Response of IAP/ LASG GOALS Model to the Coupling of Air–Sea Fresh Water Exchange①

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(Received December 17, 1999; revised March 17, 2000)

ABSTRACT

The process of air–sea fresh water exchange is included successfully in the Global–Ocean–Atmosphere–Land–System model developed at the State Key Laboratory of Atmospheric Sciences and Geophysical Fluid Dynamics (LASG). The results of the coupled integration show that the climate drift has been controlled successfully. Analyses on the responses of ocean circulation to the changes of surface fresh water or salinity forcing show that the ocean spin–up stage under flux condition for salinity is the key to the implementation of air–sea fresh water flux coupling. This study also demonstrates that the Modified–Monthly–Flux–Anomaly coupling scheme (MMFA) brought forward by Yu and Zhang (1998) is suitable not only for daily air–sea heat flux coupling but also for daily fresh water flux coupling.

Key words: Fresh water flux, Air–sea coupling, Thermohaline circulation

1. Introduction

Water pervades all of the physical and dynamic structures within the Earth climate system through a myriad of hydrological processes. The hydrological cycle influences climate in a variety of ways, and there exists robust water flux exchange between ocean and atmosphere (Zhou et al., 1999). The exchange of fresh water between atmosphere and ocean through the air–sea interface is one of the least elements understood, but now is considered as one of the most important elements of the climate system, especially for ocean circulation changes on decadal to millennial time scales (Schmitt, 1995). At the present time, the air–sea coupled climate system model is a powerful tool in the community of climate change and climate variability studies. Consequently, the including of air–sea fresh water exchange (i.e. Evaporation minus Precipitation, hereafter referred to as E–P) in a coupled model is undoubtedly important for the improvement of the model in the capability of reproducing the actual climate system.

In nature, the process of air–sea coupling in a climate model is the interaction of model

①This work was co–supported by the National Key Project (Grant No.96–908–02–03), the Excellent National Key Laboratory Research Project (Grant No.49823002) and Chinese Academy of Sciences (CAS) under grant "Bai Ren Ji Hua" for "Validation of Coupled Climate Models".

atmosphere and model ocean at their interface. For the AGCM, it needs the forcing of SST (sea surface temperature) and sea ice from OGCM; for the OGCM, it needs the forcing of heat flux, momentum and fresh water flux from the AGCM. The coupling of air–sea fresh water flux is a process linked with the internal feedback mechanism of the model system. The oceanic general circulation models are sensitive to the forcing of fresh water at sea surface, a small disturbing of fresh water flux may be large enough to cause the overturning of the oceanic stratification, and leads to a tremendous change of the ocean circulation pattern. Thus, the coupling of air–sea fresh water flux in a climate model is an extremely difficult task but key problem at the forefront of climate research. The development of air–sea coupling skills involving the treatment of air–sea fresh water exchange is regarded as one of the most important problems that need to be dealt with in the near future by the community of the coupled climate system model research.

At the State Key Laboratory of Atmospheric Sciences and Geophysical Fluid Dynamics (LASG), Institute of Atmospheric Physics (IAP), great progress has been made in developing the Global–Ocean–Atmosphere–Land–System model (hereafter referred to as GOALS) (Wu et al., 1997). In the previous versions of GOALS, only momentum and heat fluxes' exchanges have been included in the coupling schemes, i.e. the fresh water flux is not exchanged between the ocean and the atmosphere in the coupling process. For the purpose of perfecting this process, the authors have made much effort in developing the air–sea fresh water coupling skill and solved the problem to some extent, though not completely. This paper is served as a preliminary summary of the work on this aspect and organized as follows: the treatment of surface fresh water forcing for single OGCM and the air–sea fresh water flux coupling procedure for IAP / LASG GOALS model are described in Section 2. The responses of the model, especially the oceanic circulation, to the changes of surface fresh water flux forcing are discussed in Section 3. Section 4 is a brief concluding remark.

2. The boundary conditions for salinity in OGCM and air–sea fresh water flux coupling procedure for GOALS model

Generally speaking, the Ekman pumping velocity is about 30 times larger than the vertical velocity associated with precipitation and evaporation in the world ocean (Huang, 1993). A simple scaling analysis indicates that the E−P driven barotropic mass fluxes are about 1 Sv① or less, which is only about a few percent of the wind–driven or the thermally driven circulation. Thus, the fresh water forced circulation used to be ignored. However, a close examination reveals that there is no role for the salinity in the aforementioned analyses. If there is no salinity difference between the rainwater and seawater, the former theory would be a perfect description for the E−P driven circulation. However, the E−P induced dilution / concentration of salt gives rise to a very strong and complicated baroclinic structure of the saline circulation. A small amount of precipitation, together with the role of vertical mixing, can drive a strong meridional circulation. Thus, for the nonce, it is believed that the fresh water forcing is as important as the thermal forcing in determining the ocean circulation, especially the thermohaline circulation. Accordingly, the treatment of air–sea fresh water forcing in OGCM or CGCM (Coupled General Circulation Model) is an important problem in climate numerical modeling community (Huang, 1993).

In oceanic general circulation models, it is appropriate for representing the fresh water

①Sv = 10^6 m^3 / s

flux at the ocean surface as a surface boundary condition on salinity. Letting θ, λ, z and t be co-latitude, longitude, depth and time respectively, the primitive forecast equation for salinity can be written as

$$\frac{dS}{dt} = A_{\mathrm{HH}} \Delta S + \frac{\partial}{\partial z}\left(A_{\mathrm{HV}} \frac{\partial S}{\partial z} \right) + c \ , \tag{1}$$

where A_{HH} and A_{HV} are the horizontal and vertical viscosity coefficients, respectively, c represents the vertical convection adjustment if unstable stratification occurs, and

$$\frac{d}{dt} = \frac{\partial}{\partial t} + \vec{v} \cdot \nabla + w\frac{\partial}{\partial z} \ ,$$

$$\nabla = \frac{\partial}{a\partial\theta} \vec{e}_\theta + \frac{\partial}{a\sin\theta\partial\lambda} \vec{e}_\lambda \ ,$$

$$\Delta = \frac{1}{a^2 \sin\theta} \frac{\partial}{\partial\theta} \sin\theta \frac{\partial}{\partial\theta} + \frac{1}{a^2 \sin^2\theta} \frac{\partial^2}{\partial\lambda^2} \ ,$$

$\vec{e}_\theta$ and $\vec{e}_\lambda$ are the unit vectors along θ and λ directions respectively. Other variables are conventional.

Equation (1) can be expressed in Eulerian format

$$\frac{\partial S}{\partial t} = -\left(\vec{v} \cdot \nabla S + w\frac{\partial S}{\partial z} \right) + A_{\mathrm{HH}} \Delta S + \frac{\partial}{\partial z}\left(A_{\mathrm{HV}} \frac{\partial S}{\partial z} \right) + c \ . \tag{2}$$

In Equation (2), the vertical diffusion term is linked directly to the fresh water forcing from the sea surface, and can be expressed in terms of

$$\frac{\partial}{\partial z}\left(A_{\mathrm{HV}} \frac{\partial S}{\partial z} \right)_{k=1} = \frac{1}{\Delta z}\left[A_{\mathrm{HV}} \left.\frac{\partial S}{\partial z}\right|_{k=\frac{1}{2}} - A_{\mathrm{HV}} \left.\frac{\partial S}{\partial z}\right|_{k=\frac{3}{2}} \right] , \tag{3}$$

$$A_{\mathrm{HV}} \left.\frac{\partial S}{\partial z}\right|_{k=\frac{1}{2}} = \frac{S_0}{\rho_0} \cdot \rho_0 \cdot (E - P) = S_0 (E - P) \ , \tag{4}$$

where ρ_0 and S_0 are the uppermost-layer reference density and reference salinity, respectively, the subscript k denotes the model layer, E and P denote evaporation and precipitation respectively, and the difference between evaporation and precipitation is the so-called air-sea fresh water flux.

In practice, for the convenience of reproducing the salinity distribution that fits observations at sea surface, the most commonly used boundary condition for salinity is a Newtonian relaxation condition. The air-sea fresh water flux can be written in terms of salinity as

$$E - P = \frac{\Delta Z}{T} \cdot \frac{1}{S_0} (S^* - S_1) = \frac{\mu}{S_0} (S^* - S_1) \ , \tag{5}$$

where $\mu = \frac{\Delta z}{T}$, Δz is the uppermost model layer, T is the restoring time coefficient, S^* is the reference salinity and the climatological mean sea surface salinity is often used for it, and S_1 is the uppermost model layer (sea surface) salinity. Thus, Eq.(4) can be expressed as

$$A_{\mathrm{HV}} \left.\frac{\partial S}{\partial z}\right|_{z=\frac{1}{2}} = \mu(S^{*} - S_1) \ . \tag{6}$$

Formula (6) is the most commonly used format of the Newtonian cooling type of restoring condition for salinity.

The so–called Newtonian type of restoring boundary condition was originally constructed for the forcing of atmospheric heat flux. The lag of SST behind the seasonal cycle of insolation is on the order of six weeks, and is conventionally parameterized in ocean models as a response to the changing atmospheric conditions. The dependence of longwave emission, sensible heating and atmospheric humidity (hence latent heat fluxes) on temperature allows the use of a simple, linear, Newtonian damping boundary condition, and the upper layer of the model ocean is restored to an appropriate reference temperature on a fast time scale (Haney, 1971).

While there is a great deal of evidence justifying the use of the linear restoring boundary condition for the temperature field, and there is less physical justification for the use of the condition for salinity. In the natural world, evaporation is mainly a function of air–sea temperature difference, while the distribution of precipitation depends upon complicated small and large–scale atmospheric processes. The use of Newtonian condition on salinity infers that the amount of precipitation or evaporation at any given place depends on the local sea surface salinity, which is clearly incorrect. Furthermore, the relaxation condition on salinity implies a definite time scale for the removal of salinity anomalies, which is not observed and is selected arbitrarily.

In ideal circumstances, one would like to use a fixed salinity flux, which is based on the observed evaporation and precipitation rate. This, however, is impractical since it has been argued that the convergence of numerical model salinity fields, when a flux boundary condition is used, is extremely slow (Bryan, 1986; Bryan et al., 1975). As a consequence, for the advantage of reproducing the ocean climate, the Newtonian condition on salinity is still used commonly. Nevertheless, the restoring condition on salinity is not suitable for climate change studies, since the variability of ocean circulation is too weak under the strong negative feedback. To make things even worse, the reference salinity S^{*} is unknown for climate conditions different from the current climate.

For the convenience of climate variability studies, the so–called flux condition for salinity is designed (Bryan, 1986; Weaver and Sarachik, 1991). The most commonly used procedure is spinning up the model ocean to a steady state by restoring the surface temperature and salinity to the current climatology, diagnosing the corresponding salt flux required to maintain the steady state, and then switching the restoring boundary condition on salinity to the diagnosed salt flux. Thus in further integration, this diagnosed surface salinity flux is used as a fixed–flux surface boundary condition. The flux condition can be written as

$$A_{\mathrm{HV}} \left.\frac{\partial S}{\partial z}\right|_{z=\frac{1}{2}} = \mu(S^{*} - S_1^{R}) \ . \tag{7}$$

The salt flux is often called "virtual salt flux", which can be expressed in terms of fresh water flux as,

$$A_{\mathrm{HV}} \left.\frac{\partial S}{\partial z}\right|_{z=\frac{1}{2}} = \mu(S^{*} - S_1^{R}) = \overline{S}(E - P) \ . \tag{8}$$

The diagnosed air–sea fresh water flux is given by

$$E-P=\frac{\mu(S^{*}-S_1^R)}{\bar{S}}\ , \tag{9}$$

where S_1^R is the surface salinity diagnosed from the quasi–equilibrium attained by parameterizing the upper boundary conditions for temperature and salinity in terms of relaxation conditions, $\bar{S}$ is the global mean surface salinity, usually taken as 34.7 psu or 35.0 psu. Strictly speaking, $\bar{S}$ should be the local salinity, but for the convenience of keeping global salinity balance, the local surface salinity is replaced by the globally averaged surface salinity.

This kind of flux condition for salinity can represent the actual surface fresh water flux to some extent. The rationale for this approach is that by spinning up the model using some specified climatological surface restoring fields, one can obtain an equilibrium in which the surface fields of temperature and salinity are climatologically correct. Theoretically speaking, the diagnosed fresh water flux should then yield the climatological sea surface salinity field.

For the coupling of air–sea fresh water exchange in numerical models, if one ran an OGCM under the restoring condition for salinity to a quasi–equilibrium state, and then coupled it directly to an AGCM, the great change of surface forcing would lead to the collapse of the ocean circulation. Since the implied surface fresh water forcing of the flux condition is similar to that provided by an active atmosphere (i.e. AGCM), a preferred method for conducting air–sea fresh water flux coupling is to run an OGCM under the flux condition for salinity to a quasi–equilibrium state, and then to couple it to an AGCM. Under this condition, the relative weaker variation of surface forcing may not be robust enough to cause the occurrence of climate drift in theory.

Upon switching from the restoring condition to the salt flux condition, however, ocean models displayed idiosyncratic behavior, ranging from a polar halocline catastrophe (Bryan, 1986), to " flush" (Weaver and Sarachik, 1991), low frequency oscillation, decadal and inter–decadal scale oscillation (Marotzke and Willebrand, 1989; Cai and Godfrey, 1995). The oceanic thermohaline circulation is found to be extremely sensitive to the perturbations of the salt flux: a small perturbation can cause a flip of the ocean circulation from equilibrium with North Atlantic Deep Water to one without. According to the study of Tziperman et al. (1994), the realistic OGCM solution is near the stability transition point with respect to the flux boundary condition. This proximity to the transition point allows the model to make a transition between the unstable and stable regimes induced by a relatively minor change in the surface fresh water flux and in the interior solution. Changing the salinity restoring time, which is used to obtain the steady model solution under restoring conditions, may induce such a change in the surface flux. Thus, the steady solution of OGCM under restoring conditions may be either stable or unstable upon transition to the salt flux boundary condition, depending on the magnitude of the salinity restoring time used to obtain this steady solution. The criterion for choosing the right values of the restoring coefficients is that, the restoring coefficients should be chosen such that at the steady state, the root–mean–square deviation between the model and observed surface fields is of the order of the combined observational error (e_{obs}) and model error (e_{mod}) (Tziperman et al., 1994). That is, the restoring coefficients for temperature and salinity should be chosen such that

$$\left\{\begin{array}{l}\sqrt{T_{\mathrm{ss,mod}}-T_{\mathrm{ss,obs}}}\\ \sqrt{S_{\mathrm{ss,mod}}-S_{\mathrm{ss,obs}}}\end{array}\right.\approx e_{\mathrm{obs}}+e_{\mathrm{mod}}\ , \tag{10}$$

where $T_{ss,mod}$ and $T_{ss,obs}$ are the model and observed SST, respectively, $S_{ss,mod}$ and $S_{ss,obs}$ are the model and observed SSS (sea surface salinity), respectively. The increased salinity restoring time produces seemingly more reasonable results for the fresh water flux, it does not cause too large deviations from the specified surface salinity, and results in a solution that seems stable upon transition to the flux boundary condition of salinity.

Thus, the air–sea fresh water coupling in GOALS is conducted in the following ways:

First, run the ocean component under the restoring boundary condition (that is, force the model toward sea surface salinity climatology) for 1700 years with a restoring time of 20 days, which can speed–up the convergence of the ocean.

Second, for the purpose of avoiding the collapse of thermohaline circulation, increase the salinity restoring time to 150 days and run the ocean model for further 8000 years to a steady state.

Third, diagnose the salt flux required to maintain the steady state of the 4700th model year of the restored spin–up and further integrate the ocean model for 2000 years under the diagnosed surface salinity flux.

Fourth, couple the OGCM with the atmospheric component from the 1960th model year of flux forcing spin–up stage by the so–called Modified–Monthly–Flux–Anomaly coupling scheme (MMFA) (Yu and Zhang, 1998), and then finish another 100 years long–term coupled integration. The daily coupling is conducted here instead of monthly coupling. The OGCM spin–up procedure for air–sea fresh water flux coupling is depicted in Fig. 1. Note that the Haney type formula for temperature is used throughout the ocean spin–up stage. Before conducting air–sea coupling, the equilibrated solution for the atmosphere has already been obtained by spinning up the AGCM for 20 years. During the air–sea coupling stage, the MMFA scheme is used not only for the fresh water flux but also for the heat and momentum fluxes. The couple frequency is daily coupling.

3. Analyses on the responses of GOALS model to the air–sea fresh water flux coupling

In order to examine the responses of ocean overturning circulation to the changes of

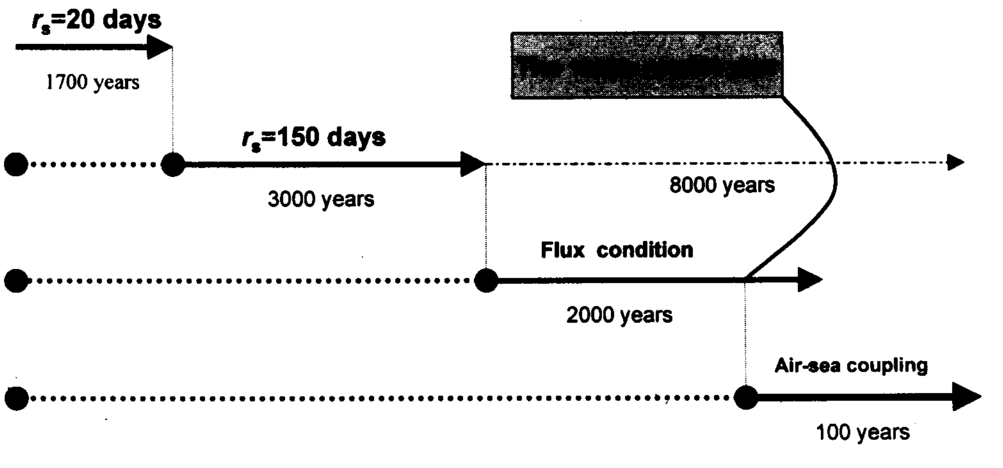

Fig. 1. The OGCM spin–up procedure for conducting air–sea fresh water flux coupling in IAP / LASG GOALS. r_s means the restoring coefficient for salinity.

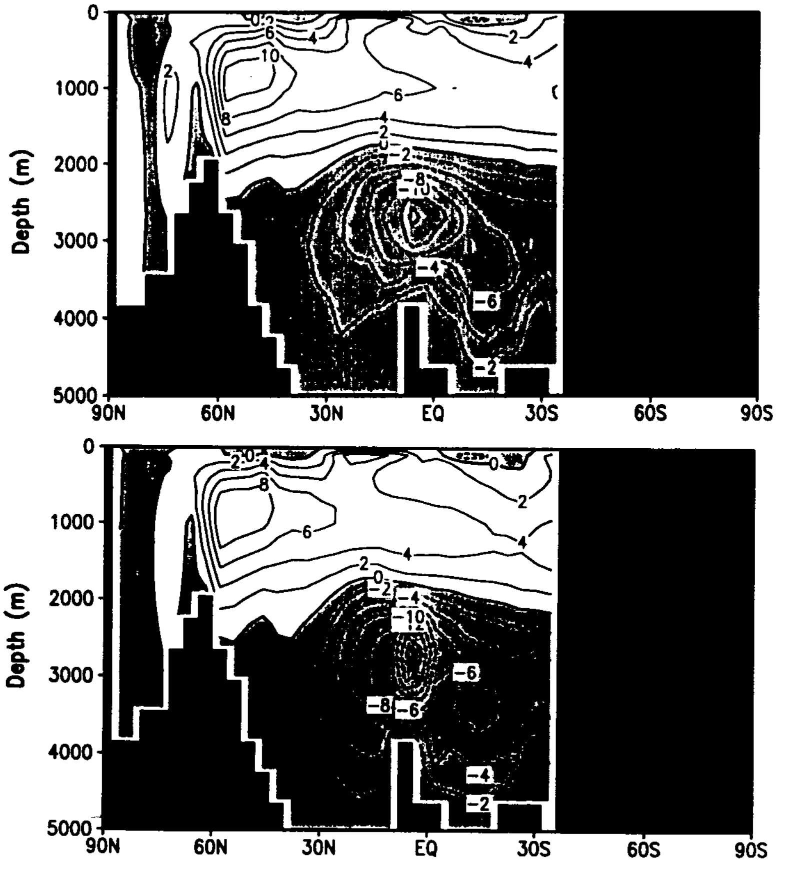

Fig. 2. The zonal mean Atlantic overturning stream functions (in Sv) for the 1700th (a) and 4700th (b) model years of the restored spin–up stage. From the 1700th year, the restoring coefficient for salinity is increased from 20 days to 150 days.

surface fresh water or salinity forcing, the zonal mean Atlantic meridional stream functions (in Sv) for the 1700th (a) and 4700th (b) model years under the restoring condition are given in Fig. 2. The time series of North Atlantic overturning circulation index (hereafter referred to as THC, which is defined as the maximum value of the down–welling branch around 60°N of Fig. 2) during the ocean spin–up stage and the air–sea fresh water coupling stage are shown in Fig. 3. The dominant feature in Fig. 2 is that when the constraint on salinity is relaxed, that is, the restoring coefficient for salinity is increased from 20 days to 150 days, the maximum value of down–welling branch of North Atlantic Deep Water (NADW) decreases by 2.0 Sv. In contrast, the strength of Antarctic Bottom Water intensifies. In the following 4000 years or so, as illustrated in Fig. 3a, the variation of NADW is weak and the model ocean comes up to an equilibrium with an intensity perturbation of less than 0.2 percent per thousand years.

Upon switching from the restoring forcing to flux condition, the North Atlantic meridional stream function undergoes a process of impetuous adjustment, but it does not collapse. The success in controlling stability transition under flux condition should be attributed to the selection of a relative larger salinity relaxation coefficient during the restored spin–up stage. As shown in Fig. 3b, the intensity of NADW first increases from 10.0 Sv to 21.0 Sv sharply, then decreases to 6.0 Sv in 100 years or so, and then once again increases to 20.0 Sv

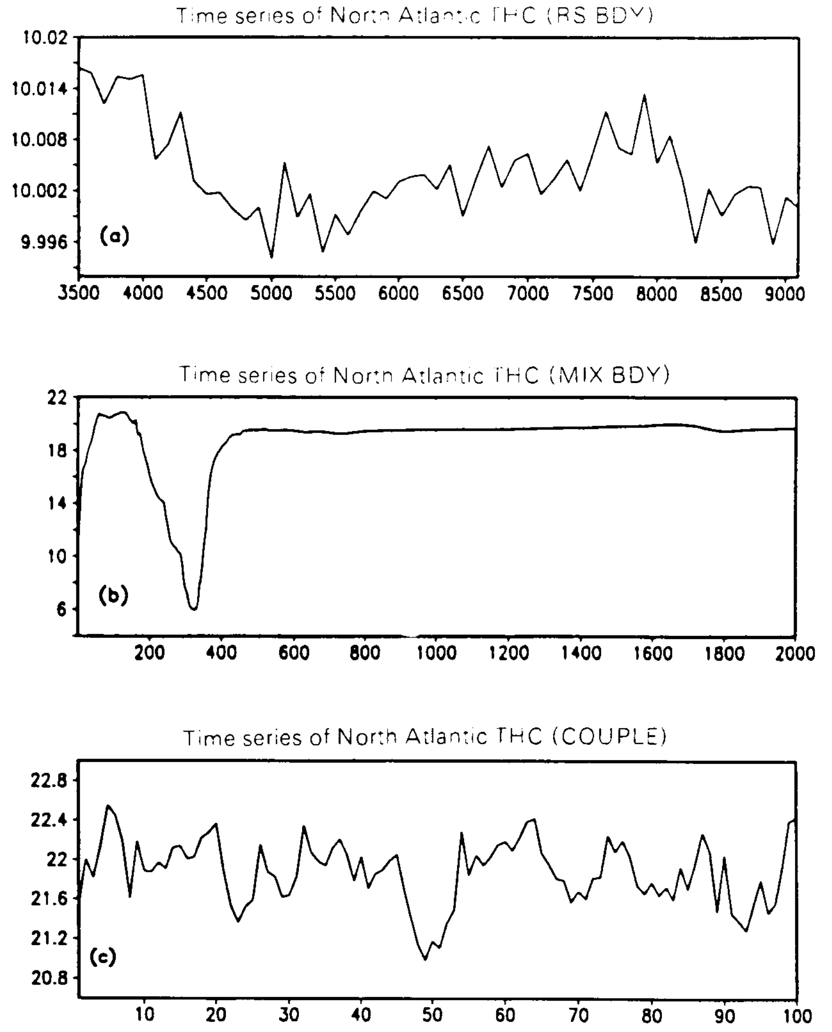

Fig. 3. Time series of the North Atlantic overturning circulation (The intensity index is defined as the maximum value of the down–welling branch around 60°N of the zonal mean stream function, units are in Sv). (a) Under restoring condition on salinity with a relaxation coefficient of 150 days, (b) under a flux condition on salinity, (c) after coupling with the atmosphere model.

in dozens of years. After the adjustment process of " amplifying–bating–amplifying" , the North Atlantic thermohaline circulation comes up to a near–steady state. In the following 1500 years, the intensity of THC maintains on the value of 20.0 Sv or so, which is a little stronger than the observed estimate of about 18.0–20.0 Sv (Boville and Gent, 1998). Thus, from the 1960th model year of spinup stage under the flux condition, an active atmosphere replaces the forcing of flux condition on salinity, that is, the ocean is coupled with the atmospheric model by using the MMFA scheme. Note that not only the fresh water flux but also the momentum and heat fluxes are included in the daily coupling.

As indicated in Fig. 3c, during the coupled 100 years integration, the strength of the North Atlantic overturning circulation is stable. The thermohaline circulation oscillates on interannual and decadal time scales with an annual mean intensity of 22.0 Sv or so. Thus, the

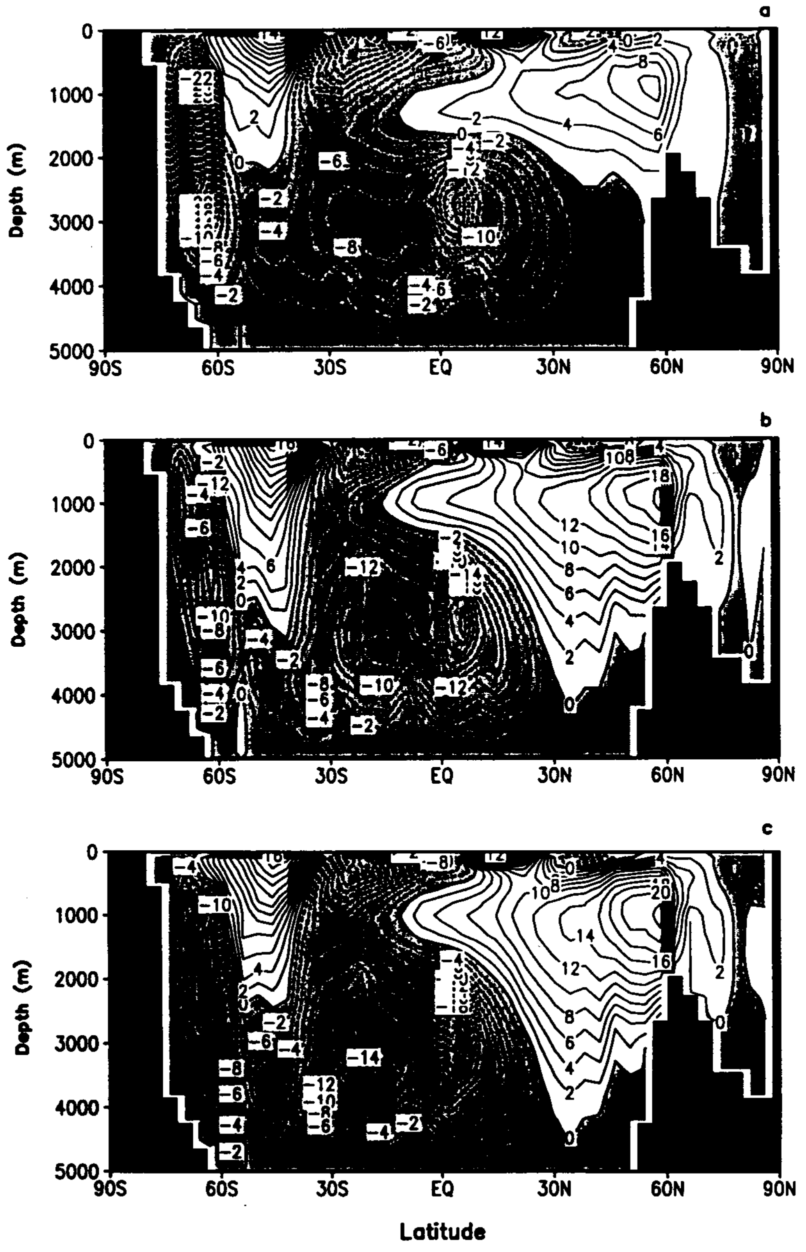

Fig. 4. Zonal Mean stream functions of the global ocean meridional overturning transport. (a) Model equilibrium under restoring condition, (b) mean of the last 50 years under flux condition, (c) mean of the last 50 years of the coupled experiment. The units are in Sv.

procedure designed for air–sea fresh water coupling of GOALS is fairly successful in controlling climate drift.

We can see from Fig. 3 that upon switching from restoring forcing on salinity to flux condition and then to that provided by an active atmosphere, the strength of the North Atlantic overturning circulation changes abruptly. Under restoring conditions, the maximum value of THC is 10.0 Sv. Under flux conditions, from the 1910th to 1960th model year, the 50 years mean intensity of THC is 20.0 Sv. In the coupled simulation, the mean intensity of THC for the last 50 years is 22.0 Sv. Accordingly, the strength of the North Atlantic thermohaline circulation under restoring conditions is far weaker than the observed estimation, while that under the flux condition or coupled simulation is somewhat stronger.

The zonal mean stream functions of the annual mean global ocean meridional overturning circulation for the restoring condition, the flux condition and the coupled simulation are shown in Fig. 4. The influences of surface forcing on the Antarctic Bottom Water (AABW), the sinking of water along the Antarctic continent and the Deacon cell are also apparent. Upon switching from restoring forcing to flux condition, the sinking of water along the Antarctic continent decreases from 32.0 Sv to 16.0 Sv, the bottom depth of the Deacon cell increases from 2000 m to near 3000 m, the center of AABW intensifies from 18.0 Sv to 22.0 Sv. Upon switching from the flux condition to an active atmosphere, both the sinking of water along the Antarctic continent and the intensity of AABW change mildly. Figure 4 exposes apparently that the sinking in the Northern Hemisphere (almost exclusively in the Atlantic) increases gradually from the restoring forcing to flux forcing and then to the coupled condition, which means the intensifying of the North Atlantic thermohaline circulation. The response of ocean circulation to the change of surface fresh water or salinity forcing is nearly the same as that of NCAR CSM (Bryan, 1998), though both the OGCM spin–up procedure and the coupling scheme are different from each other.

In essence, the variations of ocean circulation as noted above are responses of the model ocean to the changes of surface fresh water flux forcing. Under the flux condition, the inclusion of virtual air–sea fresh water flux leads to great changes of the distribution of surface salinity, which in turn causes the variation of meridional density gradient. The amplification of the North Atlantic overturning means an increasing of the meridional density gradient. From flux forcing to coupled condition, the response of the North Atlantic overturning is not the same drastic as that upon switching from restoring forcing to flux condition. This kind of difference may be partly due to the fact that the virtual air–sea fresh water flux under the flux condition is close to that provided by the active atmosphere, while the scheme of anomaly coupling may be another reason.

In order to verify the speculation mentioned above and show the similarity, the distributions of net fresh water flux implied by the salinity flux condition and that provided by the uncoupled atmosphere are depicted in Fig. 5. Also shown in Fig. 5c is the observed fresh water flux, which is denoted from the observational evaluation of Zhou et al. (1999). Negative value of E–P means a net gain of fresh water for the ocean, and vice versa. Visual inspections reveal obviously that the actual major features of the water cycle shown in Fig. 5c can also be seen clearly in Figs. 5a and 5b. For instance, in the tropics, rainfall dominates over evaporation within the Inter–tropical Convergence Zone (ITCZ). The subtropics is characterized by an excess of evaporation, except for the South Pacific Convergence Zone (SPCZ), a band of net precipitation trending to southeast away from the western equatorial Pacific. For the North Indian Ocean, evaporation dominates in the Arabian Sea, and precipitation in the Bay of Bengal. However, though the feature of net precipitation in sub–polar latitudes can be seen in Figs. 5a and 5b, there exist differences in the sub–polar North Atlantic, where the surface water downwellings and forms the North Atlantic Deep Water. The corresponding zonal profiles for Fig. 5 are shown in Fig. 6. We can see that though the zonal distribution of fresh water implied by flux condition is generally in qualitative agreement with that provided by the

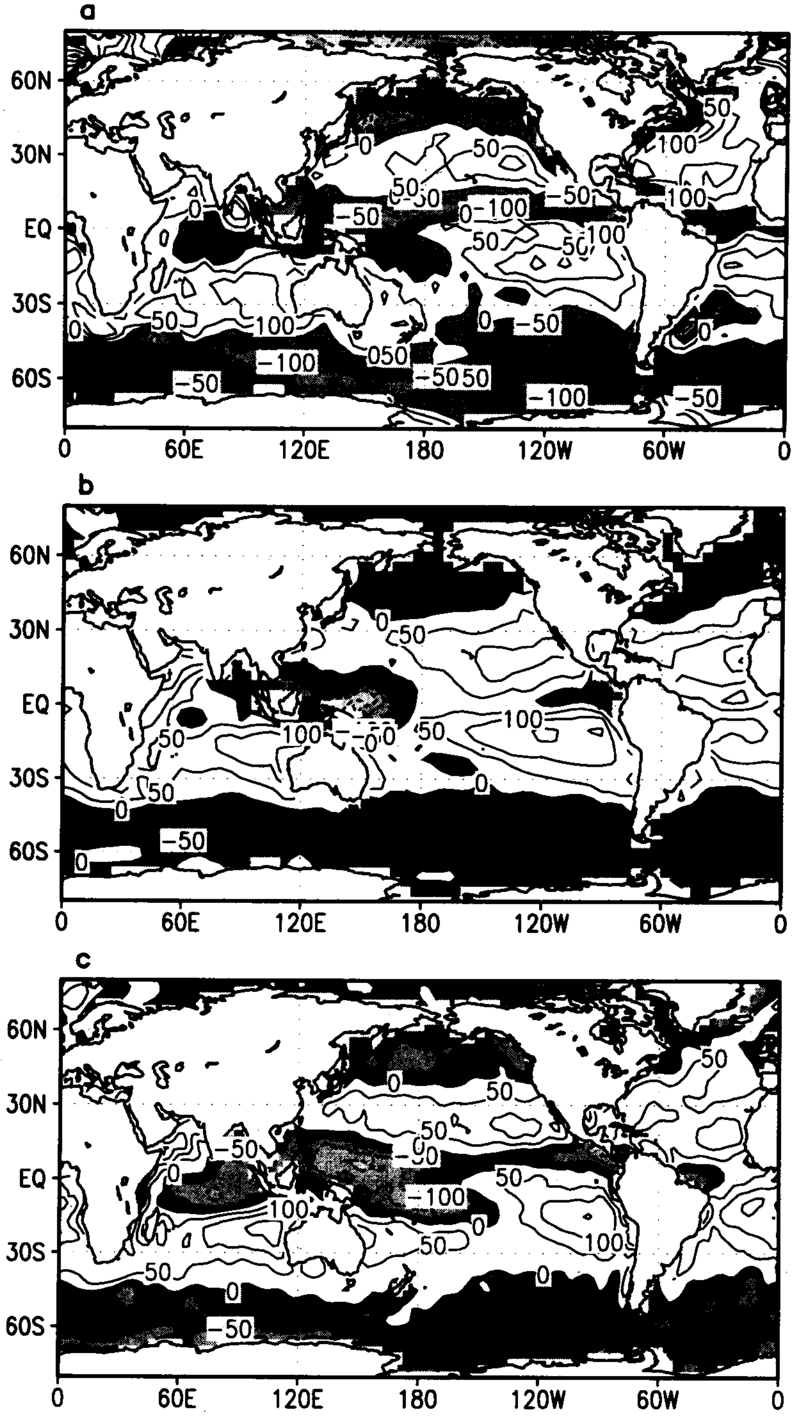

Fig. 5. Annual mean implied fresh water fluxes under flux condition (a), that provided by the uncoupled atmosphere (b) and the case for observation (c). (c) is denoted from the evaluation of Zhou et al. (1999). The units are in cm / a.

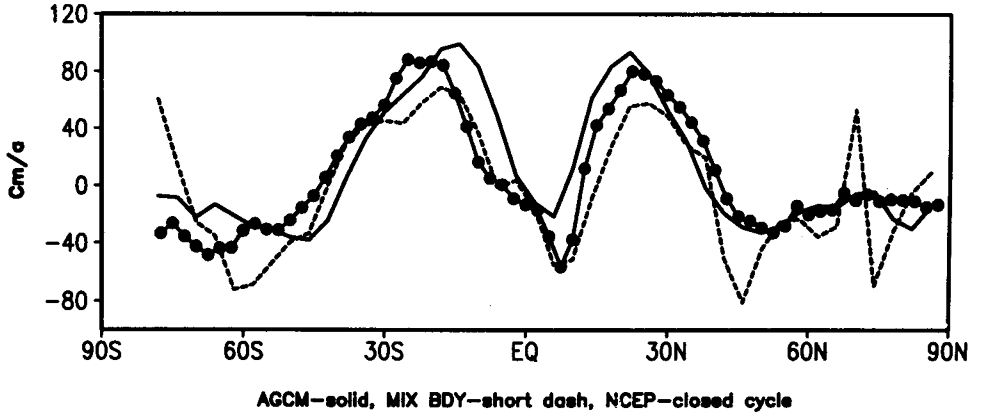

Fig. 6. Zonal profiles of annual mean fresh water flux of NCEP (closed cycles), under flux condition (short dash) and that provided by the uncoupled atmosphere (solid). The units are in cm / a.

uncoupled atmosphere, quantitative discrepancy does exist, especially in the high latitudes of both hemispheres. The impact of this kind of discrepancy on seawater density is inevitable, which will in turn affect the intensity of the thermohaline circulation.

From the time series of THC in the 100 years fully coupled integration shown in Fig. 3c, we can see our success in controlling the sea surface climate drift. This is only part of the case. The most commonly used criterion for detecting sea surface climate drift in coupled models is the globally averaged sea surface temperature. Accordingly, the time series of globally averaged monthly SST for 100 years coupled simulation is illustrated in Fig. 7a. Visual inspection shows that over the 100 years coupled integration, the globally averaged sea surface temperature is remarkably stable and exhibits no obvious drift trend. There are significant seasonal variations superimposed on the climatological mean state. The coupled simulation has strong variability on interannual and multiyear time scales, but no significant surface temperature trends.

The monthly Southern and Northern Hemisphere ice areas over the 100 years coupled integration are shown in Figs. 7b and 7c, respectively. The areas covered by sea ice in the Southern and Northern Hemisphere both nearly stabilize for the 100 years coupled simulation. The interannual and seasonal variability of sea–ice area is realistically simulated, and no climate drift exists. Since in the air–sea coupling process, the AGCM needs the forcing of sea ice from the OGCM, the steady–going sea–ice area provided by the OGCM in the coupled simulation contributes greatly to the success in controlling climate drift of AGCM.

4. Concluding remarks

The process of air–sea fresh water exchange is included successfully in the GOALS model, which is a great progress in the development of the fully coupled Global–Ocean–Atmosphere–Land–System model in IAP / LASG. Analyses on the responses of ocean circulation to the changes of surface fresh water or salinity forcing reveal that the ocean spin–up stage is the key to the implementation of air–sea fresh water flux coupling. Since the thermohaline circulation is sensitive to the change of surface forcing, the procedure for single OGCM spin–up is very important.

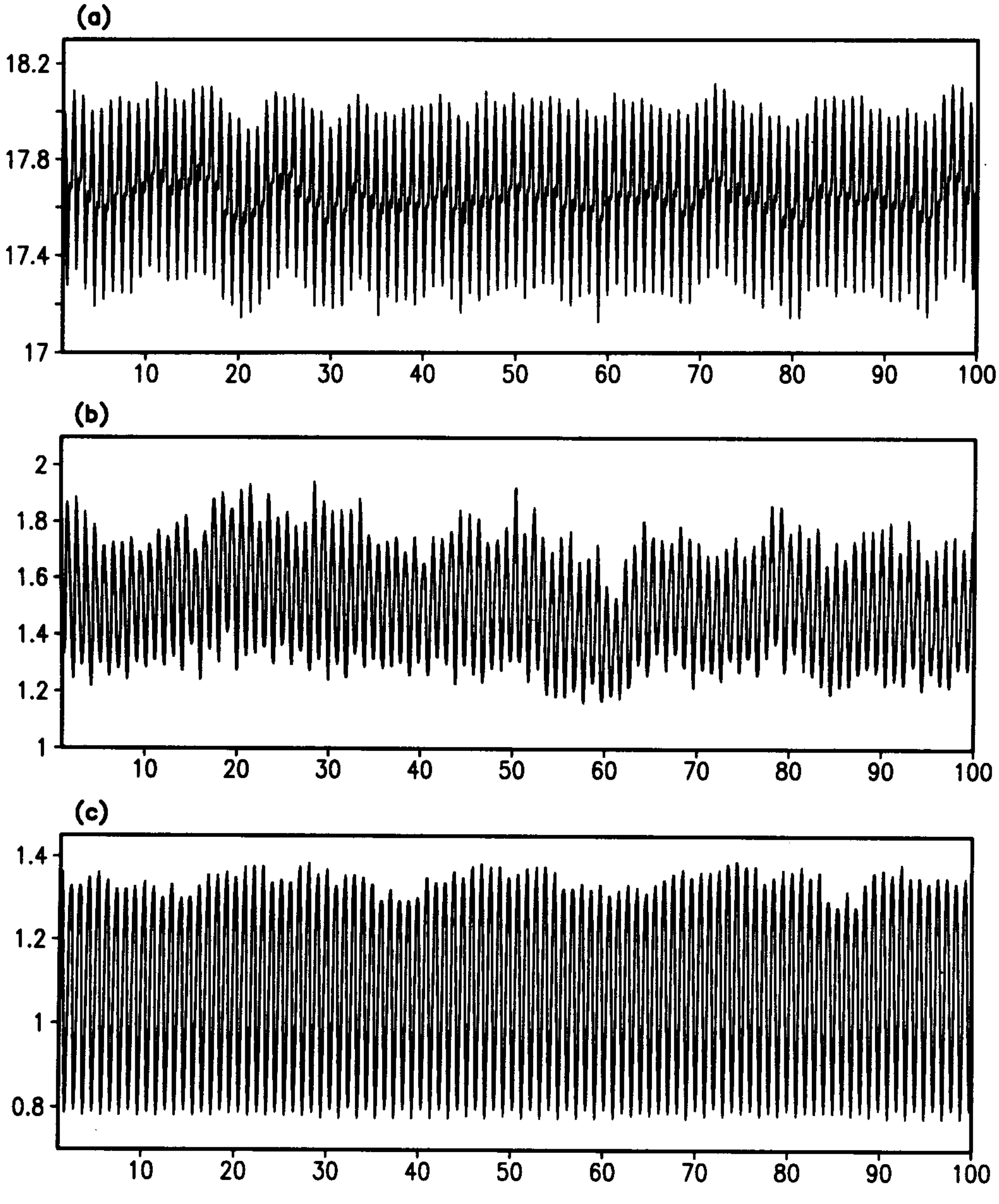

Fig. 7. Globally averaged monthly sea surface temperature (a, in °C), monthly Southern Hemisphere ice area (b) and monthly Northern Hemisphere ice area (c) (units are in 10^{13} m^2) for 100 years coupled simulation.

The essentials of the single OGCM spin–up procedure used in the present study are the stage under the flux condition for salinity. Since the virtual air–sea fresh water flux under the flux condition is close to that provided by the active atmosphere, the spin–up stage under the flux condition for salinity before conducting air–sea coupling is crucial to the successful control of climate drift in the coupled integration. Before the transition from the restoring condition to the flux condition for salinity, the spin–up with a relaxed restoring coefficient for salinity is helpful to the stability of the ocean circulation, and a relatively larger coefficient is preferred.

The present study also demonstrates that the Modified–Monthly–Flux–Anomaly coupling scheme (MMFA) brought forward by Yu and Zhang (1998) is suitable not only for heat flux coupling but also for fresh water flux coupling. In addition, it should be noted that although the process of air–sea fresh water flux has been included successfully in IAP / LASG GOALS, the role of runoff in the coupled system is another issue of the global hydrological cycle waiting for further investigations.

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