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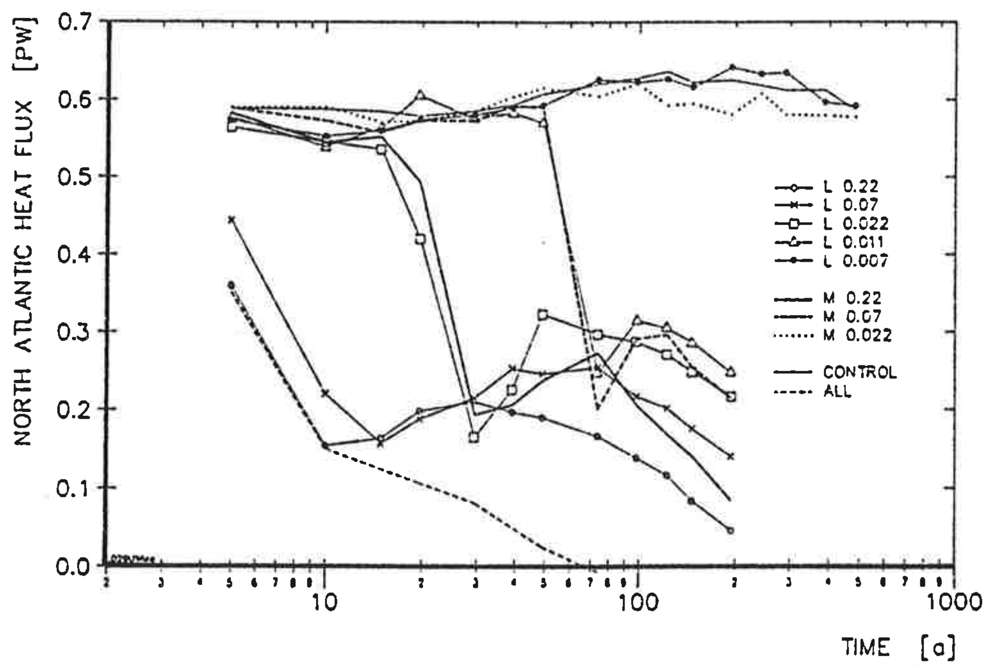

EXPERIMENTS WITH AN OGCM ON THE CAUSE OF THE YOUNGER DRYAS

by

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Experiments with an OGCM on the Cause of the Younger Dryas

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ABSTRACT

We investigate the effect of enhanced riverine freshwater input on the circulation of the northern Atlantic. It has been suggested by several authors that during the last deglaciation the melt water may have blocked the formation of deep water in the Atlantic and, consequently, reduced the meridional heat transport. The subsequent cooling of this region is identified with the Younger Dryas epoch. We found that the freshwater input rates of the deglaciation were more than sufficient to change the Atlantic circulation significantly. A doubling of the present St.Lawrence river could also be critical for the present circulation.

1. INTRODUCTION

It has been known for a long time that the last deglaciation was not a monotonic development (cf. Ruddiman and Duplessy 1985, and Broecker *et al.*, 1988). Particularly in the region of the northern Atlantic, it was characterized by an oscillatory behavior which was most pronounced between 10,000 and 11,000 years before present. The return of the glaciation, in the form of the Younger Dryas event, was preceded by a strong release of melt water into the northern Atlantic, probably due to a deflection of the melt discharge from the Mississippi to the St.Lawrence river. It has been suggested (e.g.Broecker *et al.*, loc.cit.; Berger and Killingley 1982) that the reduction of deep water formation by the resulting strong stratification could drastically reduce the meridional heat transport of the Atlantic and, consequently, provide a substantial cooling of the Arctic. The basic mechanism of this process is clear. However, it is an open question whether the actual amount of glacier melt was sufficient to trigger such an event.

The Atlantic is widely acknowledged to be a very sensitive ocean with respect to the details of forcing mechanisms. Recently experiments by Bryan (1986) with an idealized model of the Atlantic have shown that with a given forcing of the circulation with freshwater fluxes, the ocean can develop at least two completely different stable states of circulation, depending on the initial state. This result was postulated on the basis of a careful examination of geological evidence by Broecker *et al.* (1985). It must be noted, however, that the design of Bryan's basic experiment was not entirely realistic: there is no way to increase the salinity of the northern Atlantic suddenly by so much as 2 permille.

A rough estimate of the mean deglaciation rate yields a glacial melt of about 0.1 Sv. The major part of this originated from the continental ice sheets of the northern hemisphere, with about 60 % attributable to the Laurentide ice sheet (cf. Hughes *et al.*, 1981). It is unlikely that the glacial melt reached more than twice this value at intermediate episodes (Ruddiman *et al.*, loc.cit.). In this paper, we describe some experiments in which the assumed melt rate of the Laurentide Ice sheet is varied in strength and in the location of inflow to the Atlantic.

The conventional way to run OGCMs is to determine the surface fluxes of momentum, heat, and freshwater by prescribing climatological fields of wind stresses, and by requiring the surface values of temperature and salinity to adapt to the prescribed climatological surface data. For a simulation of the glacial ocean, such a procedure is not suitable, as we do not know the surface salinity of former climatic states with sufficient accuracy. However, as the ice edge in the preceding Alleroed epoch was at almost the same position as it is today, it can reasonably be assumed for the present experiment that the ocean circulation in the Alleroed period was similar to that of today. Thus we introduced the freshwater inflow as a perturbation of the present day circulation, driven by present day wind stress, atmospheric temperature, and freshwater fluxes. By comparison of this experiment with a control experiment without freshwater input, the impact of the glacier melt can be identified.

2. THE MODEL

The model is based on the standard set of equations used in numerical models. These are the conservation laws for salt, heat, and momentum, the latter in a linearized form. The equation for the vertical component of momentum is replaced by the hydrostatic approximation. The discretisation in time is written in a rigorously implicit way: all equations are taken as Euler-backward differences. The resulting systems of linear equations for the velocity field are solved iteratively for the velocity shear between neighboring layers (baroclinic modes) and by elimination for the vertically integrated transport (barotropic mode). Details are given in Maier-Reimer *et al.* (1989). This formulation almost completely filters out gravity waves which necessitate a rather short time step in conventional circulation models.

The equatorial Kelvin waves are formally included, but because of the coarse grid and the long time step, they are strongly damped. Outside the equatorial region, the motion is essentially geostrophic, with frictional effects near the boundaries. Adjustment of the density field to the forcing is provided almost entirely by Rossby waves, as proposed by Hasselmann (1982).

For the experiments described in this paper, the model was run with a horizontal resolution of 3.5 degrees in the zonal and meridional directions and with 11 levels of depth. The topography is as realistic, as far as possible in such a coarse grid. The minimum depth of shelf regions was assumed to be 200 m. The model was run with a time step of 30 days. In the basic run, the model was driven by the Hellerman data of monthly wind stress (Hellerman and Rosenstein 1983), and by the freshwater flux resulting from a soft restoring to the observed annual mean (Levitus 1982). The thermal driving was given by a similar restoring relation of the surface temperature to the monthly mean values of atmospheric temperature from the COADS data set (Woodruff *et al.*, 1987). At high latitudes, the heat flux is reduced by the presence of ice which is simulated by an ice model (Stefan 1891) with simple thermodynamics and advection with viscous rheology.

Originally, the model was spun up from a state of homogeneous water with a salinity of 34.6 and temperature of 2.5°. After 10,000 years of integration, the model ocean reached an almost stationary state: the global mean temperature in the lowermost level changed by 1.7 milliKelvin during the last millennium; the corresponding salinity change was 4.1×10^{-8} permille/year. From this run, the effective fresh water fluxes resulting from the boundary condition were stored. In the following experiments, the ocean was driven by the same wind field and atmospheric temperature as in the basic run; the salinity changes now result from an explicit prescription of freshwater fluxes. As the model now contains no stabilizing restoring term for salinity, it must be expected that this change in the boundary condition will cause the model to drift away from the former state, particularly as the convective adjustment is a highly nonlinear process which acts as an effective stochastic forcing. The drift, however, is so slow that within the first centuries of integration the deviations can barely be detected. We compare this control run with the anomaly runs, in which the melt water input is prescribed as an additional freshwater input at isolated points.

3. THE CONTROL EXPERIMENT

In the control experiment, the circulation is characterized by a pronounced difference between the circulation patterns of the major oceans, as observed in the present day real ocean. Figure 1 shows the zonally integrated meridional mass transport stream function. In the Atlantic, the cross equatorial transport of water reaches 20 Sv. This large overturning yields a substantial heat transport across the equator, as deduced also from atmospheric observations (Hastenrath 1982). In the Pacific, the meridional circulation is dominated by the equatorial Ekman cell and a slow deep ventilation from the Antarctic. The resulting heat transport is almost antisymmetric around the equator (Fig.2).

These meridional circulation patterns cannot be directly measured in the real ocean, but are deduced from the distribution of geochemical tracers, most clearly from the field of natural ^{14}C. The ^{14}C distribution can be simulated fairly realistically with a simple passive

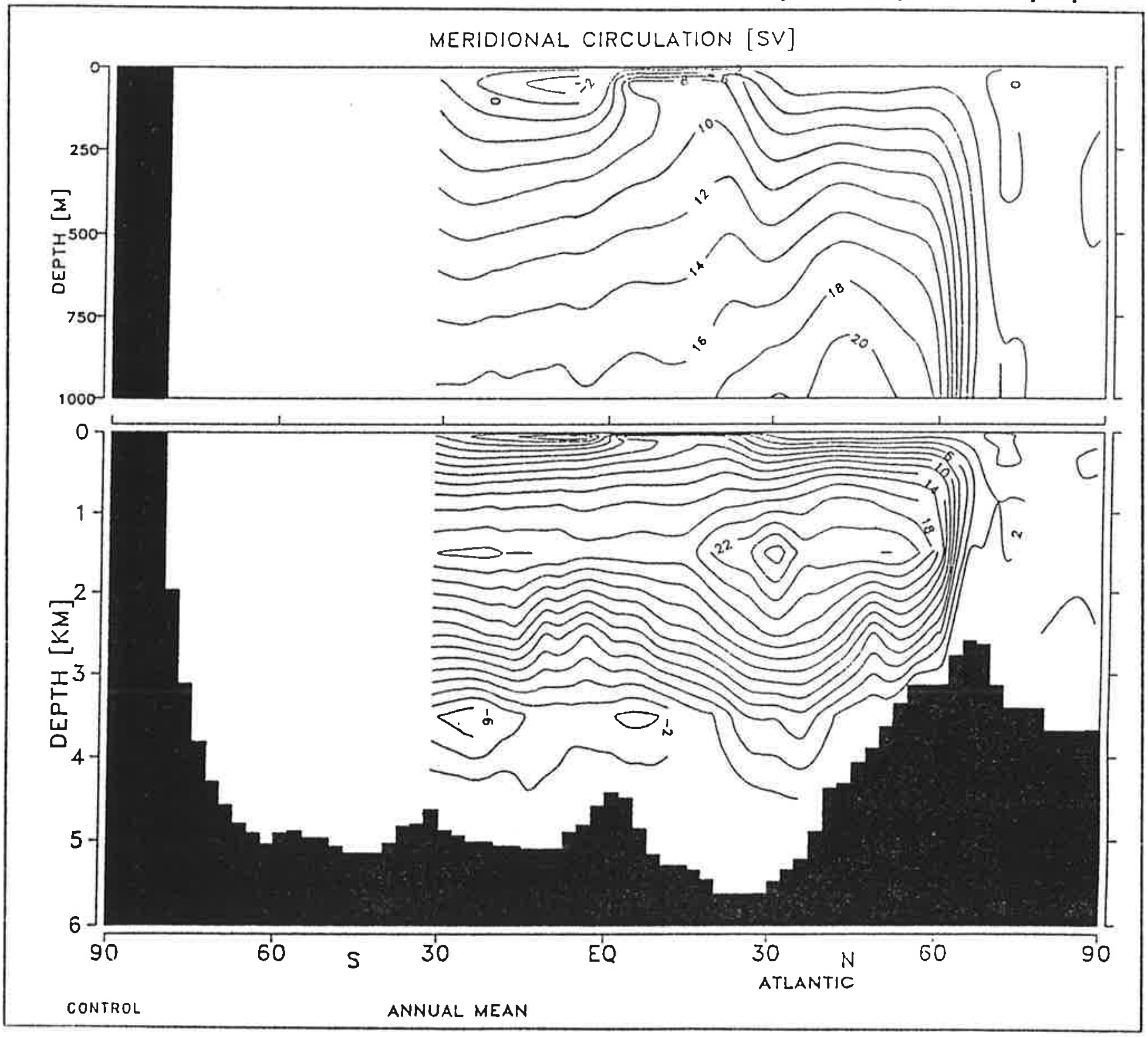

Figure 1a. Zonally integrated mass transport stream function in Atlantic. Contour interval is 2 Sv.

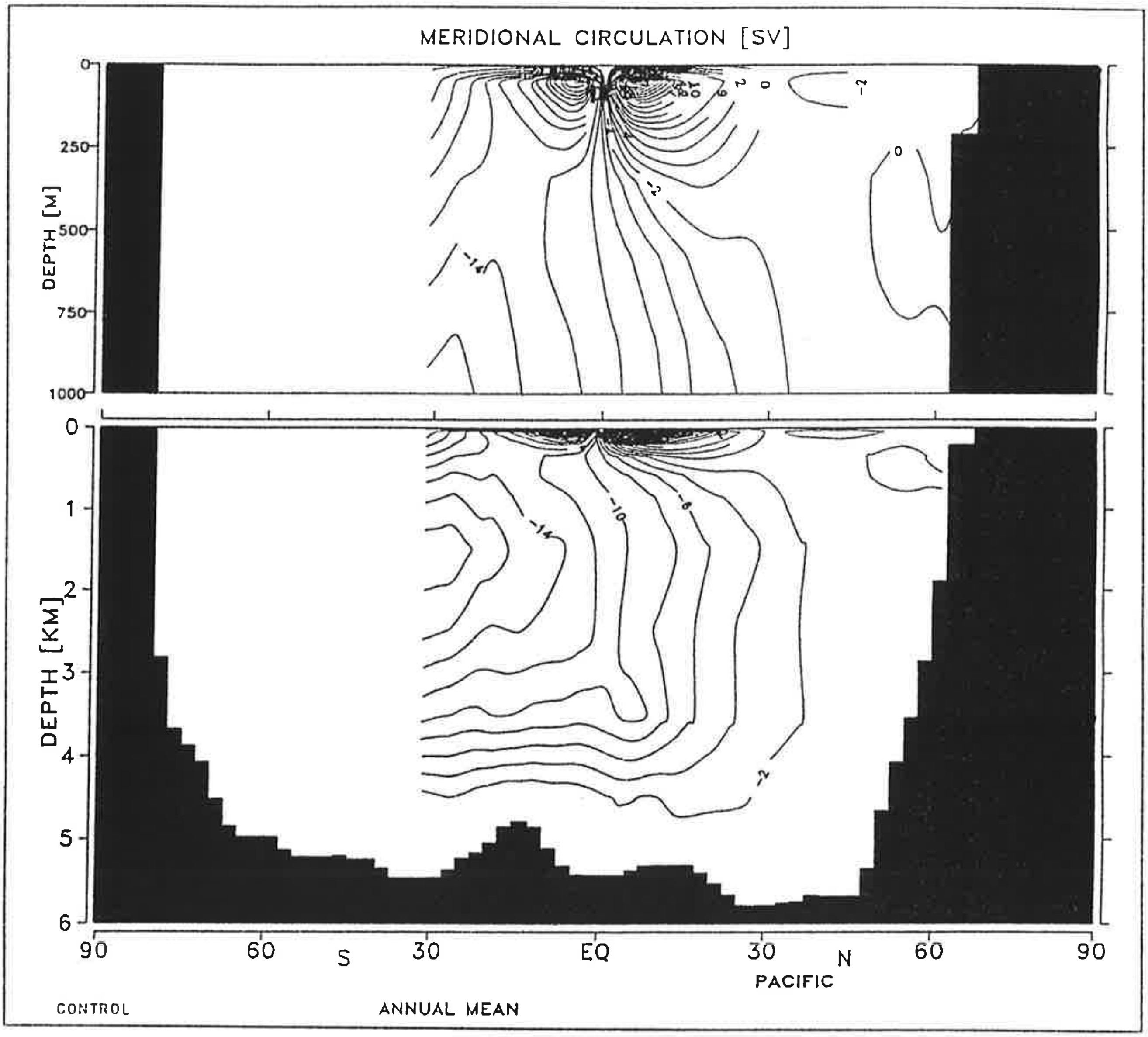

Figure 1b. Zonally integrated mass transport stream function in Pacific. Contour interval is 2 Sv.

tracer model. The differences between this simulation and more complex ocean carbon cycle models (Maier-Reimer and Hasselman 1987; Bacastow and Maier-Reimer 1989) are minor in the deep ocean. We prescribe the gas exchange with the atmosphere by a constant piston velocity of 6 m/year and treat the advection and radioactive decay as a stationary problem. The resulting set of equations is solved by a simple iterative procedure which yields the stationary distribution in a few minutes of computer time. This "quick ^{14}C " computation has proven to be a very effective diagnostic tool for the validation and discrimination of different circulation fields obtained with different choice of boundary conditions. Fig. 3 shows the resulting distribution for the present day ocean. There is good agreement with the corresponding GEOSECS sections (Stuiver and Östlund 1980; Östlund and Stuiver 1980).

Fig. 4 shows a block diagram of the large scale transports in the form commonly used in box models. The upper panel shows the horizontal transports in the top 1,500 m, the meridional components in the laterally

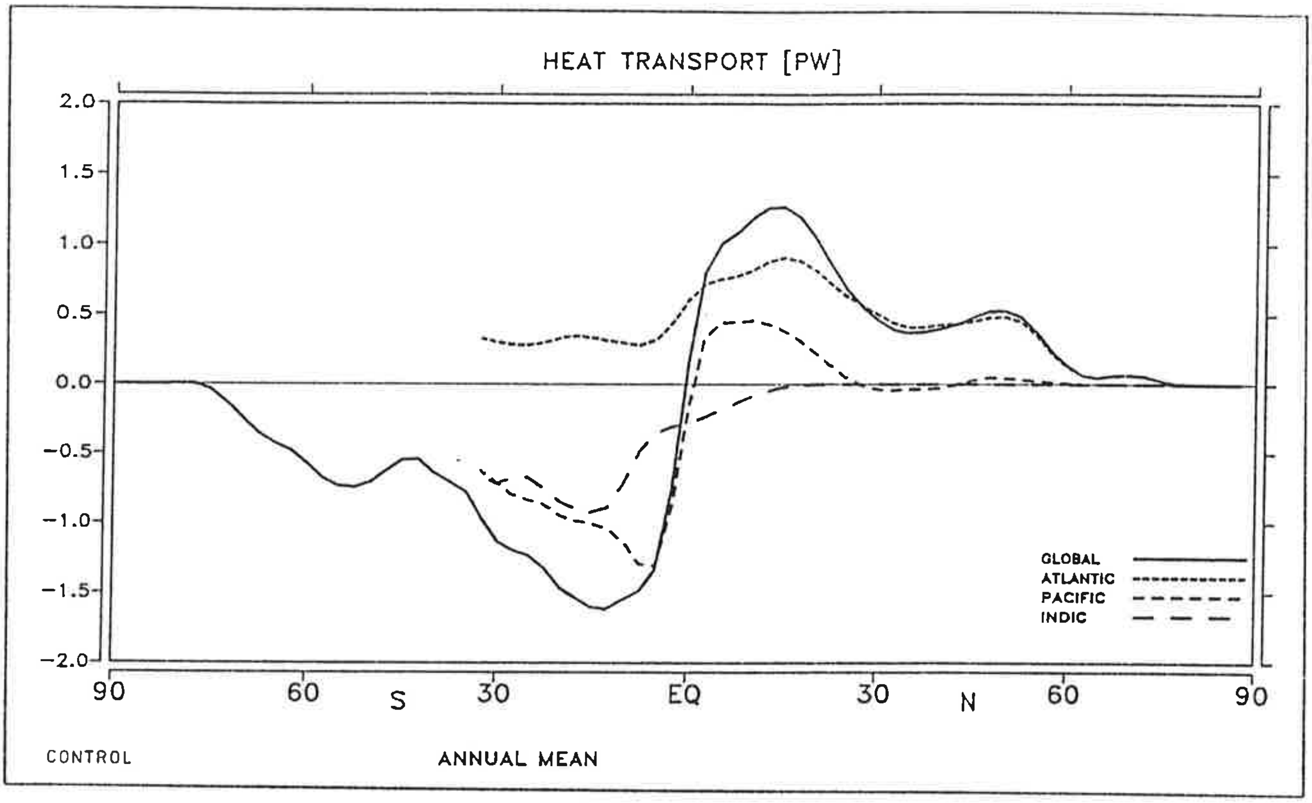

Figure 2. Heat transport of the major oceans in the control run

bounded oceans across 10 and 40° N and S, resp., and the zonal component in the southern circumpolar ocean. The middle panel shows the vertical component across the level 1,500 m, integrated over the areas defined by these bounds. The lower panel shows for the deep layers the values corresponding to the upper panel . This summary shows that the model clearly reproduces the structure of the conveyor belt (Gordon 1985): the salty water from the northern Atlantic spreads into the deep layers of the global ocean. With the model topography, the transport through the Banda strait is only 1.8 Sv. The return flow occurs almost entirely through the Drake Passage.

The deep circulation is driven primarily by the spreading of dense water produced in the regions of strong cooling. Figure 5 shows the locations and strengths of the production zones of deep water in the control run. The most important regions are the Irminger sea and the shelves around Antarctica.

4. THE MELT EXPERIMENTS

In a first experiment, we prescribed a very large melt water inflow (six times the estimated mean postglacial melting rate), distributed equally at three points: Gulf of Mexico, St. Lawrence River, and Norwegian Sea, respectively. After 10 years of integration, the thermohaline circulation in the Atlantic had completely collapsed . The heat transport of the Atlantic was directed southwards everywhere. As this assumption of melt water input is probably unrealistic, we will not discuss it further.

In a series of further experiments, we examined the influence of the individual locations of melt water inflow, reducing the flow rate from experiment to experiment. The main features of these experiments are summarized in fig. 6, which shows the development in time of the

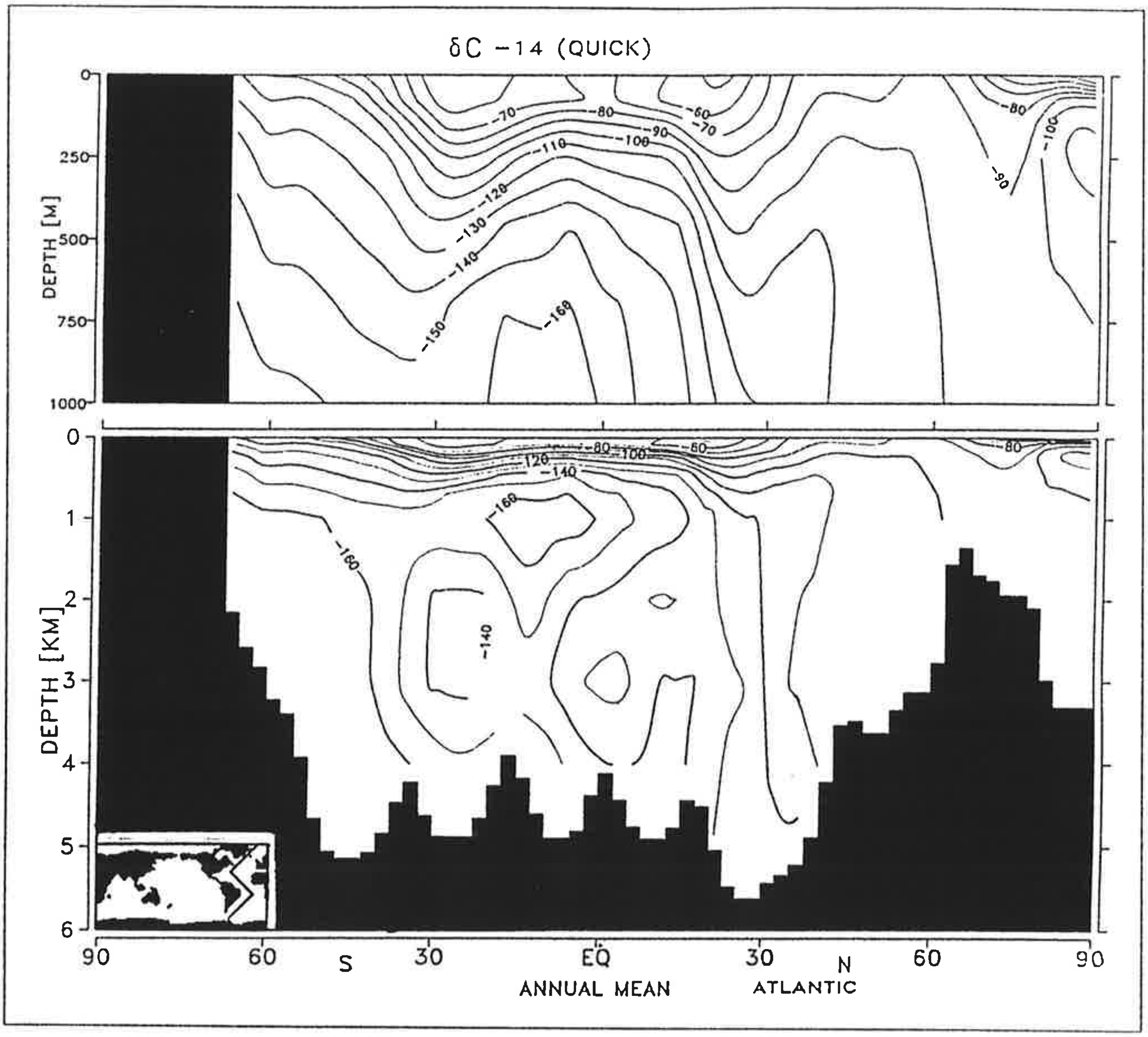

Figure 3a. Distribution of natural ^{14}C for the control run obtained with a constant piston velocity of 6 m/y. Atlantic.

surface heat flux, integrated over the region north of 30° N in the Atlantic, including the Arctic Ocean. In a stationary state, this quantity is balanced by the meridional heat transport in the Atlantic across 30° N. The heat flux responds immediately after the moment when the surface salinity is reduced to the point where the brine released during ice formation can no longer penetrate the underlying haline water. In all experiments, the switch off of the driving mechanism of the heat transport, when it occurs, is found to be a rapid event. It is followed by an oscillation with a period of some twenty years which may be explained by the generation of large amplitude Rossby waves on the arrival of the applied distortion at greater depths. We have not yet analyzed these oscillations in detail.

Fig. 7a,b show the resulting meridional stream function and heat transport after 200 years of integration for the experiment M 0.22 in which an additional release of 0.22 Sv was introduced at the Mississippi River. The contrast to the reference run is striking. The heat transport is completely reversed; the Atlantic no longer exports heat from the Arctic across the equator. The climatic feedback of this reversal would have been dramatic. These figures are not, of course, to be interpreted as stationary states. The integration time was too

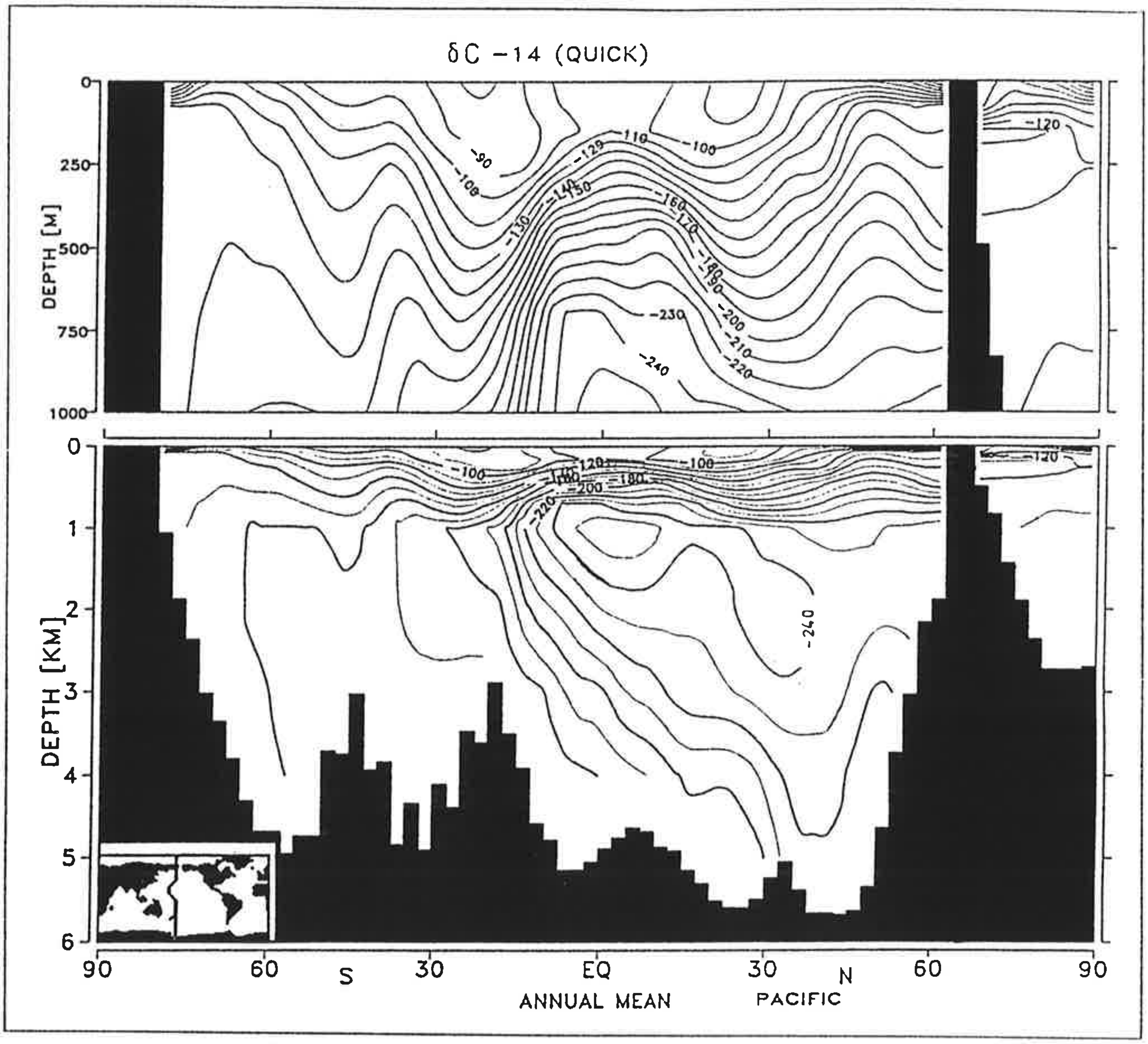

Figure 3b. Distribution of natural ^{14}C for the control run obtained with a constant piston velocity of 6 m/y. Pacific.

short for a new equilibrium to be established. They represent rather a snapshot of a changing circulation.

As the Mississippi outflow is transported to the polar front by the Gulf Stream, the overall perturbation introduced by this outflow is not as would be immediately inferred from an examination of the local stratification. One major difference lies in the timing: the Mississippi water needs a longer time to reach the sensitive region of deep water production. On the way to this region it is diluted, and part of it recirculates in the subtropical gyre. However, a melting rate of only 0.07 Sverdrup at either of the two locations is sufficient to return the Atlantic circulation to almost the same mode as in the first melt experiment within 200 years of integration. For the Mississippi case, the rapid change of the heat flux is delayed by some 60 years relative to the St.Lawrence case. With a melting rate of 0.02 Sv, released via the St.Lawrence river, we still obtain a strong reduction of the heat flux to about one third of the undisturbed state for the St.Lawrence outflow after 30 years. The same melting rate, released at the Mississippi, produces no

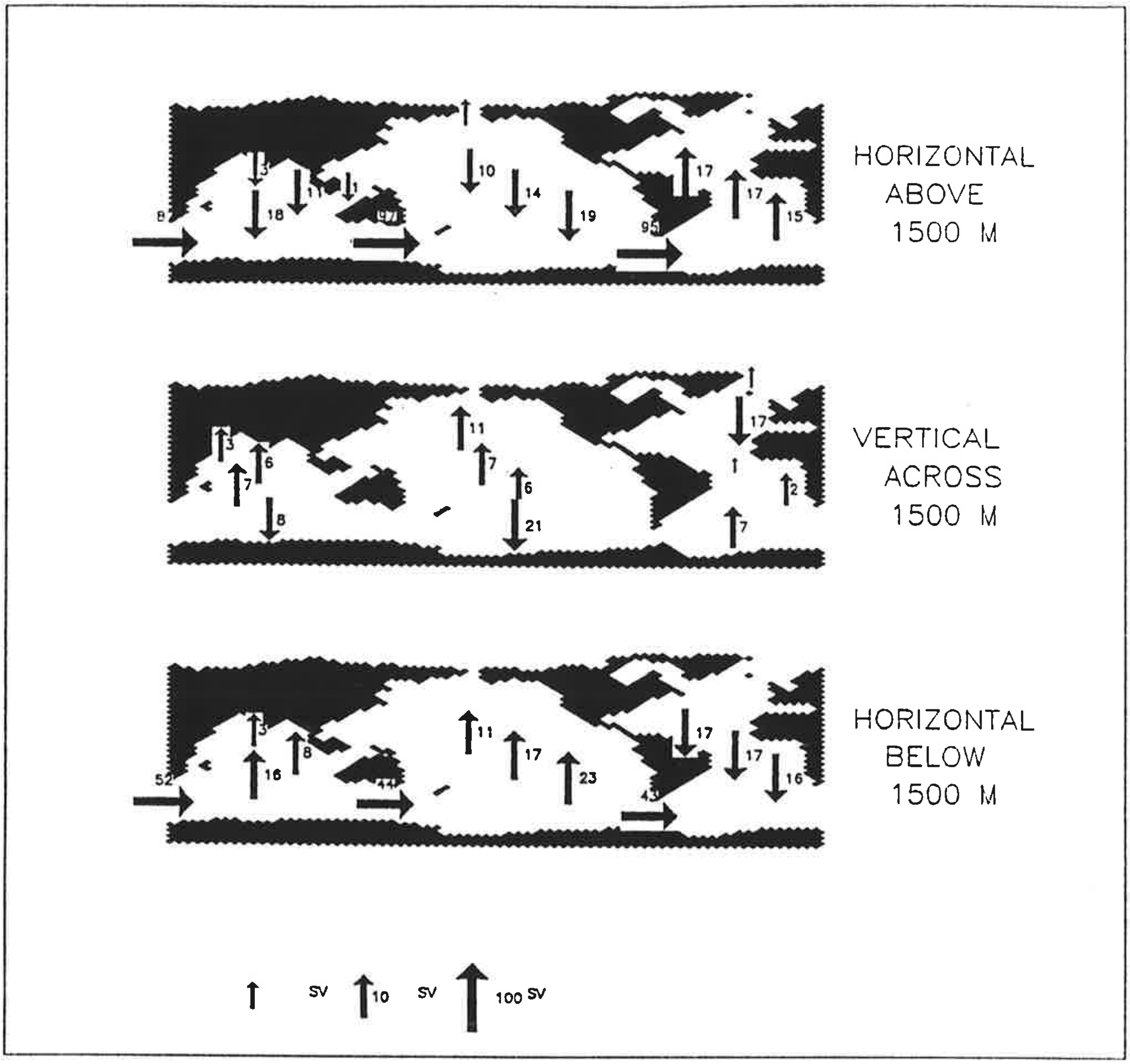

Figure 4. The conveyor belt of the control run

substantial changes of the heat flux and the circulation even after 500 years of integration. A similar cutoff of efficiency is found with a St.Lawrence output only after the meltwater input had been reduced to 0.007 Sv. (It should be emphasized again that most of the experiments assume a smaller melting rate even than the long term average of the deglaciation. Today the Mississippi is estimated to carry 0.017 Sv, the St. Lawrence river 0.01 Sv (Baumgartner and Reichel 1975). These modern river discharges are reflected in the fields of surface salinity which were used to force the basic run).

When comparing the effect on the heat flux of these experiments with the total amount of freshwater input, we observe a maximum efficiency at the moderate melting rates. Very weak releases are compensated by dilution; the stabilizing effect in the regions of deep water formation is not strong enough. On the other hand, very strong melt water outlets cannot do more (within the framework of this study) than block the exchange of surface water with the deep sea, and it takes some time for the additional fresh water to reach the region of deep water formation. For the St. Lawrence outflow, a

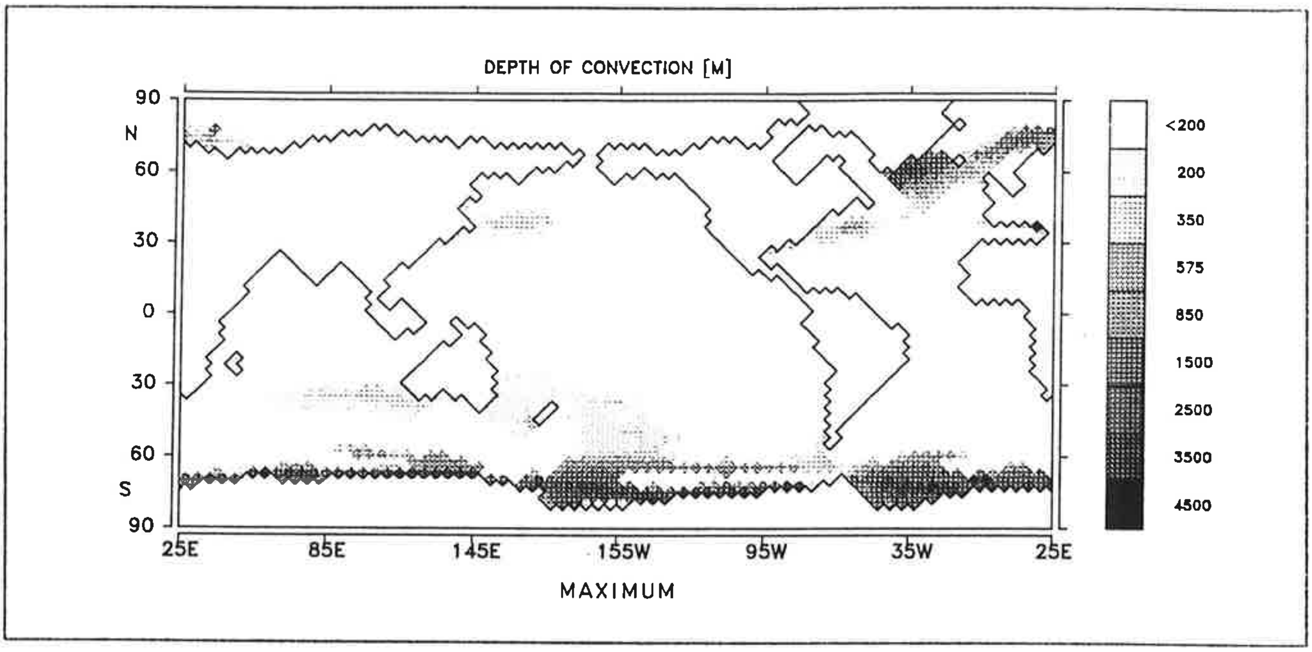

Figure 5. Deep water formation of the control run.

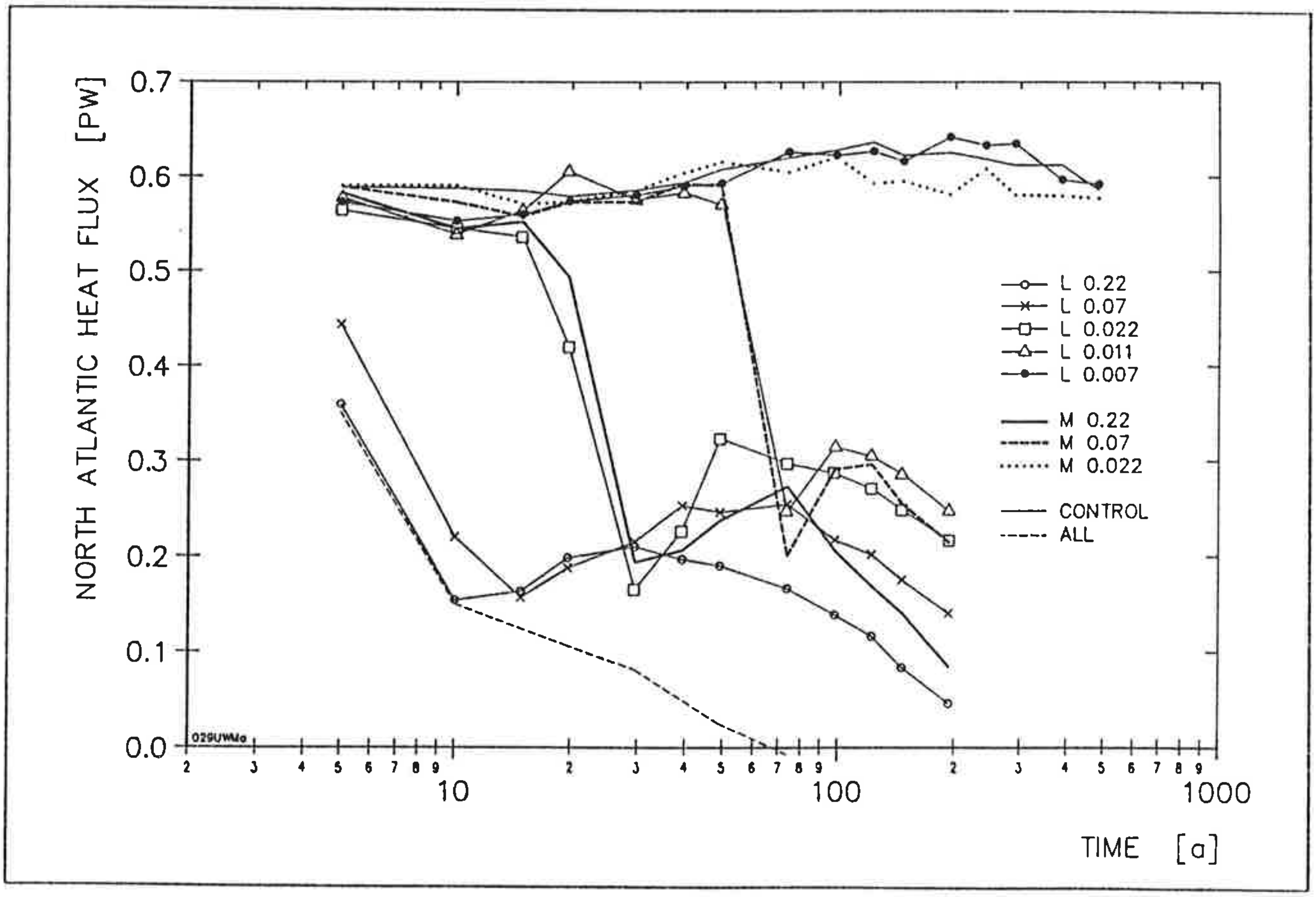

Figure 6. Time series of the North Atlantic heat flux from the melt Experiments. L and M denote release at St.Lawrence River and Mississippi, resp. The numbers indicate the strength of the melt water inflow in Sv. "All" denotes the experiment with six times the mean deglaciation rate.

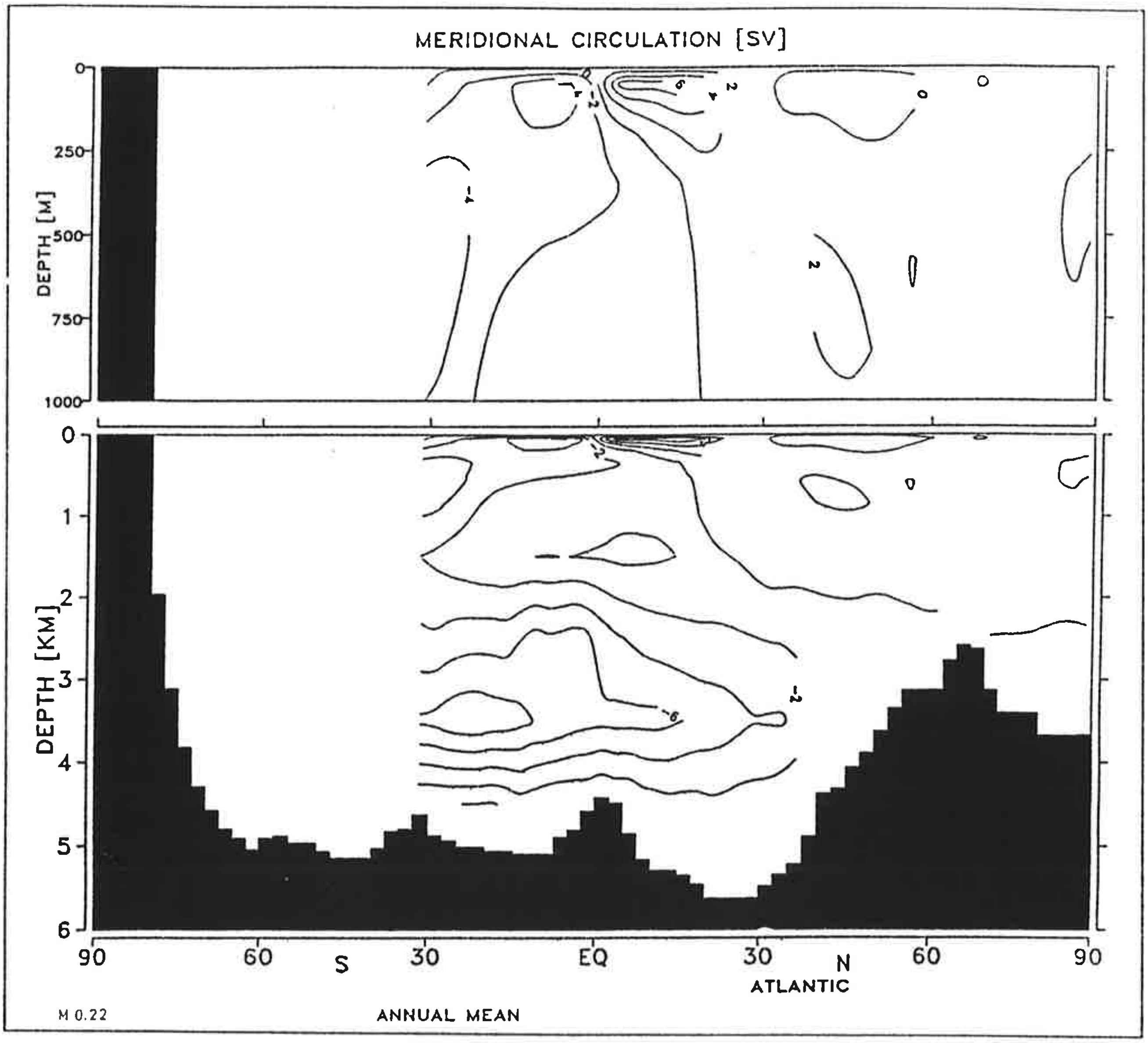

Figure 7a. Meridional stream function of the experiment M 0.22 after 200 years.

threshold value lies between 0.007 and 0.011 Sv. The reduction of the melt water input by a factor of 3 extends the period in which the heat flux remains stable by at least an order of magnitude. For the Mississippi outlet, the corresponding threshold lies between 0.02 and 0.035 Sv.

So far, we have discussed only the heat flux. A more direct climatic feedback is seen in the ice coverage. The melt experiments were forced by exactly the same temperature field as the control run. The creation of a fresh water lens at the surface makes it more difficult for the brine released by the ice formation to penetrate to greater depths. As a consequence, the melting of sea ice by heating from below is inhibited, and the ice coverage increases. Figure 8a shows the March ice coverage of the control run. The polar front is clearly seen north of Iceland. The ice edge is represented by the last 20 cm isoline. The Labrador sea ice is bounded by a line from Cape Farwell to Newfoundland; the Irminger sea is ice free. Fig 8b shows the same for the experiment M 0.22. Now, the Irminger sea is ice covered in March, and the polar front lies across Iceland.

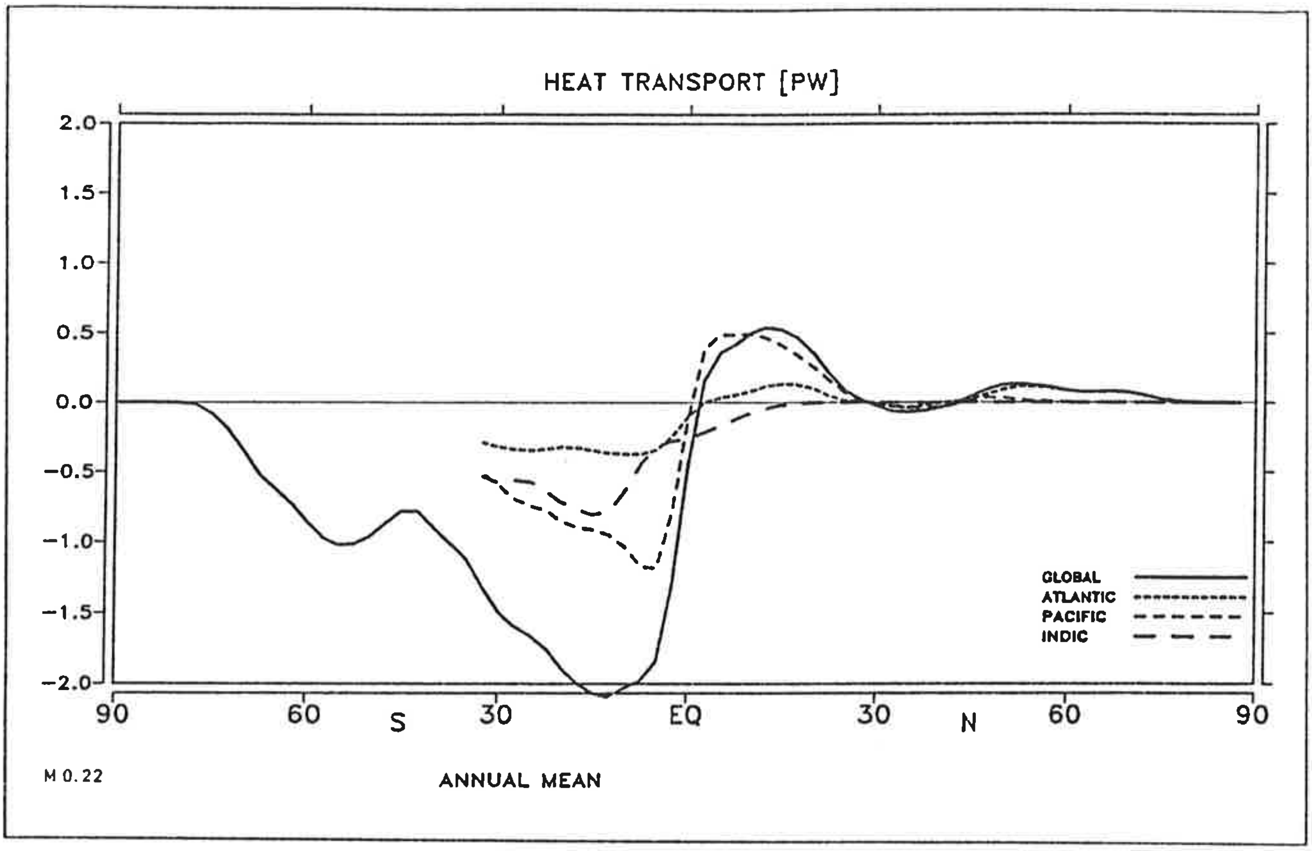

Figure 7b. Heat transport of the experiment M 0.22 after 200 years.

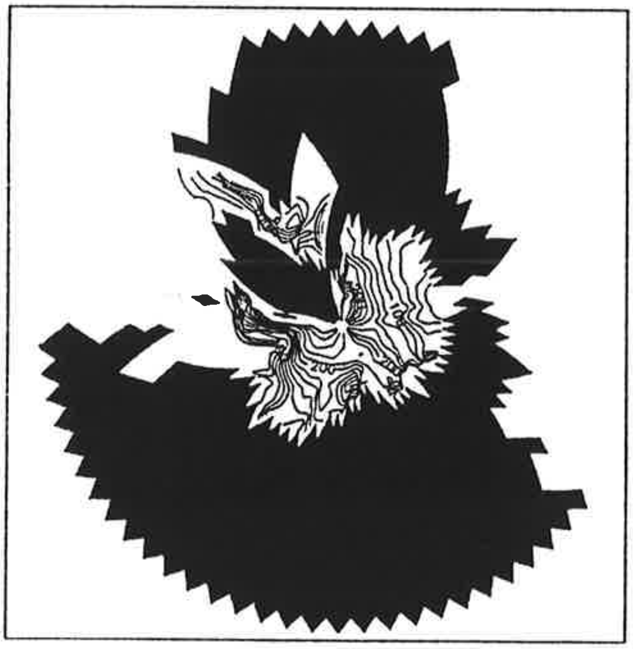

Figure 8a. March sea ice coverage of the control run after 200 years.

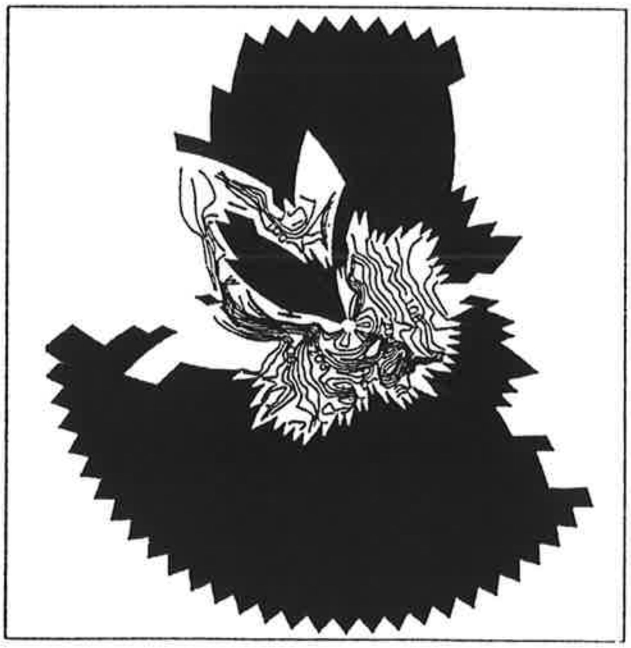

Figure 8b. March sea ice coverage of the experiment M 0.22 after 200 years.

5. CONCLUSION

We have found again that the circulation in the northern Atlantic is extremely sensitive with respect to the assumed boundary conditions. A perturbation significantly smaller than assumed in Bryan's experiment is sufficient to create substantial changes of the circulation patterns. For the interpretation of the Younger Dryas event, the present experiments are not, of course, the whole story: the assumption of constant atmospheric temperature and constant freshwater exchange with the atmosphere is not realistic for a changing ocean circulation. With decreasing oceanic heat transport, the overlying atmosphere must become colder and act as a negative feedback. However, we have demonstrated that a realistic estimate of melt water inflow is more than sufficient to trigger some form of substantial climatic fluctuations. We have also confirmed Broecker's hypothesis that the location of the inflow plays a critical role. The deflection of the Mississippi melt water at a rate of 0.011 Sv in our experiment was sufficient to switch off the meridional heat transport of the Atlantic within 200 years.

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6. REFERENCES

BAUMGARTNER,E. and E.REICHEL. 1975. Die Weltwasserbilanz/the World Water balance (bilingual). München, Wien.

BACASTOW, R. and E.MAIER-REIMER. 1989. Circulation model of the oceanic carbon cycle. Climate Dynamics, in press.

BERGER, W.H. and J.S.KILLINGLEY. 1982. The Worthington effect and the origin of the Younger Dryas. J. Mar. Res.40, Suppl., 27-38.

BROECKER,W.S., M.ANDREE, W.WOLFLI, H.OESCHGER, G.BONANI, J.KENNET, and D.PETEET. 1988. The chronology of the last deglaciation: implications to the cause of the Younger Dryas event. Paleoceanogr. 3, 1-19.

BROECKER, W.S., D.M.PETEET and D.RIND. 1985. Does the ocean atmosphere system have more than one stable mode of operation? Nature 315, 21-25.

BRYAN, F. 1986. High latitude salinity effects and interhemispheric thermohaline circulations. Nature 305, 301-304.

GORDON, A.L. 1986. Interocean exchange of thermocline water. J.Geoph. Res. 91, 5037-5046.

HASSELMANN, K. 1982. An ocean model for climate variability studies. Progr. Oceanogr. 11,69-92.

HASTENRATH, S. 1982. On meridional heat transports in the world ocean. J. Phys. Oceanogr. 12, 922-927.

HELLERMAN,S.and M.ROSENSTEIN. 1983. Normal monthly wind stress over the world ocean with error estimates. J.Phys.Oceanogr. 13, 1093-1104.

HUGHES,T.J., G.H.DENTON, B.C.ANDERSEN, D.H.SCHILLING, J.L.FASTOOK, and C.S.LINGLE. 1981. The last great ice sheets: a global view. in: G.H.Denton and T.S.Hughes (Eds.) *The Last Great Ice Sheets*. J.Wiley & Sons,N.Y. : 263-317.

LEVITUS, S. 1982. *Climatological Atlas of the World Ocean*. NOAA Professional Paper 13. Rockville Md.

MAIER-REIMER, E. and K.HASSELMANN. 1987. Transport and storage of CO2 in the ocean - an inorganic ocean-circulation carbon cycle model. Climate Dynamics 2, 63-90

MAIER-REIMER,E., U.MIKOLAJEWICZ, and K.HASSELMANN. 1989. On the sensitivity of the global ocean circulation to surface forcing and initial conditions. (In prep.)

ÖSTLUND, H.G. and M.STUIVER. 1980. GEOSECS Pacific radiocarbon. Radiocarbon 22, 25-53.

RUDDIMAN, W.F. and J.C.DUPLESSY. 1985. Conference on the last deglaciation: timing and mechanism. Quat.Res. 23, 1-17.

STEFAN, R. 1891. The theory of ice formation especially in the Arctic Ocean. Ann. Phys. Chem. N.F.42, 269-278.

STUIVER,M. and H.G.ÖSTLUND. 1980. GEOSECS Atlantic radiocarbon. Radiocarbon 22, 1-24.

WOODRUFF, S.D., R.J.SLUTZ, R.L.JENNE, and P.M.STEURER 1987. A comprehensive ocean-atmosphere data set. Bull.Am.Met.Soc. 68, 1239- 1250.