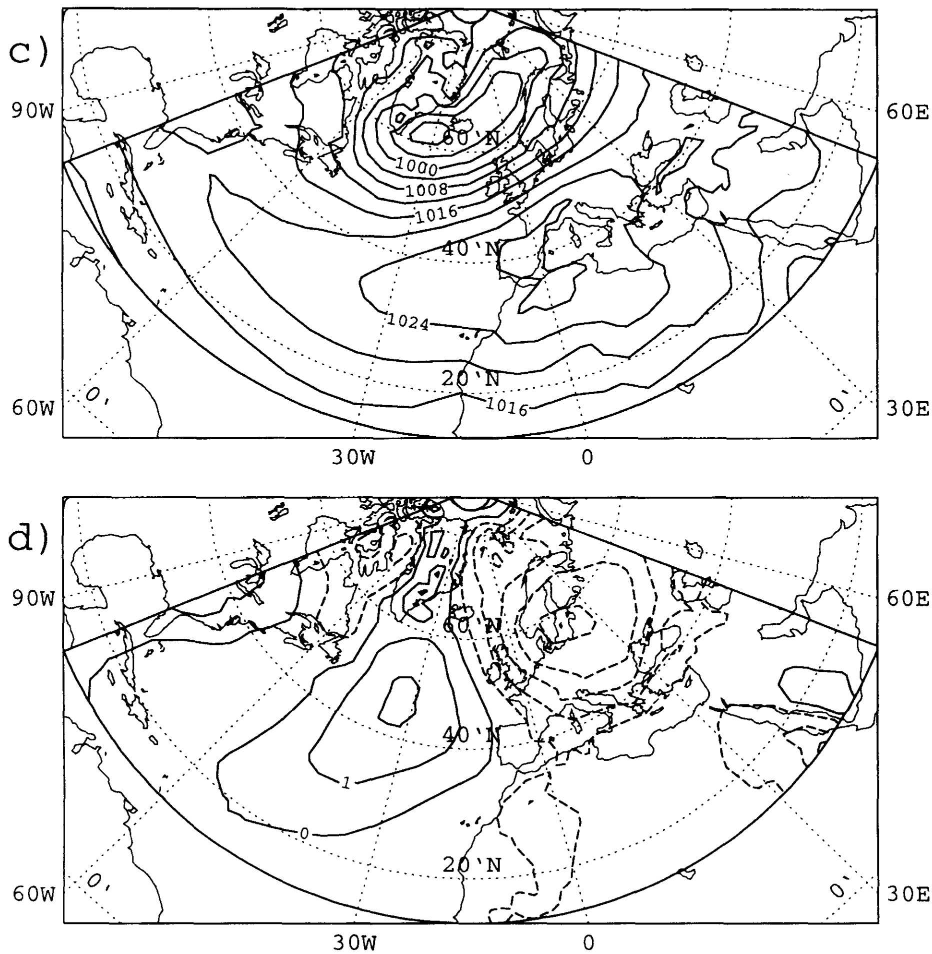

Fig. 1 (cont'd).

MSLP between the winters 1991–1995 (high phase) and the winters (1986–1990) (low phase) is presented in Fig. 1b. The pattern in this figure is similar to the NAO pattern given in Hurrell (1995). This shows that, to a certain extend, these two 5-year periods represent the two opposite modes of the NAO, which is confirmed by the NAO index values given in Hurrell (1995).

The model experiments contain only 6 complete winters (DJF). This is clearly too short to simulate the inter-decadal variability of the NAO. The mean winter MSLP field of the climate model compares reasonably well with the climate of the ECMWF analyses. However, the MSLP field in the control experiment (Fig. 1c) compares better with the ECMWF analyses for the 1991–1995 period (Fig. 1a) than with the analyses for the 1986–1990 period or with the 1955–1993 winter mean of the DNMI pressure fields (not shown). This means that the mean winter MSLP field of the control experiment seems to correspond to the phase of the NAO with a strong zonal flow. The largest dissimilarity between the control experiment and the analyses for the 1991–1995 period is that the Azores high is slightly more intense and is located over the Iberian Peninsula in the

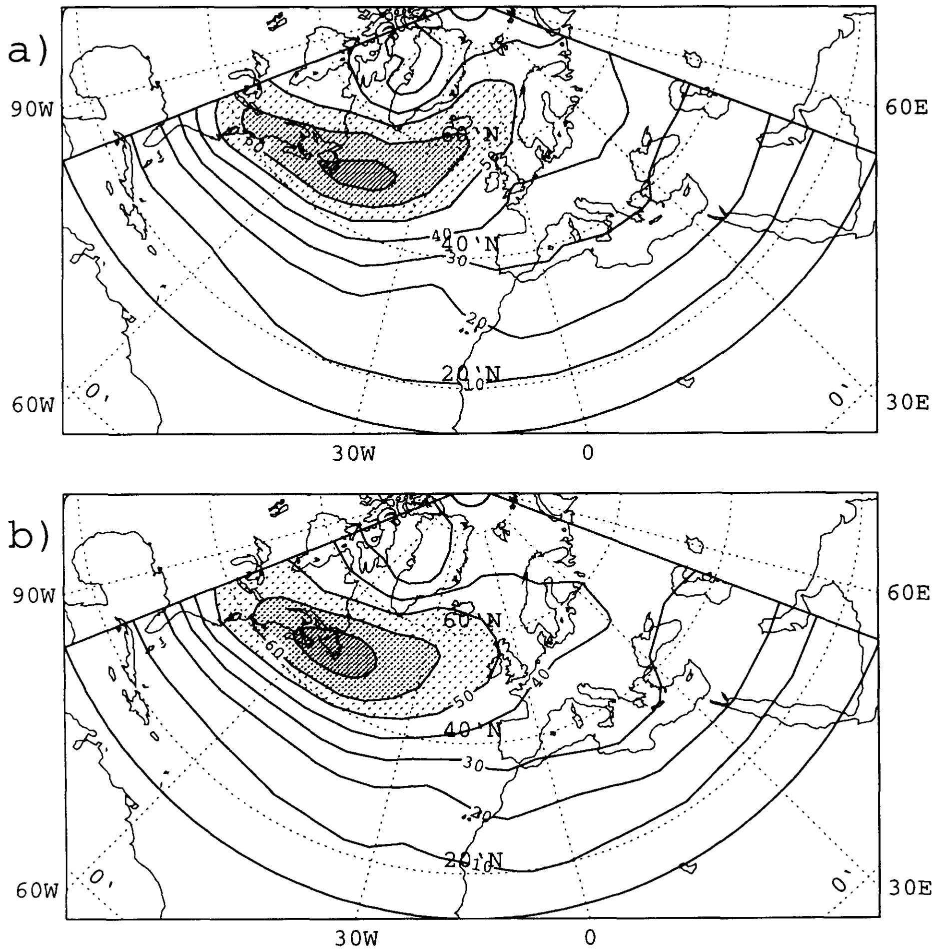

Fig. 2. Standard deviation of the band-pass (2.5–6 days) filtered 500 hPa height (DJF). (a) ECMWF analyses: 1991–1995; (b) ECMWF analyses: 1986–1990; (c) ECHAM3/T106: control experiment (6 winters); (d) ECHAM3/T106: $2 \times CO_2$ experiment (6 winters). Values over 50 m are shaded.

control experiment. It should be noted, however, that the SST's in the model control experiment were taken from the period 1979–1988, when the average NAO index is somewhat smaller than in the 1991–1995 period.

3.2. *500 hPa height variability*

A traditional measure of baroclinic activity is the time filtered variability of the 500 hPa height field (Blackmon, 1976). Figs. 2a, b show the standard deviation of the band-pass (2.5–6 days) filtered 500 hPa height fields from ECMWF analyses for the winters (DJF) of the years 1991–1995 and 1986–1990 respectively. The two 5-year periods which roughly represent the two modes of the NAO show an associated effect with respect to the intensity and location of the North Atlantic storm track. In the most recent period the highest intensity is somewhat smaller but the storm track extends much further into the Norwegian Sea. A similar relation between the NAO and the North Atlantic storm track was found by Rogers (1990).

The North Atlantic storm track in the 6 winters of the control experiment is given in Fig. 2c. The simulated storm track compares quite well with the storm tracks in the ECMWF analysis. The east-side of the storm track seems to consist of two branches: a northern branch with southwest-northeast direction extending into the Norwegian

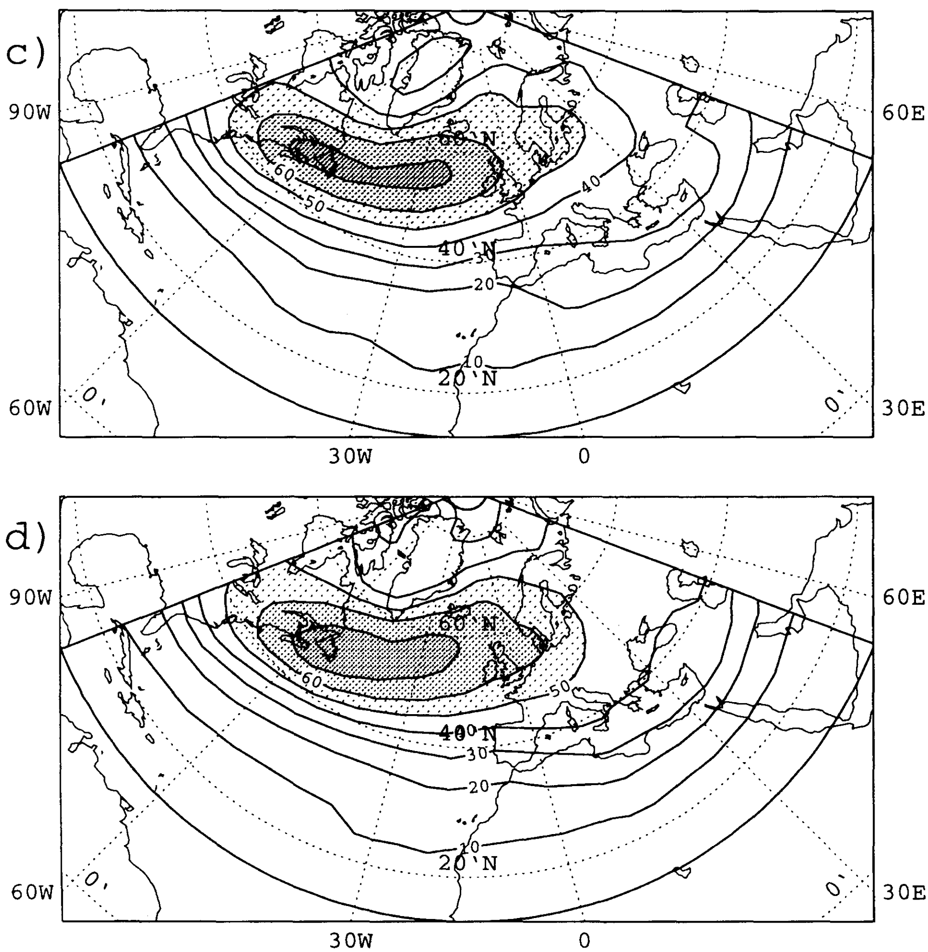

Fig. 2 (cont'd).

Sea and a southern branch with a west-east orientation. The area of maximum intensity extends further eastwards than for either of the two periods of ECMWF analyses. It should be noted that earlier versions of the ECHAM model with lower resolutions were not able to produce maximum values over 70 m (von Storch et al., 1993).

3.3. *Surface winds*

In our analysis of the 10 m wind we distinguish average wind speed and extreme wind speed. The extreme wind speed is calculated as the 95th percentile of twice daily wind speeds. The fields of 95th percentile for the individual winters are averaged to obtain a spatially smoothed field. The fields of the average wind speed and the extreme wind speed in the control experiment compare fairly well with the ECMWF analyses and are therefore not shown. Both fields compare slightly better with the ECMWF analyses of the winters 1991–1995 than those of the winters 1986–1990. The major deviations are described briefly below. Compared to the ECMWF analyses, the control experiment underestimates average wind speeds and extreme wind speeds in the Atlantic between 30° and 50°N respectively 5–15% and 10–20%. Average wind speeds are overestimated 5–15% along the British and Scandinavian coasts.

The differences in both the average wind speed and the 95th percentile associated with the NAO are largest (up to 15%) along the Scandinavian coast. Consistent with the larger pressure gradient between Iceland and the Azores, the winters of 1991–1995 show a 1–2 m/s higher average 10 m wind speed over the Atlantic between south Greenland and Scandinavia than the winters of 1986–1990. Further, a decrease of the extreme wind speeds near the Azores is observed when the NAO phase is high (1991–1995) which is in agreement with the intensification of the Azorian high. It should be noted, however, that in 1991 the boundary layer scheme of the ECMWF model was changed resulting in a reduction of the wind speed.

3.4. *Frequency of storms*

The simulation of storms is examined using two different approaches. In the first approach, the number of depressions with different intensities, as measured by their core pressure, is determined over all 12-h time steps for the winter seasons. In this way the duration of a storm is taken into account, as well as the number of storms. In the second approach the number of storms is examined by counting for each grid point the number of storm events. A storm event (or run event) is defined as a series of consecutive wind speed values exceeding a certain threshold. Since storms are generally separated by a period of calmer weather a storm event is regarded as an appropriate measure of a physical storm when a large enough threshold is applied. A threshold of 15 m/s has been used which corresponds with wind force 7 on the Beaufort scale and which is described as near gale.

Table 1 shows the average number of depressions per winter, as obtained by the first method, for both model experiments and for the two 5-year periods of ECMWF analyses. The numbers in the table refer to all depressions with a certain intensity in the North Atlantic area. The table shows that the climate model is able to realistically simulate the frequency of depressions with a core pressure between 960 and 975 hPa in the whole area. The number of deep depressions (with a core pressure less than 960 hPa) is even overestimated

Table 1. *Frequency of depressions (per winter) grouped by their core pressure in the whole North Atlantic for the ECMWF analyses, the control experiment and the $2\times CO_2$ experiment*

Core pressure of depression (P(hPa))	ECMWF 1986–1990	ECMWF 1991–1995	Control	$2\times CO_2$
$975<P<990$	304	296	313	366
$960<P<975$	110	117	118	117
$945<P<960$	42	39	52	43
$P<945$	6	6	9	10
$P<990$	462	458	492	536

in the model by about 30%. During 5 simulated winters one depression occurred with a core pressure less than 930 hPa while 4 of these depressions occurred in the 1991–1995 ECMWF analyses. This result would be difficult to achieve with a lower resolution model.

A similar analysis was carried out with the DNMI data. The analysis of cyclone frequency in this case was limited to the north-western North Atlantic area (50–70°N, 20°W–20°E) including the Irish Sea, the North Sea and the southern part of Norwegian Sea, hereafter called NWNA area. In this region the quality of the DNMI data seems to be better than in other areas through the whole time period 1955–1993. Table 2 summarizes the cyclone statistics in the NWNA area for the DNMI data, separately for each decade, and for the GCM experiments. The intensity of the storms is evaluated in terms of their core pressure as well as the average geostrophical wind around the core of the lows, obtained from the average pressure gradient around the core. The number of detected cyclones and their intensity distributions in the DNMI data and the control experiment are comparable. The number of deep depressions is a little larger in the control experiment but the difference is much smaller than for the whole area. A slightly increased number of deep ($P<975$ hPa) and strong ($U>15$ m/s) depressions in the last 10 years of the DNMI data set, is accompanied by a much stronger zonal flow in this period compared with the previous 10-year periods. The pressure gradient in the 1984–1993 DNMI data is of similar magnitude as in the 1991–1995 ECMWF analyses.

Fig. 3a presents the number of storm events in the ECMWF analyses of the 1991–1995 winters as calculated from the second method. As expected, there are very few exceedances of 15 m/s over land, due to stronger surface friction over land. The effect of the NAO variability on the number of storm events is given in Fig. 3b. The largest differences in the number of storm events are found between south Greenland and Scandinavia. In this area there are up to 8 storm events per winter more in the 1991–1995 period than in the 1986–1990 period which is in agreement with the shift of the storm track and the stronger mean zonal flow. The maximum number of storm events over the North Atlantic in the model control experiment (Fig. 3c) is smaller than the corresponding maximum number in each of the analysed periods.

Comparing results from both approaches an interesting paradox appears. Using the first approach we have seen that the control run overestimates the number of deep depressions, but in the second approach we obtained an underestimate of the number of storm events. To explain this apparent discrepancy we have considered a number of possibilities. First of all we have investigated whether there was a significant difference between the ECHAM3 control run and the ECMWF analyses with respect to the duration of the storm events. This was not the case. Next, we determined the pressure gradients associated with deep depressions both for the ECMWF analyses and the control run. Again no systematic difference was obtained. From this, one then is led to infer

Table 2. *Frequency of depressions (per winter) grouped by their core pressure and by the mean geostrophical wind around the core in the north-western part of the North Atlantic (50–70°N, 20°W–20°E) for the DNMI data, separately for each decade, the control experiment and the $2\times CO_2$ experiment*

Core pressure of depression (P(hPa))	DNMI 1955–1963	DNMI 1964–1973	DNMI 1974–1983	DNMI 1984–1993	Control	$2\times CO_2$
$990<P$	47	49	49	47	43	52
$975<P<990$	62	64	69	66	73	67
$P<975$	54	52	49	58	55	50
Mean geostrophic wind (U(m/s))						
$U<7.5$	21	24	21	20	19	20
$7.5<U<15$	78	77	79	76	76	77
$15<U$	64	64	67	75	77	72
total	163	165	167	171	172	169

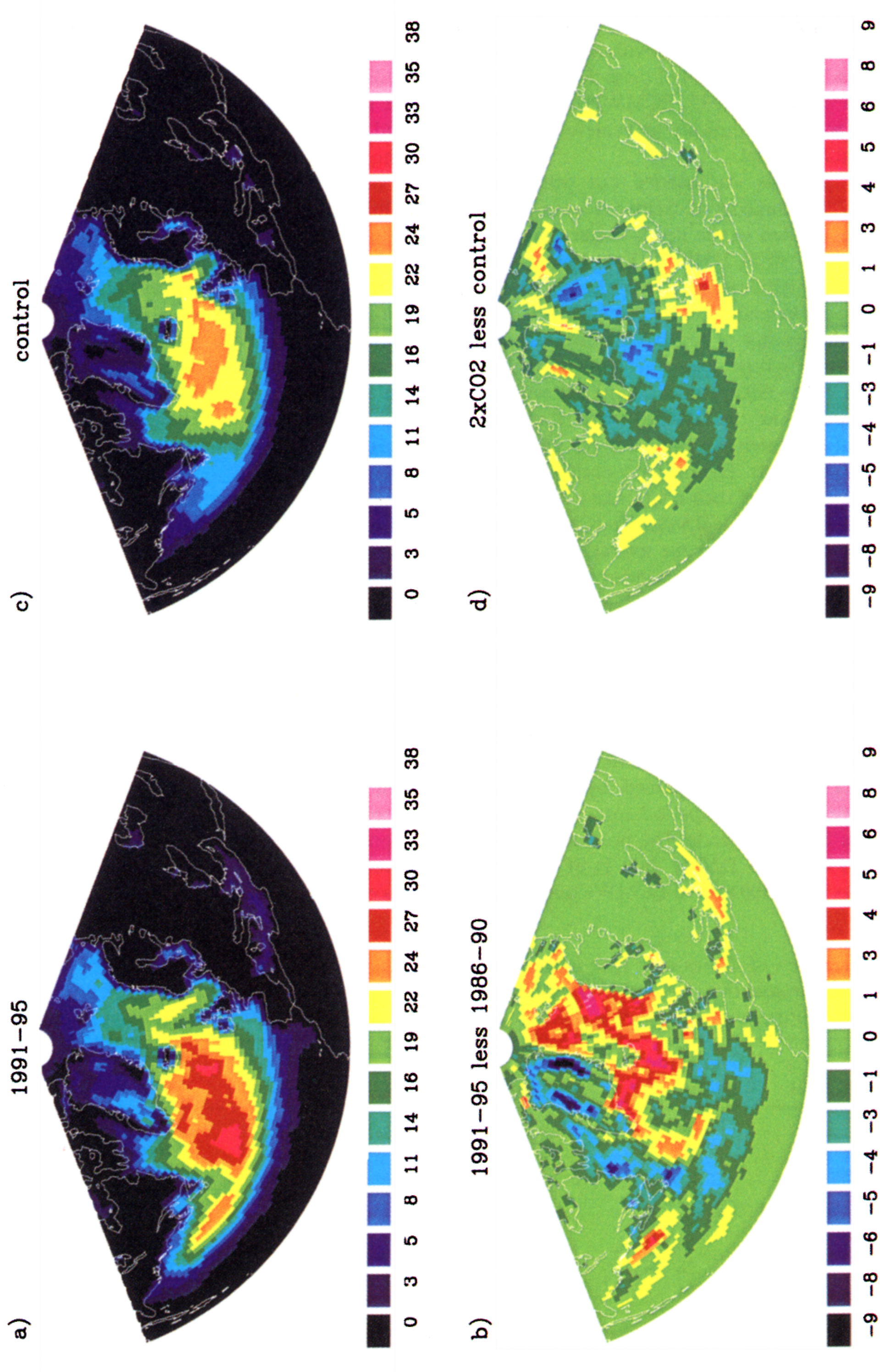

Fig. 3. Number of storm events per winter (see text for description). (a) ECMWF analyses: 1991–1995; (b) ECMWF analyses: 1991–1995 less 1986–1990; (c) ECHAM3/T106: control experiment (6 winters); (d) ECHAM3/T106: $2\times CO_2$ — control (6 winters). The results have been smoothed in space with a moving 4 grid point window.

that ECHAM3 and ECMWF compute different winds for given pressure gradients, which suggests that the discrepancy is caused by differences in the physics of both models. One such difference is in the way the horizontal diffusion is parametrized. Another difference is the value of the Charnock parameter, which is 0.018 in the ECMWF model but was taken as 0.032 in ECHAM3 (DKRZ, 1992). This high unphysical value of the Charnock constant may well be responsible for the reduced number of storm events in the control run. It would also account for the underestimation of extreme winds noted above. However, it is difficult to settle this matter without performing additional model runs, which is outside the scope of this paper.

In summary, we have seen that the ECHAM3/T106 control simulation is basically realistic in terms of mean pressure distribution, variability of the 500 hPa height, average wind speed and storm frequency. Deviations have been described above. In general, the decadal variability of the ECMWF analyses is larger than the differences between the control simulation and the analysis. The particular control run that has been made is closest to the ECMWF analysis 1991–1995. Extreme winds and the number of storm events seem somewhat low in the control simulation, but this may be related to the high value of the Charnock constant used in in this version of ECHAM3.

4. The climate of the $2 \times CO_2$ experiment

The aim of this paper is not only to establish whether the control experiment is realistic, but also whether the climate produced by the $2 \times CO_2$ experiment is different from that produced by the control experiment. This section is therefore devoted to an analysis of the wind and storm climate in the North Atlantic in the $2 \times CO_2$ experiment. Again the focus is on the MSLP, the 500 hPa geopotential height variability and the surface winds using the same diagnostic evaluation as for the control run.

4.1. Mean sea level pressure

In the $2 \times CO_2$ experiment the north-south pressure gradient (not shown) is of comparable size as in the control experiment which means that also the $2 \times CO_2$ experiment appears to correspond with a high phase of the NAO. Fig. 1d illustrates the difference in the average MSLP between the $2 \times CO_2$ experiment and the control experiment. The anomaly pattern shows a dipole with negative differences of up to 6 hPa over Scandinavia and positive differences of up to 3 hPa over the central North Atlantic. It is associated with a slight extension to the west of the Azores high and an extension to the east of the Icelandic low. The pattern looks almost orthogonal to the one seen in Fig. 1b, showing the difference of the average MSLP for two different 5-year periods of ECMWF analyses, which is associated with an increase in the zonal index. When t-tests are applied to the MSLP differences in both 5-year ECMWF analyses and to the differences between the $2 \times CO_2$ and control experiment the percentage of significant grid points is only half of the significance level for the ECMWF anomalies and twice the significance level for the $2 \times CO_2$ – control anomalies. In the latter case the significant grid points occur mainly in two areas south of 30°N where the interannual variability of the MSLP is small. From this it may be concluded that the MSLP pattern in the $2 \times CO_2$ experiment does not markedly differ from that in the control experiment. In other words, it is unlikely that a similar signal would be reproduced in the model integrations made with slightly different initial conditions.

4.2. Variability of the 500 hPa height

Fig. 2d shows the time-filtered variability of the 500 hPa height field from the $2 \times CO_2$ experiment for the 6 winters available. Compared with that of the control experiment (Fig. 2c) the area of the storm track appears to have become slightly more intense, with standard deviations of 75–80 m over Newfoundland. Furthermore, the storm track is somewhat extended towards the east. The largest increases of the 500 hPa variability are found over the North Sea, the Bay of Biscay and central Europe. For example, over the North Sea the standard deviations of the filtered 500 hPa field are 55–60 m in the control experiment, but in the $2 \times CO_2$ experiment the values are 60–65 m which is an increase of almost 10%. The slight increase of the 500 hPa variability is likely related to the enhanced pole-to-equator temperature gradient in the upper troposphere causing in turn a slightly increased subtropical jet in the double CO_2 scenario (Bengtsson et al., 1996).

4.3. Surface winds

From geostrophic balance a pressure (gradient) anomaly as in Fig. 1d should be accompanied by a wind speed anomaly. The pressure gradient anomaly in Fig. 1d, however, is almost orthogonal to the climatological pressure gradient over the British Isles and therefore its effect on the average wind speed in this area is small. In fact, the differences in the average 10 m wind speed and the 95th percentile over the British Isles are less than about 5% (not shown). Over central Europe, where the anomalous pressure gradient adds to the climatological gradient, increases of 10–15% are found for both average and extreme wind speeds. The latter increase is in agreement with the increase in the 500 hPa height variability.

A clearer signal over the British Isles is found in the changes in the wind *direction*. Fig. 4 shows the distribution of wind directions for 5 winter seasons for an area North of Scotland (60.0–64.5°N, 9.0°W–0.0°). The figure shows that in general there is a clockwise shift in the distribution of the wind directions. In the control experiment, the dominant direction is SW to W, while the peak in the $2 \times CO_2$ experiment is shifted somewhat to a more westerly direction. When the winds are categorized according to wind strength, this trend for the directions to veer is present in all categories (not shown).

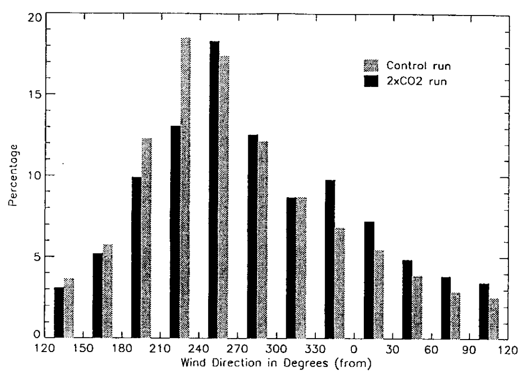

Fig. 4. Distribution of wind directions (DJF) for an area north of Scotland (60.0–64.5°N, 9.0°W–0.0°). Control experiment and $2 \times CO_2$ experiment.

4.4. Frequency of storms

The frequency of storms is examined with the same two approaches that were used in Section 3 to analyse the control run. In the first approach the number of depressions was counted. The results are given in Table 1 for all of the (extra-tropical) North Atlantic and in Table 2 for the NWNA area. In both cases the number of depressions with a core pressure below 975 hPa is lower in the $2 \times CO_2$ scenario than in the control simulation. In the NWNA area the number of very weak depressions ($P > 990$ hPa) increases while there is a small decrease of the total number of depressions. In addition there is a change towards smaller geostrophic winds (Table 2). In the whole North Atlantic area the number of weak depressions, with a core pressure between 975 and 990 hPa decreases by 17%. These results suggest a shift towards less developed depressions in the $2 \times CO_2$ scenario.

In the second approach the spatial distribution of the number of storm events is determined. The results are presented in Fig. 3d which gives the difference between the number of storm events in the $2 \times CO_2$ experiment and in the control experiment. This figure shows that the number of storm events decreases in the $2 \times CO_2$ experiment over much of the North Atlantic area. A considerable number of the deep depressions identified in Table 1 occur around the southern tip of Greenland and to the west of Iceland. It is also in this area that the largest decrease is observed, namely up to 6 storm events per winter (Fig. 3d). Over the North Sea and the Bay of Biscay there is an increase. In all cases the changes are relatively small, in particular when compared with the difference between the two ECMWF analyses periods which result from the NAO (low frequency) variability (Fig. 3b).

From the combined results it can be concluded that there is a shift in the distribution of depression strength towards weaker depressions. As a result the number of storm events, which is directly related to the strength of the depressions, decreases in most areas. The increase of the number of storm events over the North Sea and the Bay of Biscay is in agreement with the enhanced 500 hPa height variability in these areas. The changes in these two small areas are obviously masked in the analysis of the depression frequencies because the areas used for the analysis of the depression frequencies are much larger.

5. Discussion and conclusions

We have analysed the time-slice experiments with the ECHAM3 model, at a horizontal resolution of T106. These are among the best high resolution climate model data available for studying intense developments in a greenhouse-gas induced climate. An important caveat of these simulations is however their length. It is hardly possible to make stable estimates about the changes in extremes from simulations of only 5-years. Observations show that variability on interannual to decadal time scales plays an important role in the North Atlantic region. Due to the presence of this low-frequency variability, simulations with a high resolution fully coupled ocean-atmosphere model covering at least a few decades are required to make meaningful inferences about the effects of enhanced greenhouse gas concentrations. This has so far not been possible from the point of view of computational costs.

The control experiment was compared with the operational ECMWF analyses from the 10-year period 1986–1995 and with a set analyses for the period 1955–1993 prepared by DNMI. A striking feature in the observations is the low frequency variability. This variability is associated in part with the North Atlantic Oscillation, but one can not exclude that inhomogeneities in the preparation of the analyses also contribute. It will therefore be interesting to repeat the analysis when more homogeneous data become available. In any case, the control simulation compares best with observed data in the period 1991–1995, a period characterised by a strong mean zonal flow over the North Atlantic. The simulation was found to have many realistic characteristics. In particular, the variability of the 500 hPa height (with standard deviations exceeding 70 m) and the frequency of deep depressions were simulated well. This result, which would be difficult to achieve with a lower resolution model, justifies the use of the time-slice runs for an analysis of the extra-tropical storm climatology.

When comparing the $2 \times CO_2$ experiment and the control experiment our main conclusion is that differences are small. The $2 \times CO_2$ experiment has, relative to the control experiment, a negative pressure anomaly of 5 hPa over Scandinavia and a weaker positive anomaly over the central North Atlantic. Associated with this pressure anomaly pattern there is a veering of the mean wind direction over the eastern North Atlantic. Also, there is a slight intensification of the variability of the 500 hPa height, which is most pronounced in the Bay of Biscay, the North Sea and over central Europe. The frequency of deep depressions decreases slightly while the frequency of weak depressions increases. This corresponds with a spatial pattern in which the number of mid-latitude storm events decreases somewhat in a large part of the North Atlantic, but increases in the North Sea and the Bay of Biscay. However, it is not possible to conclude that these differences are the result of the CO_2 doubling because they are small with respect to the natural variability.

The absence of a strong $2 \times CO_2$ doubling signal in the extra-tropics contrasts with the results of Bengtsson et al. (1996). These authors found a significant reduction of the number of tropical storms (though not of their strength) in the $2 \times CO_2$ run compared to the control run. These two results are not at variance since the physical mechanism for the generation of extra-tropical storms is baroclinic instability, which is quite different from the mechanism responsible for the generation of tropical cyclones. One of the factors affecting baroclinic instability is the meridional temperature gradient. As discussed in Section 1, model simulations of the climate response to an increase in greenhouse gas concentration show the largest winter surface warming near the poles. The resulting decrease in the pole-to-equator temperature gradient could weaken the midlatitude baroclinic eddies. Most models, including ECHAM3, however, also predict an important competing change, i.e., an increase in the pole-to-equator temperature gradient in the upper troposphere. The numerical simulations analysed here suggest that the decrease in the lower tropospheric temperature gradient and the increase in upper tropospheric gradient balance approximately, in such a way that the net effect on mid-latitude storminess is small.

6. Acknowledgements

The authors would like to thank Professor L. Bengtsson, Max-Planck-Institute for Meteorology in Hamburg, for making the ECHAM3/T106 time-slice experiments available. The DNMI sea level pressure data were kindly provided by Dr. M. Reistad. The authors are grateful to Dr. H. von Storch for valuable discussions and suggestions made during the work. They also thank Drs. T. A. Buishand, A. Kattenberg and G. P. Können and an anonymous reviewer for constructive comments. This work is part of the EU sponsored WASA project (Grant EV5V-CT94-0506).

REFERENCES

Bacon, S. and Carter, D. J. T. 1991. Wave climate changes in the North Atlantic and North Sea. *Int. J. Climatol.* **11**, 545–558.

Beersma, J. J. 1994. *Storm activity over the North Sea and the Netherlands in two climate models compared with observations.* KNMI Scientific report WR 94-02, 43 pp., Royal Netherlands Meteorological Institute (KNMI), De Bilt, The Netherlands.

Bengtsson, L., Botzet, M. and Esch, M. 1995. Hurricane-type vortices in a general circulation model. *Tellus* **47A**, 175–196.

Bengtsson, L., Botzet, M. and Esch, M. 1996. Will greenhouse gas-induced warming over the next 50 years lead to higher frequency and greater intensity of hurricanes? *Tellus* **48A**, 57–73.

Blackmon, M. L. 1976. A climatological spectral study of the 500mb geopotential height of the Northern Hemisphere. *J. Atmos. Sci.* **33**, 1607–1623.

Bouws, E., Jannink, D. and Komen, G. J. 1996. The increasing wave height in the North Atlantic Ocean. *Bull. Amer. Meteor. Soc.* **10**, 2275–2277.

Brinkop, S. 1991. Inclusion of cloud processes in the ECHAM PBL parameterization. In: *Studying climate with the ECHAM atmospheric model. Large-Scale atmospheric modelling*, R. Sausen (ed.). Report No. 9, 5–14. Meteorologisches Institut der Universität Hamburg, Germany.

Brinkop, S. 1992. Parameterisierung von Grenzschichtwolken für Zirkulationsmodelle. *Berichte aus dem Zentrum für Meeres- und Klimaforschung*, Reihe A. Meteorologie, nr. 2. 77pp., Meteorologisches Institut der Universität Hamburg, Germany.

Cubasch, U., Hasselmann, K., Hoeck, H., Maier-Reimer, E., Mikolajewicz, U., Santer, B. D. and Sausen, R. 1992. Time-dependent greenhouse warming computations with a coupled ocean-atmosphere model. *Climate Dynamics* **8**, 55–69.

Deser, C. and Blackmon, M. L. 1993. Surface climate variations over the North Atlantic Ocean during winter: 1900–1989. *J. Climate* **6**, 1743–1753.

DKRZ (ed.) 1992: *The ECHAM 3 atmospheric general circulation model.* DKRZ Technical Report No.6, 184 pp., Deutsches Klimarechenzentrum (DKRZ), Hamburg, Germany.

Dronia, H. 1991. On the accumulation of excessive low pressure systems over the North Atlantic during the winter seasons November to March) 1988/89 to 1990/91. *Die Witterung im Übersee* **39**, 27.

Held, I. M. 1993. Large-scale dynamics and global warming. *Bull. Amer. Meteor. Soc.* **74**, 228–241.

Hurrell, J. W. 1995. Decadal trends in the North Atlantic Oscillation: Regional temperatures and precipitation. *Science* **269**, 676–679.

Hurrell, J. W. and van Loon, H. 1995. Analysis of low-frequency climate variations over the Northern Atlantic using historical atmospheric data. Preprints *Sixth Symp. on Global change studies*, Dallas, TX, Amer. Meteor. Soc., 174–179.

Kattenberg, A., Giorgi, F., Grassl, H., Meehl, G. A., Mitchell, J. F. B., Stouffer, R. J., Tokioka, T., Weaver, A. J. and Wigley, T. M. L. 1996. Climate models: projections of future climate change. *Climate change 1995, The science of climate change*, Houghton, J. T., Meira Filho, L. G., Callander, B. A., Harris, N., Kattenberg, A. and Maskell, K (eds.): 572 pp., Cambridge University Press, Cambridge, UK.

Kaas, E., Li, T.-S. and Schmith, T. 1996. Statistical hindcast of wind climatology in the North Atlantic and northwestern European region. *Climate Research* **7**, 97–110.

Louis, J. F. 1979. A parametric model of vertical eddy fluxes in the atmosphere. *Bound.-Layer Meteor.* **17**, 187–202.

Miller, M. J., Palmer T. N. and Swinbank, R. 1989. Parameterization and influence of sub-grid scale orography in general circulation and numerical weather prediction models. *Met. Atm. Phys.* **40**, 84–109.

Miller, M. J., Beljaars A. and Palmer, T. N. 1992. The sensitivity of the ECMWF model to the parameterization of evaporation in the tropical oceans. *J. Climate*, **5**, 418–434.

Nicholls, N., Gruza, G. V., Jouzel, J., Karl, T. R., Ogallo L. A. and Parker, D. E. 1996. Observed climate variability and change. *Climate change 1995, The science of climate change*, Houghton, J. T., Meira Filho, L. G., Callander, B. A., Harris, N., Kattenberg, A. and Maskell, K (eds.): 572 pp., Cambridge University Press, Cambridge, UK.

Palmer, T. N., Shutts G. J. and Swinbank, R. 1986. Alleviation of a systematic westerly bias in general circulation and numerical weather prediction models through an orographic gravity wave drag parameterization. *Quart. J. Roy. Meteor. Soc.* **112**, 1001–1031.

Rogers, J. C. 1990. Patterns of low-frequency monthly sea level pressure variability (1899–1986) and associated wave cyclone frequencies. *J. Climate* **3**, 1364–1379.

Schinke, H. 1993. On the occurence of deep cyclones over Europe and the North Atlantic in the period 1930–1991. *Beitr. Phys. Atmosph.* **66**, 223–237.

Schmidt, H. and von Storch, H. 1993. German Bight storms analysed. *Nature* **365**, 791.

Stein, O and Hense, A. 1994. A reconstructed time series of the number of extreme low pressure events since 1880. *Meteorologische Zeitschrift* **N.F.3**, 43–46.

Von Storch, H., Guddal, J., Iden, K. A., Jonsson, T., Perlwitz, J., Reistad, M., de Ronde, J., Schmidt, H. and Zorita, E. 1993. *Changing statistics of storms in the North Atlantic?* MPI Report No. 116, 19 pp. Max-Planck-Institut für Meteorologie, Hamburg, Germany.

Wallace, J. M. and Gutzler, D. S. 1981. Teleconnections in the geopotential height field during the northern hemisphere winter. *Mon. Wea. Rev.* **109**, 784–812.