PRECIPITATION: A PARAMETER CHANGING CLIMATE AND MODIFIED BY CLIMATE CHANGE

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Abstract. This paper discusses two aspects of climate modeling, the deep water formation in the North Atlantic and precipitation changes due to climate change caused by anthropogenic emissions of greenhouse gases. The deep water formation is strongly influenced by the precipitation, and the precipitation is affected by the concentration of the greenhouse gases in the atmosphere and by the atmospheric and oceanic circulation. The experiments discussed here have been performed independently to test the stability of the thermohaline circulation of the North Atlantic and to investigate changes in precipitation due to anthropogenic greenhouse gas emissions. The precipitation changes in a climate change environment are sufficient in some simulations to decrease the thermohaline circulation noticeably. However, it appears that the amount of freshwater needed to bring the circulation to a collapse is magnitudes larger than the anticipated change in precipitation due to anthropogenic activities within the next 100 years. The precipitation changes, on the other hand, might change regionally quite drastically towards more extreme situations, thereby putting additional stress on vegetation and enhancing soil erosion.

1. Introduction

In the past, the groups working on the stability of the thermohaline circulation and those dealing with the issue of anthropogenic global change have been working in parallel and quite independently from each other for several reasons. One reason is that the first topic has been considered the concern of oceanographers, the other an atmospheric science issue. Secondly, the time scales involved were considered to be at least one order of magnitude apart, i. e. decades to one century for anthropogenic climate change, and centuries to millennia for the North Atlantic circulation. In recent years, a number of studies indicated that anthropogenic global change is quite likely to influence the thermohaline circulation (THC) (Manabe et al., 1990; Cubasch et al, 1992). At the same time, observational evidence has been published indicating that the THC in the North Atlantic region undergoes a number of fluctuations even on decadal time scales (see e.g., Dickson et al., 1996 for a review). The aim of the present paper is firstly to demonstrate the current state of knowledge of the stability of the THC using one example, secondly to describe the precipitation change under anthropogenically changed climate conditions, and finally to discuss the potential of both parameters to be represented by impulse response function and their ability to serve as indicators for critical states of the climate system.

Climatic Change **46:** 257–276, 2000.

2. The Stability of the North Atlantic Deep Water Formation

Geological records show some strong climatic events with large temperature changes happening over a period of a few decades. The most prominent example is the Younger Dryas cold event during the last deglaciation about 13,000 years ago, when Europe and the North Atlantic experienced a sudden return to almost glacial conditions, which lasted for about 1300 years (Mayewski et al., 1993).

It has often been speculated (e.g., Rooth, 1982) that changes in the THC of the ocean and associated changes in oceanic heat transport might be responsible for these large climate changes. The speculations are built on the fact that, at least in certain parameter regimes, ocean models can have multiple steady states of the thermohaline circulation, particularly in the North Atlantic (Stommel, 1961; Bryan, 1986; Marotzke and Willebrand, 1991; Manabe and Stouffer, 1988; Mikolajewicz et al., 1993; Hughes and Weaver, 1994; Rahmstorf and Willebrand, 1995).

Maier-Reimer and Mikolajewicz (1989) showed with an ocean model, in which the atmospheric forcing was represented only by a relaxation to observed temperature and by a prescribed fresh water flux (so called “mixed boundary conditions”) that the addition of small amounts of meltwater to the North Atlantic was sufficient to cause a transition from the present day (conveyor-) mode of ocean circulation with strong formation of North Atlantic Deep Water (NADW) to another mode without NADW formation. Furthermore, they showed that the stability of the THC strongly depended on the location of the freshwater input, thus supporting the hypothesis that the Younger Dryas was caused by meltwater input from the Laurentide ice sheet (Rooth, 1982; Broecker et al., 1985). The unrealistically weak stability of the conveyor mode of the thermohaline circulation in ocean general circulation models was found recently to be an artifact of the use of “mixed boundary conditions” (Zhang et al., 1993; Mikolajewicz and Maier-Reimer, 1994; Rahmstorf and Willebrand, 1995).

3. The Sensitivity to Freshwater Input

Since a comprehensive review of all methods analyzing the stability of the THC would exceed the scope of this paper, the principle of these studies is demonstrated on a single example. The experiments were performed with the ocean-atmosphere general circulation model (GCM) ECHAM3/LSG. The atmospheric component consists of the general circulation model ECHAM3 (Roeckner et al., 1992) in a T21 resolution. Vertically, the model is discretized in 19 levels. The ocean general circulation model LSG (Maier-Reimer et al., 1993) has 11 layers and uses a similar grid spacing as the Gaussian grid of the atmosphere model. Both models are coupled periodically synchronously (Voss and Sausen, 1996; Voss et al., 1997).

In the experiment EXP, the meltwater input to the Labrador Sea is increased linearly for 250 years to a maximum input of 0.625 Sv and then reduced over

250 years again to 0 Sv (Figure 1a). The results are compared to a corresponding control simulation without additional freshwater input (CTRL).

During the first 250 years, the increasing freshening forces a shutdown of the formation of NADW and a subsequent reversal of the thermohaline circulation of the Atlantic, filling the deep Atlantic with Antarctic Bottom Water (AABW) (Figures 1a, 1b).

The reduced flow of surface water into the Atlantic as part of the conveyor belt circulation also leads to strong changes in the surface velocity field (Figure 2). The most pronounced differences are evident in the Atlantic. In experiment CTRL, the main Gulf Stream flow penetrates northward until it reaches the GIN (Greenland-Iceland-Norway) Sea. This path is completely interrupted in the experiment with increased freshwater input. Instead, a weak Gulf Stream separates further south of the American coast and extends only to an area south of about 40°N. An intensified East Greenland Current supplies the water that flows southward along the European west coast and finally constitutes a strengthened North Equatorial Current.

Once the conveyor-belt overturning circulation has been interrupted, the Atlantic northward heat transport at 30°N is reduced from 0.7 PW to 0.1 PW. With the reduced northward oceanic heat transport, the sea surface temperature (SST) drops to the freezing point and sea-ice forms, isolating the ocean surface from the atmosphere. The SSTs of the whole North Atlantic are colder than in CTRL (over 8.5°C in the annual mean around Iceland and in the Norwegian Sea, see Figure 3). Consequently, the maximum in sea-ice extent now touches the American coast south of Newfoundland and even covers parts of the coasts of the British Isles. The Norwegian Sea is completely ice-covered. Another remarkable effect is the cooling over the North Pacific, which is accompanied by enhanced ventilation of intermediate depth water masses along the American coast (Mikolajewicz et al., 1997).

Four major feedbacks that influence NADW formation, but act in opposite directions, and thus partly cancel each other out, can be identified in this experiment:

- The changes in the overturning lead to reduced advection of saline, low-latitude surface water and significantly longer residence times of the surface waters in the northern North Atlantic and Arctic. The net precipitation in these areas causes a further reduction of the surface salinity and thus further increases the vertical stability of the water column. This effect has first been noted by Stommel (1961).
- The collapse of the conveyor belt type thermohaline circulation reduces the northward heat transport of the Atlantic. The resulting colder SSTs in the North Atlantic increase the surface density and counteract the effect of the freshening. The reduced stability of the water column allows enhanced mixing with the underlying, more salty, water masses.
- The increased meridional temperature gradient over the Atlantic leads to increased cyclonic activity over the North Atlantic and increased northward

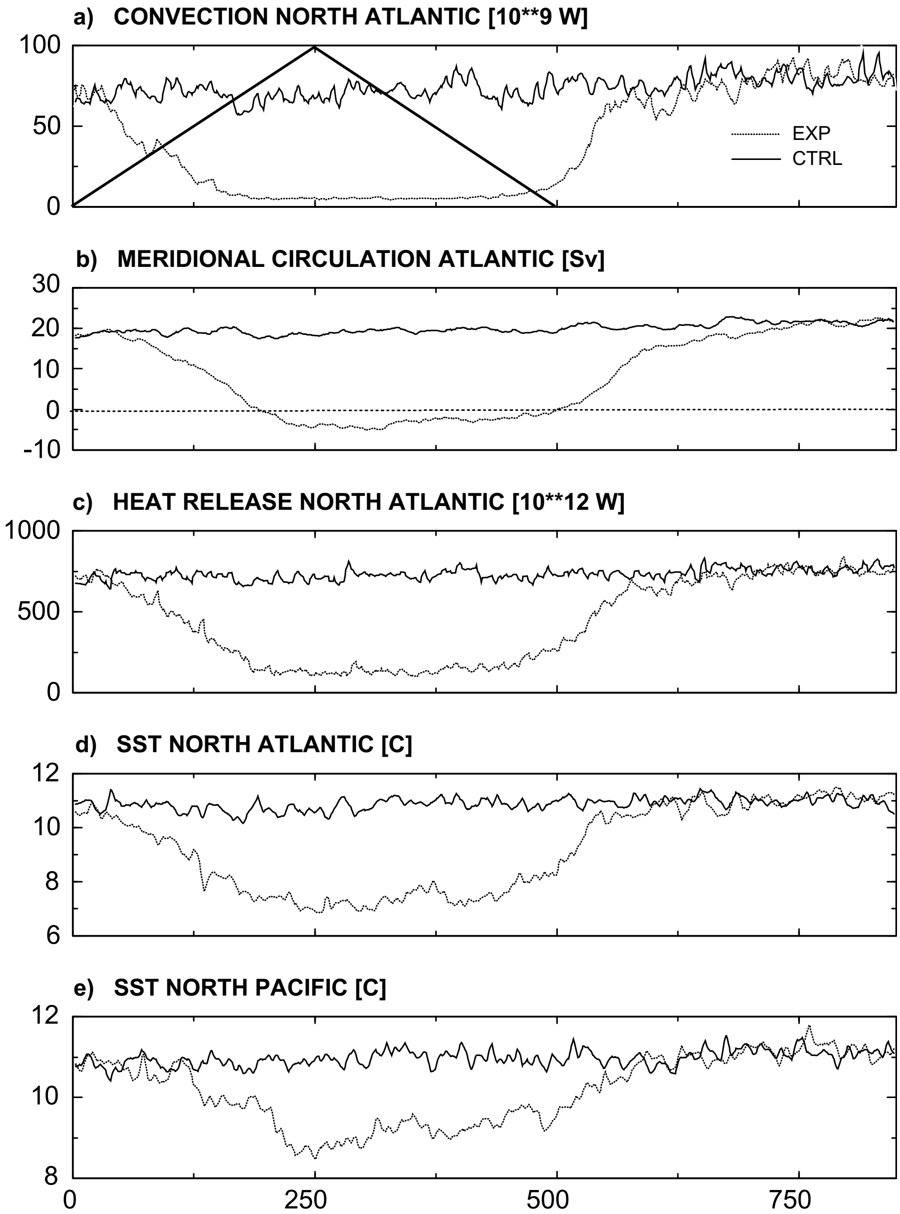

Figure 1. Time series of (a) prescribed meltwater plus freshwater input into the Arctic and North Atlantic north of 30°N (indicated schematically by the triangle) and the convection in the North Atlantic and Arctic, (b) meridional circulation ("overturning") at 30°S in the Atlantic (negative values indicate outflow of NADW to the Southern Ocean), (c) the heat release in the North Atlantic, (d) the annual mean North Atlantic SST, and in the North Pacific (e), averaged between 30 and 70°N. solid line: control experiment, dots: climate change experiment. All data are annual means without further filtering (after Schiller et al., 1996).

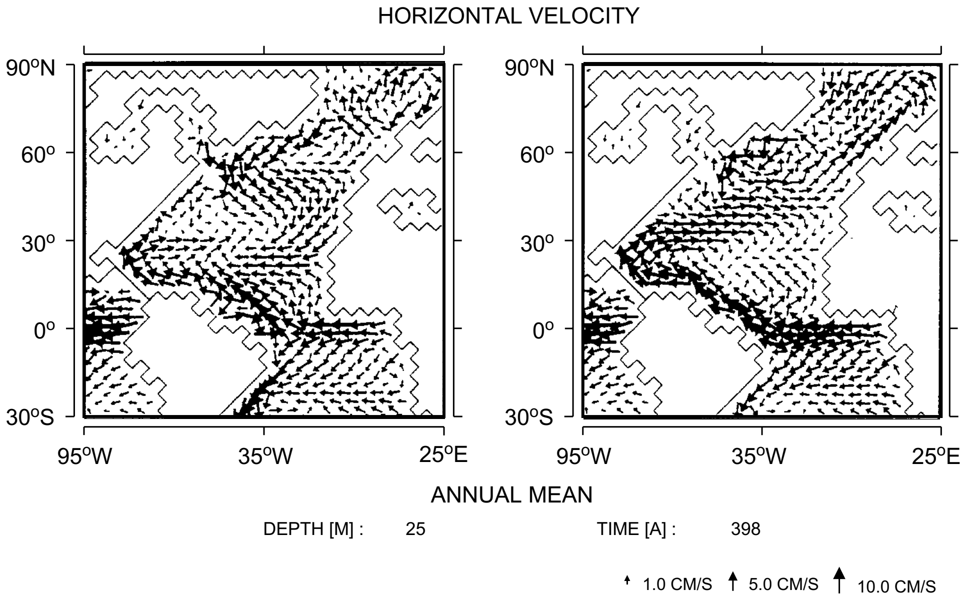

Figure 2. Surface currents in the Atlantic sector of the model. Average over the final decade of the freshwater input experiment (left) and the 25th decade of the control run (right). See lower right corner for scaling of velocity vectors (after Schiller et al., 1996).

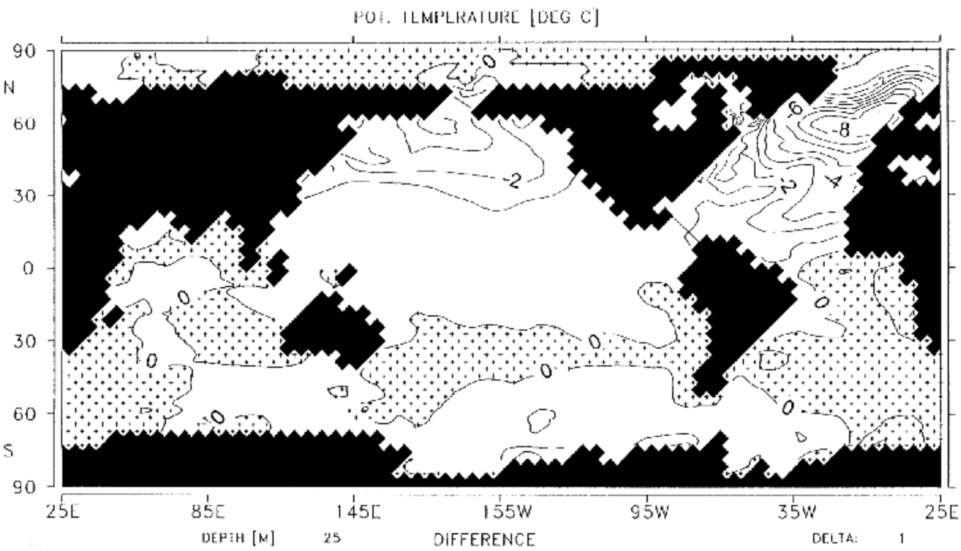

Figure 3. SST difference between the freshwater input and the control experiment, shading indicates positive values. C.I.: 1.0°C. Average over the 25th decade (years 241-250) of experiments (after Schiller et al., 1996).

atmospheric water vapor transport. This additional freshwater input to the North Atlantic dilutes the surface salinity and increases the stability of the water column (Nakamura et al., 1994).

- In the atmosphere, the deepening of the trough over the northern North Atlantic leads to an intensification of the cyclonic circulation in the subpolar gyre and intensified upwelling through Ekman-suction. Ekman transports lead to an intensified East Greenland Current that exports fresh surface water to the North Atlantic. This wind-driven feedback was found to be important for the breakdown of the stable stratification in the Norwegian Sea and the re-initiation of NADW formation after collapse of the conveyor belt circulation.

To investigate whether the sum of these compensating feedbacks allows a stable second steady state mode of the thermohaline circulation without NADW to exist, the experiment was continued, slowly reducing the meltwater input again to zero. After this, the integration was continued for 350 years without external perturbation. When the meltwater input was strongly reduced, the convection in the North Atlantic started to recover. It took about 100 years to reach a value almost as high as in the control run. The overturning at 1500 m depth at 30°S (an indicator of the outflow of NADW to the Southern Ocean) showed a matching, but somewhat delayed, response. The NADW formation and the NADW outflow have both been suppressed for more than 300 years (Figure 1a). The overall circulation at the end the experiment was clearly of the conveyor belt type (Figure 1b). The onset of NADW formation is accompanied by a rapid warming in the North Atlantic (Figure 1d).

The ECHAM3/LSG model shows a higher stability of the conveyor belt mode of the thermohaline circulation than the Manabe and Stouffer (1988) model (MS88) with annual mean insolation, and most ocean-only models. The difference in the behaviour can be explained by the weakness of stabilizing atmospheric feedbacks in the MS88 model or their total neglect in ocean-only models. The atmospheric response in the MS88 model is typically only one half of the response in our model. As the atmospheric feedbacks had a tendency to stabilize the conveyor belt circulation, this weaker atmospheric response could be responsible for the differences in stability. Additionally, the conveyor belt circulation in our model is about twice as strong as in the MS88 model. This, too, should definitely have a significant effect on its stability (compare with the simulations of Tziperman, 1997). Due to the non-linearity of the problem, the quantitative difference in the strength of single feedbacks leads to a qualitative difference in the response of the coupled system. Thus the conveyor belt circulation is definitely the preferred and probably also the only mode of thermohaline circulation in our OAGCM, in contrast to the results of MS88 and most ocean-only models.

Manabe and Stouffer (1995), with a newer version of their coupled OAGCM, also investigated the effect of meltwater input into the North Atlantic. They prescribed an input of 1 Sv over 10 years and continued the experiment. This resulted in a temporary strong reduction of NADW-formation, but with a subsequent recovery of the conveyor belt circulation. One should keep in mind that a meltwater pulse of 10 years duration is too short to allow estimates of the threshold value of the model for an infinitely long meltwater pulse. The threshold value in their model can be estimated to be 0.1 Sv, from their simulation described in Manabe and Stouffer (1997), with a prescribed meltwater input of 0.1 Sv maintained over 500 years.

4. The Sensitivity to Anthropogenic Climate Change

The question of the stability of the thermohaline circulation has also been discussed in connection with the climate response associated with the increase of greenhouse gas concentrations in the atmosphere. Many simulations investigating the transient behaviour to a slow increase of the atmospheric CO_2-concentration show a reduction of the NADW-formation and the overturning in the Atlantic (e.g., Manabe et al., 1990 or Cubasch et al., 1992), but the fast surface warming also reduces the formation of AABW. Manabe and Stouffer (1994) found a total decoupling of the deep ocean from the surface in response to a quadrupling of CO_2.

The response characteristics of the thermohaline circulation are fundamentally different for perturbations affecting only a single source of deep and bottom water (such as the meltwater experiments) or all sources (such as the CO_2 experiments).

In order to investigate the influence of possible long-term climate change, two 700-year simulations with increasing CO_2-concentration have been performed. The CO_2-concentration follows the IPCC scenario A (IPCC, 1990) until CO_2-doubling (after 60 years; experiment P2CO) or CO_2-quadrupling (after 120 years; experiment P4CO) is reached and then remains constant for the following years. A 700-year control integration with fixed present-day CO_2-concentration (PCON) has been performed for comparison.

At the time of CO_2-doubling, after 60 years, the warming is 1.4 K in the experiment P2CO (Figure 4a). In spite of the constant CO_2-concentration in the following years, the temperature increase rises to 2.5 K after 700 years. In the experiment P4CO, the warming is 3.8 K after 120 years, the time of CO_2-quadrupling. During the following centuries, the warming slowly increases up to approximately 4.9 K after 700 years. The increase of the near-surface air temperature during the periods with constant CO_2-concentration is due to adjustment processes in the ocean. In the years with fixed CO_2-concentration, the maximum rates of warming are located over the Southern Ocean and the North Atlantic where heat is transferred into the deep ocean.

The deep ocean, on the other hand, shows only a weak response during the period of CO_2-increase and a stronger warming in the following centuries (Figure 4b). The deep ocean is far from a new quasi-equilibrium state even after 700 years. This slow continuing adjustment of the ocean effects also the sea-level change due

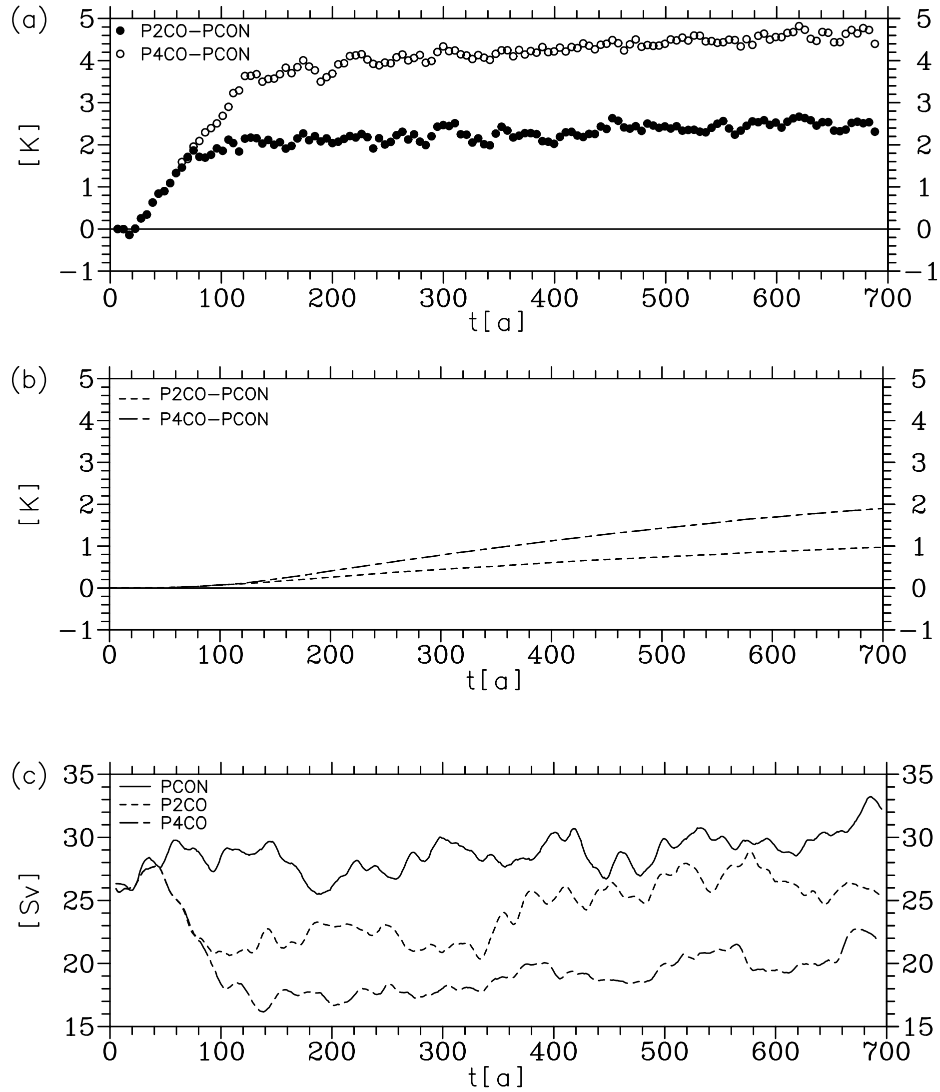

Figure 4. Response of the globally averaged near-surface air temperature (a), the horizontally averaged ocean temperature in 2000 m depth (b), and the maximum of the meridional stream function of the Atlantic (c).

to thermal expansion. After 700 years, the sea-level rise is more than 65 cm in the experiment P2CO and 135 cm in the experiment P4CO. Due to the large heat capacity of the deep ocean, the strongest sea-level changes occur during the final century of the experiments (7 cm in the run P2CO and 14 cm in the run P4CO). A possible sea-level rise due to an additional freshwater input into the ocean caused by meltwater supply from the ice sheets is not taken into account in the calculation.

The thermohaline circulation also exhibits remarkable changes. As an example, the time series of the maximum of the meridional stream function for the Atlantic are shown in Figure 4c. During the initial period, characterized by the CO_2-increase, the intensity of the thermohaline circulation weakens in both climate change experiments. In the upper layers of the North Atlantic, the temperature increases whereas the salinity decreases due to a precipitation increase. As a result, the stratification of the ocean becomes more stable and the convective activity in the northern North Atlantic decreases. In addition, the salinity of the subtropical upwelling region increases due to enhanced evaporation. The intensity of the overturning circulation weakens during the first 100 years of the experiment P2CO and during the first 150 years of the experiment P4CO. In the CO_2-doubling experiment P2CO, the intensity almost recovers during the following centuries, whereas the CO_2-quadrupling experiment P4CO shows only a slight increase and is about 30% weaker compared to the control run when the integrations were stopped. In both climate change experiments, an Atlantic "conveyor-belt"-type overturning pattern is present throughout the full simulation period. The overturning cell becomes shallower (i.e., the water masses in the northern North Atlantic do not penetrate as deep as in the control simulation) and the AABW penetrates further to the north.

While our CO_2-doubling experiment behaves similarly to the equivalent one performed by Manabe and Stouffer (1994) (MS94), our CO_2-quadrupling experiment does not show such a definite collapse of the NADW as in the MS94 simulations. Also, there are indications that even at $4\times CO_2$ the North Atlantic circulation recovers while it stays "dead" in MS94. One of the reasons for this different behaviour may be that our experiment starts from a higher level of overturning than the MS94 simulations that might be too strong to be completely suppressed. Moreover, the GFDL-model used by MS94 has a higher climate sensitivity than our model, causing a faster and greater warming of the atmosphere than our model, thus increasing the stability of the oceans due to a stronger warming and stronger freshwater input.

Contrary to the experiments in which only the freshwater input into the North Atlantic had been increased, the CO_2-experiments do not show a cooling in the North Atlantic, rather a reduced warming compared to the zonal mean (Figure 5). The horizontal pattern of the SST change is fairly similar in both simulations and shows a reduced warming in the North Atlantic. In none of the experiments (nor of the MS94 simulations), however, does the North Atlantic show a cooling, let alone catastrophic cooling as anticipated from the pure fresh-water input experiments. This means that the global warming due to greenhouse gases more than counterbalances the cooling by the reduced advection of warm water by the Gulf stream. Scenarios in which Europe experiences an ice-age in an otherwise warmer global climate (e.g., Rahmstorf, 1997) are not supported by our simulations.

The issue of the response of the thermohaline circulation to global warming is debated very much in the scientific community (Rahmstorf 1999, Latif et al., 2000). It depends on the formulation of the ocean model used or on the strength

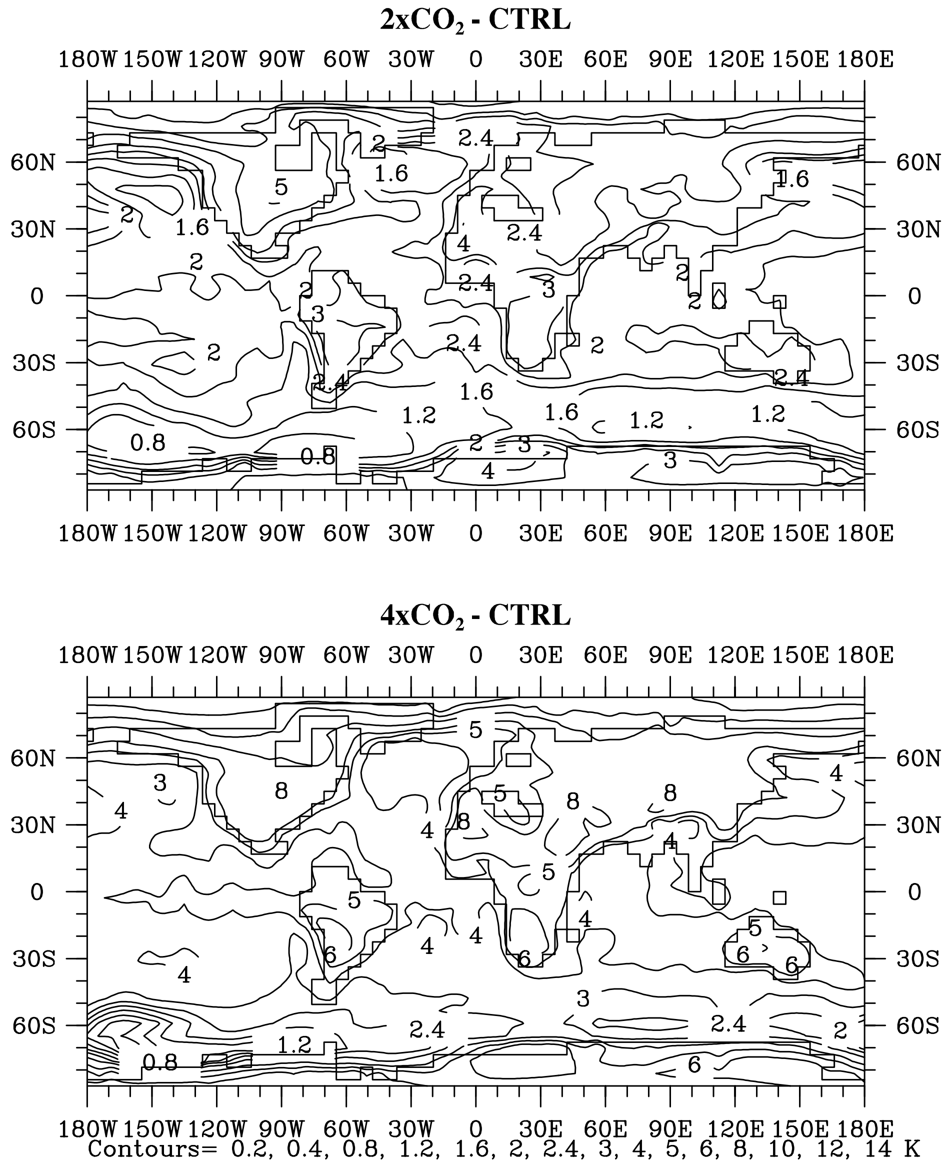

Figure 5. Annual mean sea-surface temperature (SST) change of the $2 \times CO_2$ (top) and the $4 \times CO^2$-experiment (bottom). Average over the years 651-700 of the experiments.

of the thermohaline circulation modeled whether it is a temperature effect or a freshwater flux effect, if the thermohaline circulation is affected at all. It would be premature at present to state that a final and generally accepted theory has been established.

5. Precipitation and Precipitation Intensity

The experiments discussed so far have dealt with time scales of centuries and longer. However, mankind is more interested in the more immediate future. Therefore several climate change simulations have been performed (for an overview see IPCC, 1990; 1992; 1996) in which the changes of climate under the influence of the anthropogenic emission of greenhouse gases have been studied. These studies have generally indicated that the hydrological cycle, i.e., the evaporation and the precipitation, increases on the order of 5% (Cubasch et al., 1994) within the next 100 years in a "business-as-usual" scenario. The global warming appears also to be affecting the THC in the North Atlantic, but not in all simulations. Manabe et al. (1990) indicate a 30% decrease in 100 years, caused mostly by the mechanism discussed before, while Roeckner et al. (1998) find almost no decrease due to a large atmospheric moisture export from the tropical Atlantic to the Pacific. On the whole, in this short time-period, the THC has not collapsed in any sensible experiment.

On the regional scale, the effects of the precipitation changes are much more drastic. As storm tracks and intensities change, snowfall comes later in the year, and snowmelt earlier, a number of local changes can be anticipated. A number of problems exist: the simulations are carried out with models which are too coarse to resolve the local topography, the experiments are too short and there are too few to place any statistical significance on the results of such a spatially and temporally highly variable quantity like precipitation. Only an increase of the computing resources by two orders of magnitude would create the possibility to simulate regional precipitation change and the stability of the thermohaline circulation in one integration.

A number of dynamical and statistical "downscaling" techniques have been developed (for an overview see Cubasch, 1998) to bridge the gap between the different time and space scales. In this particular study, results of the so-called "time-slice" method are presented. In this experimental technique, the uncoupled atmosphere model is forced by the sea surface temperature (SST) and sea ice distribution is taken from a transient simulation with a coarse resolution (T21, i.e., a Gaussian grid of 5.6°) globally coupled ocean-atmosphere model (Cubasch et al., 1992) at the point of CO_2-doubling ($2\times CO_2$) and tripling ($3\times CO_2$), and by the corresponding changed greenhouse gas concentration (Cubasch et al., 1995; 1996a). This experimental set-up makes it impossible to study changes of the THC, but they provide a large sample with a spatially high resolution. All experiments, including a control simulation with present-day climate conditions ($1\times CO_2$) have been carried out with the ECHAM3 T42 model (T42, i.e., a Gaussian grid of 2.8°) for an integration time of 30 years. The control experiment has also been used in the AMIP-model intercomparison (Gleckler et al., 1994).

The model data are compared with the rainfall data after Legates and Willmott (1990). Five of the six regions analyzed have been recommended by IPCC (1990). The sixth region is Central and Northern Europe (for more details see Cubasch et al., 1995).

6. Changes in Precipitation

The simulated precipitation over Central North America completely misses the peak during JJA (June, July, August) (Figure 6a). The mean value is only about 50% of the observed amount. The precipitation does not significantly change in any of the climate change simulations. Differences of a maximum of 20% still fall within the interannual variability of the control simulation.

Over Southern Asia (Figure 6b), the precipitation has a marked seasonal cycle with a minimum in early spring (the winter monsoon) and a maximum in JJA. The annual cycle of precipitation is simulated quite well, even though the maximum during the summer monsoon season is underestimated by about 30%. Under $2\times CO_2$ conditions this situation is hardly altered, and changes are barely significant. Under $3\times CO_2$ conditions, the precipitation in the monsoon season increases by 10%. It has already been stressed by Lal et al. (1995a) that the enhanced hydrological cycle results in an increase of the monsoon precipitation over India. However, in more recent experiments, which took into consideration the direct effect of anthropogenic sulfate aerosols, the Indian summer monsoon is rather weakened (Lal et al., 1995b) and that casts some light on the uncertainties associated with the prediction of precipitation.

In the Sahel (Figure 6c), the precipitation maximum during JJA is simulated quite well, but does not peak so sharply as the observations. The response to an enhancement of greenhouse gases is unequivocal: it decreases at $2\times CO_2$, but increases with $3\times CO_2$. Again, the changes are in the same order of magnitude as the error.

The annual precipitation cycle over Southern Europe (Figure 6d) is simulated well. However, the absolute amount is underestimated by a factor of two. Precipitation increases during the winter season (December, January, February, DJF) in both climate change experiments, but decreases during JJA. The seasonal cycle of precipitation is therefore enhanced.

While the JJA precipitation over Australia (Figure 6e) is well simulated, the steep increase in spring and autumn simulated by the model cannot be found in the observations. The annual mean is therefore overestimated. No clear signal can be identified for the precipitation change. The change is not significant in either

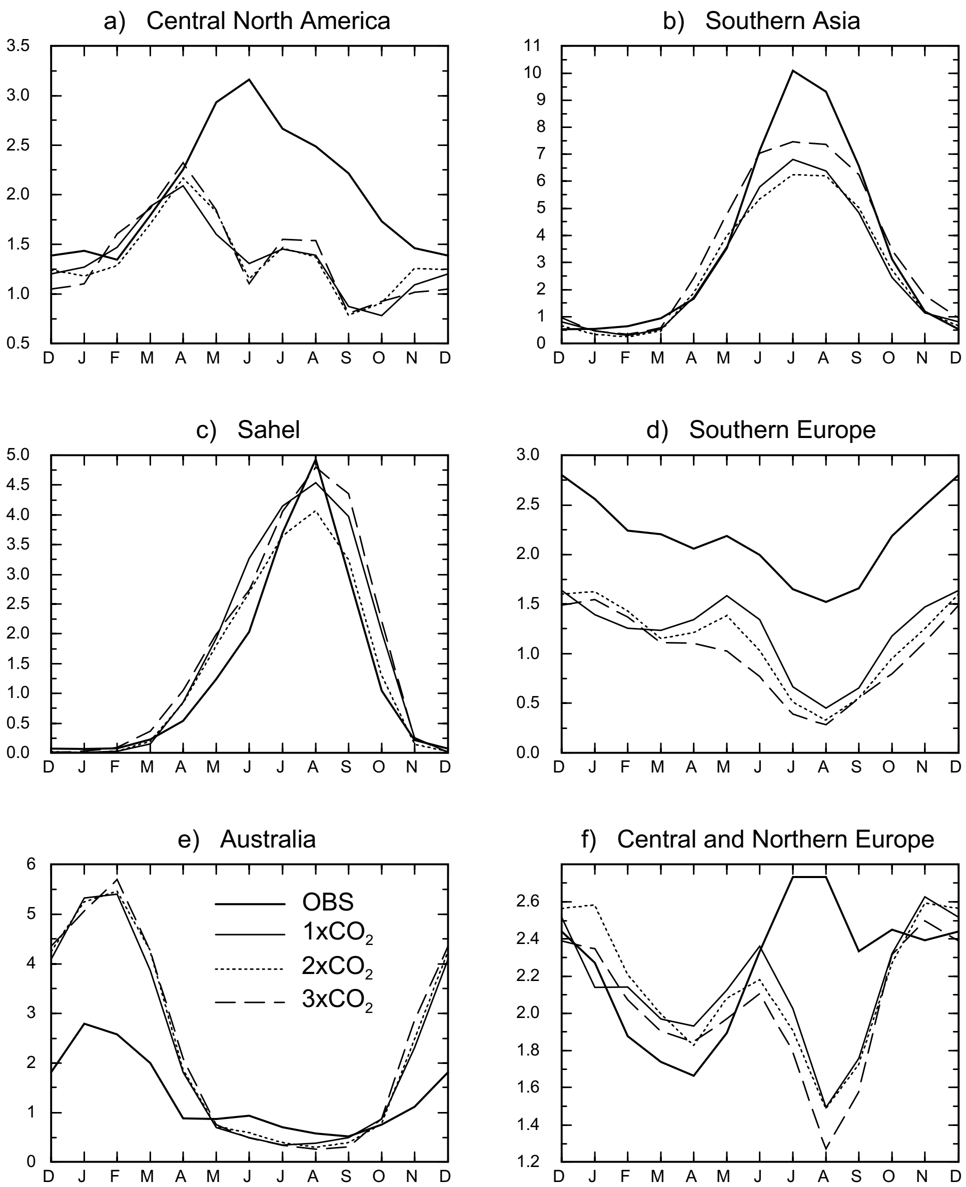

Figure 6. The annual cycle of the precipitation (mm/day) for the observation (bold solid) and the $1\times CO_2$ (thin solid), $2\times CO_2$ (dotted) and $3\times CO_2$ (dashed) integrations for Central North America (a), Southern Asia (b), the Sahel region (c), Southern Europe (d), Australia (e), and Central and Northern Europe (f) (after Cubasch et al., 1995).

the $2\times CO_2$ or $3\times CO_2$ and displays no seasonal cycle. No unequivocal significant change in the variability of precipitation can be found under the changed climate conditions.

In Central and Northern Europe (Figure 6f), the observed precipitation has a minimum in spring and a maximum during July/August. The minimum in spring is simulated, but instead of the precipitation maximum in JJA the model shows an absolute minimum. The simulated climate resembles the conditions of southern Europe rather than those of northern Europe. The responses to the climate change conditions are not very distinct.

Generally, the deviation in the precipitation simulation compared to observations is larger than the predicted climate change. Experiments using higher resolution do not necessarily show an improvement (Cubasch et al., 1996a). These problems appear to be common to all simulations of precipitation (IPCC, 1996).

7. Changes in the Precipitation Intensity and in the Frequency of Droughts

While the mean precipitation appears not to be changing much, and therefore changes of the THC are of minor concern for the anthropogenic change scientists, the extreme events are altering. This is of particular concern since they place a much greater strain on the biosphere, the lithosphere (via erosion) and industry (e.g., agriculture, insurance).

For example, Figure 7 displays those regions of the $3\times CO_2$-experiment where the heavy rain (defined as greater than 10 mm/day, 10 mm is about 20% of the monthly precipitation value in Germany) is changed significantly (determined by using the t-test) and has the same sign as in the $2\times CO_2$-experiment. The oceanic regions are the only areas where the heavy precipitation decreases, while over large areas of the globe, particularly in regions with tropical rain forest, an increase is evident. In southern Europe, in DJF particularly, the more intense precipitation increases, while the total annual amount remains almost unchanged (Cubasch et al., 1996a).

If more rain falls in less time, it also means that the number of dry days increases. This will have a greater impact if these dry days follow each other without interruption. To obtain a global overview of the absolute number of consecutive dry days, the “average waiting time for the next precipitation” on a randomly chosen day has been calculated. Figure 8b displays this quantity for the control experiment. The arid regions can easily be identified by the dark shades. The change of this quantity has been displayed in Figure 8a for the $3\times CO_2$-experiment, but again only for those points where the sign for the $2\times CO_2$ and the $3\times CO_2$-experiments coincide. The “average waiting time for the next precipitation” increases significantly in the mid-latitudes, while it decreases in the tropics and polar regions. The

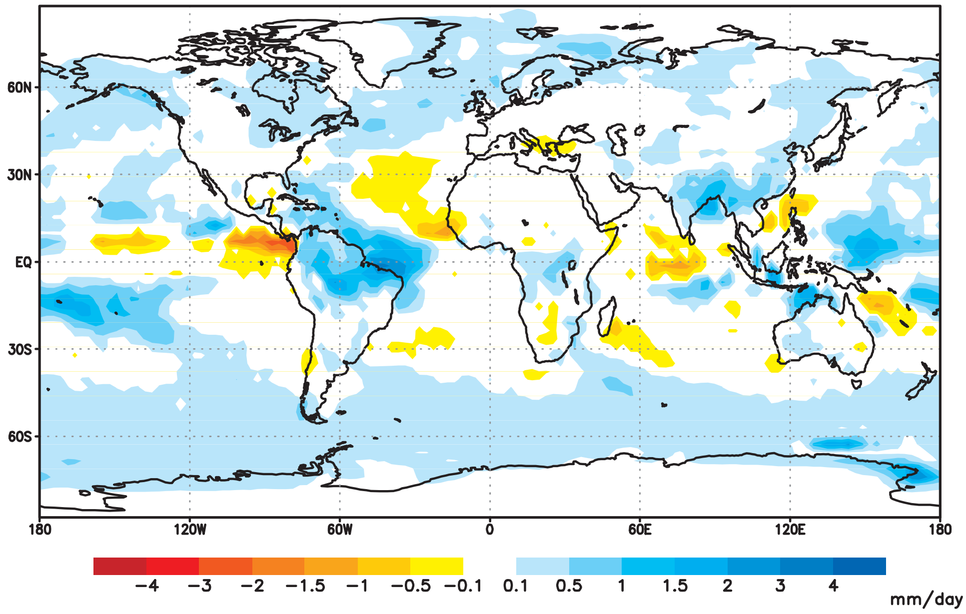

Figure 7. The regional distribution of the annual mean change of heavy rain (>10 mm/day) for the $3\times CO_2$-integration relative to the $1\times CO_2$-experiment. Regions, where the $2\times CO_2$ and $3\times CO_2$-simulation do not show the same trend or where the changes are not significant, have been blanked out (after Cubasch et al., 1995).

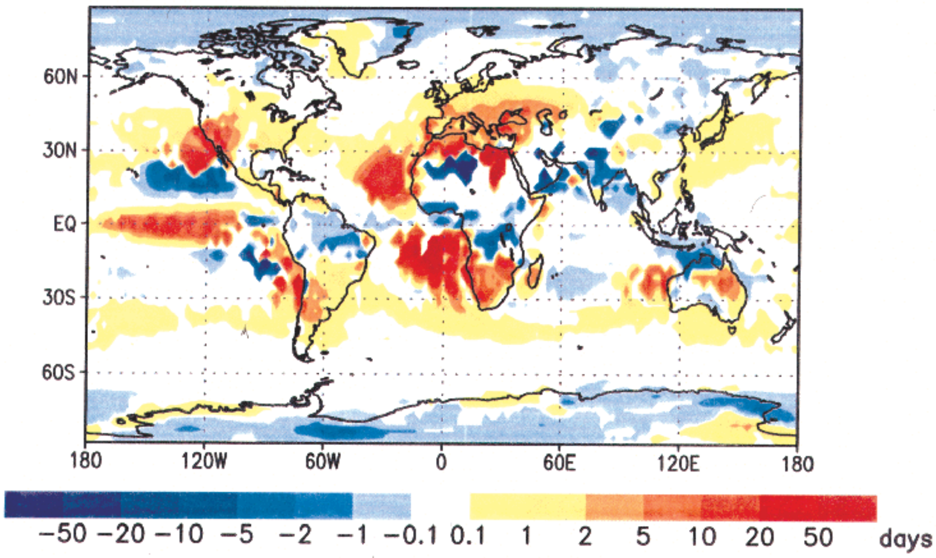

Figure 8. a) Changes in the regional distribution of the annual mean of the average waiting time for the next precipitation in the $3\times CO_2$-integration relative to the control experiment. Regions where the $2\times CO_2$ and $3\times CO_2$-simulation do not show the same trend have been blanked out (after Cubasch et al., 1995).

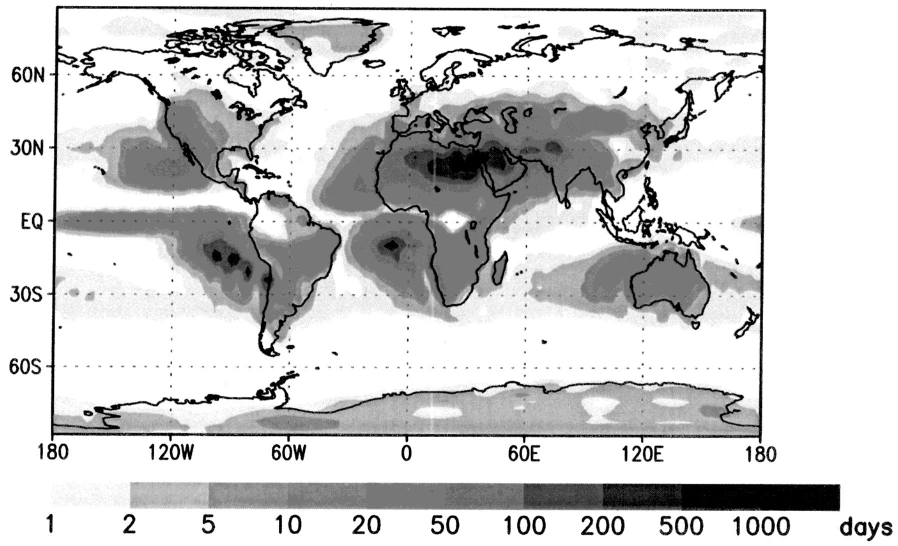

Figure 8. b) The regional distribution of the annual mean of the average waiting time for the next precipitation for the control integration in days (after Cubasch et al., 1995).

decrease of the waiting time for rain in some of the desert regions, notably the Sahara, is only of marginal importance and is not statistically significant.

8. Discussion

The wide range of time and temporal scales involved in going from the regional circulation, which is important for the precipitation, to the stability of the thermohaline circulation of the North Atlantic makes it impossible with present-day computing resources to perform simulations that consider all aspects at the same time. The results are therefore a bit “patchy”. One has to wait for improved models and increased computing resources before a comprehensive answer can be given.

Both the addition of freshwater or the global warming, cause the North Atlantic Deep Water formation to diminish. However, the coupled ocean-atmosphere system appears to have a number of feedback mechanisms that maintain or restore it. Only in extreme circumstances does the North Atlantic Deep Water formation collapse completely. In this case, however, global warming has advanced so far that the danger of an ice age over Europe (Rahmstorf, 1997) in an otherwise super warmed world seems to be unlikely (Manabe and Stouffer, 1993). Within the next 100 years, which is the typical time-scale simulated in global warming studies, a slowing down of the thermohaline circulation on the order of 20-30%, but no complete collapse can be found in most studies. It has to be stressed that most of the models have no or only a rudimentary representation of the Greenland Ice Sheet. It might be that its comprehensive inclusion might give an additional freshwater input into the North Atlantic, thereby perturbing the THC more than has been simulated up to now. It must also be noted that investigations of the stability of the thermohaline circulation have so far only been performed with models with a resolution too coarse to explicitly simulate deepwater formation. This process is usually parameterized in a simple manner. We thus have to rely on the assumption that these parameterizations react correctly to the change in the large-scale hydrographic conditions that the current generation of models are able to resolve.

For the near-surface temperature, a smooth response function can be fitted. If the North Atlantic Deep Water formation has to be described by a response function, a critical threshold can be defined.

The hydrological cycle is found to increase in all greenhouse gas simulations. Precipitation, however, is a spatially and temporally very variable parameter. It is also difficult to observe, to simulate and it does not show a clear trend under enhanced greenhouse gas conditions. It is also very sensitive to the experimental conditions: the inclusion of aerosols can change its sign over certain regions of the globe (Lal et al., 1995b; Cubasch et al., 1996b). To obtain more regional information for longer times and with a larger sample, a dynamical method has been developed, the so called time-slice approach. From the model output generated with this method, a number of derived parameters can be analyzed, such as the frequency of droughts or precipitation intensity. It is difficult to assess the quality of these simulated model quantities, as no comparable observed data on a global scale are available.

The analysis presented here indicates that the simulated change for precipitation within the next 100 years is smaller than the error of the simulation. It can also be seen that there is a tendency towards more intense precipitation and longer drought periods. This combination certainly influences vegetation and soil erosion, and it has economic consequences.

The elusiveness of precipitation, and its spatial and temporal variability, make it necessary to use various “downscaling”-techniques to be able to derive response or threshold functions. On a global scale, it seems to be difficult to define a critical threshold for precipitation, on a local scale droughts or floods are a natural choice.

Acknowledgements

The authors thank J. Waszkewitz and A. Hellbach for the technical support to this paper. J. Jones improved the English. The research has been supported by the German Ministry for Education, Research and Technology (BMBF), the Max-Planck-Gesellschaft and the European Commission Environmental program (EV5V-CT92-0123; ENV4-CT95-0102).

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(Received 30 December 1997; in revised form 18 April 2000)