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THE SALINITY DRIFT OF HADCM3: DESCRIPTION, ANALYSIS AND A SCHEME TO CORRECT THE GLOBAL AVERAGE

by

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1. Introduction

In order to obtain correct solutions for the equations governing the ocean's motion and mixing processes a realistic density structure is required. In addition, extracting information on changes in sea-level caused by thermal expansion or contraction from rigid lid ocean models relies on an accurate treatment of the ocean's density structure. The density of sea-water is a non-linear function of both temperature and salinity (the mass of dissolved salt in a known mass of sea-water), therefore, to determine the density, both of these quantities must be adequately represented in ocean general circulation models.

The ocean component of HadCM3 is based on the formulation of Cox (1984). It does not have a free water surface, but is constrained by a surface pressure field which provides the ocean with a rigid lid. This technique has the advantage of filtering out high frequency surface waves which would otherwise require the model to use a shorter computational time step. The ocean currents are determined by solving the momentum equation on a grid of 1.25° latitude by 1.25° longitude on 20 vertical levels. The momentum equation contains terms which depend on the local density and pressure. In turn, the local density is a function of pressure, temperature and salinity (Apel, 1986). In the HadCM3 ocean, salinity is calculated by the solution of a prognostic salinity equation which consists of internal ocean terms which represent: advection, diffusion, mixed layer processes, Mediterranean mixing and ocean convection. In addition, the equation contains surface salinity flux terms which account for the flux of water between the ocean and atmosphere, and between the ocean and sea-ice. Riverine inputs are treated as an addition to the precipitation flux at selected river outflow points. Neither water mass nor salt mass are conserved directly by the ocean model.

Ocean salinity is increased by a net loss of water from the ocean to the atmosphere, that is a negative precipitation minus evaporation flux. It is decreased by a net gain of water by the ocean, that is a positive precipitation minus evaporation flux or a riverine input to the ocean. Freezing of sea-water, resulting in the creation of sea-ice, raises the salinity of the surrounding water as salt is liberated from the newly formed ice and returned to the ocean. Similarly, melting of sea-ice provides a source of water which has a lower salinity than the ocean in which it resides. In the ocean model of HadCM3, sea-ice is considered to have a salinity of 6 psu; this is less than the mean salinity of the oceans which is close to 35 psu. In addition to these processes, snow falling onto sea-ice will eventually melt providing a flux of fresh water. Some snow may be first converted to white ice as the weight of snow pushes the snow line below the water surface allowing water from the ocean to be incorporated into the snow, taking with it some of the salt from the surrounding water. White ice may eventually melt and lower the ocean salinity further. Assuming the internal oceanic processes conserve salinity, any drift in the volume integrated salinity of the ocean must be caused by the surface processes.

Preliminary climate experiments using the pre-HadCM3 model to simulate a stable pre-industrial control climate showed a sizeable, and unrealistic, positive drift in the volume integrated salinity of the ocean. The size of the volume integrated salinity drift in the pre-HadCM3 control run (**aasac**) was 2.4×10^7 psu m^3 s^{-1}. In terms of mean oceanic salinity this represents a rate of salinity increase of 5×10^{-4} psu yr^{-1}, or 0.12 psu over a period of 240 years. The change in water density caused by an increase of salinity of this size corresponds to an ocean mass increase of 1.3×10^{17} kg, or the equivalent of 37 cm of water spread over the entire ocean area. This is similar in magnitude to the predicted change in mean sea-level over the same period due to the thermal expansion of the oceans and the melting of glaciers, small ice caps and the Greenland ice sheet (IPCC, 1996; Lowe and Gregory, 1997).

In this report, reasons for the salinity drift of rigid lid oceans in general and the pre-HadCM3 ocean in particular are discussed and compared with the behaviour of the HadCM2 ocean. Subsequently, a method by which the HadCM3 drift can be removed is developed and tested.

2. Conserving Salinity in a Rigid Lid Ocean Model

The change in the mean salinity of a column of water of depth h due to a surface flux of fresh water is described by Equation 1, which follows directly from the definition of salinity (Gill, 1982). Here, S is the local salinity, q is the net water mass flux into the ocean and ρ is the local density of sea-water.

$$\frac{\partial S}{\partial t} = -\frac{Sq}{h\rho} \tag{1}$$

Consider a simple ocean model consisting of two separate columns of salt water with each column being well mixed, but with no direct connection between the two columns. The ocean surface is free so that water level in each column can change. For a small loss of fresh water (x) from one ocean column, Equation 1 predicts a rise in salinity for that column which is accompanied by a lowering of the water level. If the fresh water taken from the first column is added to the second column, the water level of the second column will rise and, according to Equation 1, the salinity of the second column will fall. If the volume integrated salinity (I) for the two column ocean were calculated using the free surface of both water columns it would have the same value before and after the water relocation, as shown in Equation box 2.

$$\begin{gathered}
dS_1 = \frac{S_1 x}{\rho_1 h_1} \qquad dh_1 = -\frac{x}{\rho_1} \\
dS_2 = \frac{-S_2 x}{\rho_2 h_2} \qquad dh_2 = \frac{x}{\rho_2} \\
I = S_1 h_1 + S_2 h_2 \\
dI = S_1 dh_1 + S_2 dh_2 + h_1 dS_1 + h_2 dS_2 \\
dI = \frac{-S_1 x}{\rho_1} + \frac{S_2 x}{\rho_2} + \frac{h_1 S_1 x}{\rho_1 h_1} - \frac{h_2 S_2 x}{\rho_2 h_2} \\
dI = 0
\end{gathered} \tag{2}$$

Next, consider what would happen if a rigid lid model were used so that the water level in each column remained fixed. The effect of moving a small amount of fresh water from one column to the other on the salinity of the columns could still be calculated using Equation 1, but the volume integrated salinity may change during the water relocation. The contribution to the volume integrated salinity made by the first column is too large because the local salinity is multiplied by too great a volume (the fall in water level in the free surface model has not been accounted for in the rigid lid case). Conversely, the contribution of the second column to the volume integrated salinity is too low because it uses a volume which is too small. These errors to the volume integrated salinity are of opposite sign, but, unless the initial salinities and densities of the two columns were identical, are different in magnitude. Therefore, in a rigid lid model the volume integrated salinity will be incorrect if surface processes effectively move fresh water from one location to another location which has a different salinity, as shown in Equation box 3.

$$dS_1=\frac{S_1 x}{\rho_1 h} \qquad dS_2=\frac{-S_2 x}{\rho_2 h}$$
$$I=S_1 h+S_2 h$$
$$dI=h(dS_1+dS_2) \tag{3}$$
$$dI=x\left(\frac{S_1}{\rho_1}-\frac{S_2}{\rho_2}\right)$$

Similar arguments apply when water is lost from a single column ocean at one time and returned to the column at a later time when the salinity of the column is different. The change in the volume integrated salinity of the ocean associated with moving water from one location to another, or taking it from the ocean at one point in time and returning it later when the ocean is at a different salinity will be referred to as *water relocation salinity drift*. Both Equations 2 and 3 neglect the second order effect of density changes resulting from the changes in salinity. In the HadCM3 ocean model a constant reference density is used in salinity flux calculations.

The loss of fresh water from the ocean system will also increase the volume integrated salinity; the same amount of salt is dissolved in less water. This will be referred to later as *water loss salinity drift*.

Although the water relocation and water loss salinity drifts have been considered for an extremely simple ocean model, the same problems occur in the more complex HadCM3 ocean. The volume integrated salinity drifts take place regardless of whether water is moved around by precipitation and evaporation, river runoff, or ice processes.

3. Quantitative Assessment of the pre-HadCM3 Salinity Drift

The mean rate of drift of the volume integrated salinity of the pre-HadCM3 ocean during the control run (**aasac**) was found to be approximately 2.39×10^{7} psu m^3 s^{-1}, with an interannual standard deviation of 1.2×10^{6} psu m^3 s^{-1}. The volume integrated salinity and rate of integrated salinity drift for this experiment are presented in Figures 1 and 2, respectively, for a 50 year period.

The volume integrated salinity budget for the HadCM3 ocean can be written in the form:-

Volume Integrated salinity drift=A+B **(4)**

where **A=Drift due to non-ice processes**
B=Drift due to ice processes

The rate of volume integrated salinity drift caused by non-ice processes is determined by integrating over the surface of the ocean the time mean of the product of precipitation plus river runoff minus evaporation (P+Y-E) and the local salinity (S) divided by the reference density. This quantity is output as stash code 30280. In experiment **aasac** the time mean rate of drift of volume integrated salinity due to non-ice processes was found to be approximately 2.97×10^{7} psu m^3 s^{-1} (or 124% of the total salinity drift). Salinity changes associated with ice processes (notably, the freezing and melting of sea-ice, the production of white ice and the melting of snow which falls in icy grid boxes) are output from HadCM3 model runs as stash code 144. Integrating these effects over the surface layer of the ocean provides the rate of volume integrated salinity drift due to ice-processes. In **aasac** the time mean drift of volume integrated salinity due to the ice processes was approximately -5.7×10^{6} psu m^3 s^{-1} (or -24% of the total drift), where the negative sign indicates that their net effect was to decrease the salinity.

The volume integrated salinity drift due to non-ice processes can be further split into water relocation and water loss terms. The water relocation contribution to the total integrated salinity drift can be investigated using the covariance of the salinity (S) and the water flux out of the ocean divided by the water density ($-q/\varrho$). A large positive covariance suggest that values of salinity greater than the mean salinity occur at similar times or positions as the values of water flux out of the ocean that are greater than its mean value. This condition leads to the relocation of fresh water from a region or time of higher salinity to a region or time of lower salinity, and, in a rigid lid model, causes a drift of the volume integrated salinity of the ocean. The larger the covariance term, the greater the salinity drift due to water relocation is likely to be. The covariance of the salinity and the water flux out of the ocean has both temporal and spatial components (Equations 5 and 6, respectively).

$$\langle covar_t(S,-\frac{q}{\rho})\rangle^x = \langle\langle\frac{-Sq}{\rho}\rangle^t\rangle^x - \langle\frac{\langle S\rangle^t . \langle -q\rangle^t}{\rho}\rangle^x \qquad (5)$$

$$\langle covar_x(S,-\frac{q}{\rho})\rangle^t = \langle\langle\frac{-Sq}{\rho}\rangle^x\rangle^t - \langle\frac{\langle S\rangle^x . \langle -q\rangle^x}{\rho}\rangle^t \qquad (6)$$

In Equations 5 and 6, $covar_x(S,-q/\rho)$ and $covar_t(S,-q/\rho)$ are the spatial and temporal covariances of S and $-q/\rho$. The notation $\langle\rangle^x$ and $\langle\rangle^t$ refers to averaging over space and time, respectively.

The time averages of salinity (S) and water flux out of the ocean (-q), together with the time average of the instantaneous product of S and q were used to estimate the volume integral of the temporal covariance of S and $-q/\rho$ (Equation 5 multiplied by the ocean volume), as 1.95×10^7 psu m^3 yr^{-1}. This is sizeable when compared with the total volume integrated salinity drift due to non-ice processes (making up 66% of the non-ice processes drift) and indicates that a large part of the salinity drift due to the non-ice processes can be explained by the water relocation process. Unfortunately, instantaneous values of S and q were not available with a sufficiently high temporal resolution for the model run analysed and so it was not possible to reliably estimate the temporal integral of the spatial covariance (Equation 6).

The remainder of the drift associated with non-ice processes was caused by water which was lost from the ocean and either stored on land, discarded or returned to the ocean by an alternative route. In pre-HadCM3 run **aasac**, approximately 8.5×10^7 kg s^{-1} of water was found to accumulate on the Antarctic and Greenland ice sheets, with Antarctic receiving 80% of the total. These accumulation rates compare closely with IPCC 1990 estimates of 7.0×10^7 kg s^{-1} for Antarctica and 1.7×10^7 kg s^{-1} for Greenland (Trenberth, 1995). In reality this accumulation would be largely offset by calving of icebergs from the ice sheets. However, this process is not currently included in the model. The volume integrated salinity drift associated with this storage of water on the ice sheets is estimated using Equation 1 to be approximately $3x10^6$ psu m^3 s^{-1} (10% of the non-ice processes drift). A further $4.2x10^7$ kg s^{-1} of water was lost from the ocean during **aasac** as discarded runoff, corresponding to a further volume integrated salinity drift of $1.5x10^6$ psu m^3 s^{-1} (5% of the non-ice processes drift). This term arises because, in HadCM3, runoff not occurring within the collection areas of rivers with ocean outflow points is simply discarded. The final component of the drift due to non-ice processes, $5.7x10^6$ psu m^3 s^{-1} (20% of the non-ice processes drift), results from snow which accumulates on sea-ice rather than precipitating directly into the ocean. Once on the sea-ice, the fate of this snow is determined by ice processes and the volume integrated salinity drift associated with the melting of the snow is then included as a negative contribution in the ice processes integrated salinity drift, a term which has a magnitude of $-5.7x10^6$ psu m^3 s^{-1}.

Averaging over periods of a decade or more, the volume of sea-ice in the spun up pre-HadCM3 control run **aasac** remained approximately constant, suggesting that all of the snow accumulating on sea-ice eventually melts. The salinity drift associated with ice processes was approximately equal in magnitude to the drift due to the water temporarily lost from the ocean as snow accumulation on sea-ice (a term in the non-ice processes drift) but of opposite in sign. The relocation of water by the ice processes had little effect on the total volume integrated salinity drift.

3.1 Why the Salinity Drift was not present in HadCM2 results

The volume integrated salinity drift of the HadCM2 pre-industrial control run was much lower than the equivalent pre-HadCM3 run, less than 1×10^6 psu m^3 s^{-1} compared with 2.4×10^7 psu m^3 s^{-1}. In HadCM2 the salinity drift was offset by the water flux correction term used to maintain a realistic climatology. Unlike HadCM2, the more recent HadCM3 model is not flux corrected. The volume integrated salinity drift in HadCM2 due to non-ice processes (excluding the flux correction term) was approximately 2.5×10^7 psu m^3 s^{-1}, compared to a value of -7.1×10^6 psu m^3 s^{-1} for ice processes. The drift due to non-ice processes comprised a water relocation contribution of approximately 1.1×10^7 psu m^3 s^{-1} and a water loss contribution of approximately 1.4×10^7 psu m^3 s^{-1}. However, the flux correction resulted in a negative contribution to the volume integrated salinity drift of approximately -1.8×10^7 psu m^3 s^{-1}, causing the total drift in volume integrated salinity to be small. The differences in salinity drifts of the pre-HadCM3 and the HadCM2 control runs are presented in Table 1.

Table 1: A comparison of the volume integrated salinity drifts of the pre-HadCM3 and the HadCM2 control runs (all values are in units of 1×10^6 psu m^3 s^{-1} and given to 2 significant figures).

Drift Component	HadCM2	pre-HadCM3
Total Drift	<1.0	24
Non-Ice processes (excluding flux correction)	25	30
Flux Correction	-18	0.0
Ice processes	-7.1	-5.7

Figure 1: HADCM3 (aasac) mean oceanic salinity (psu)

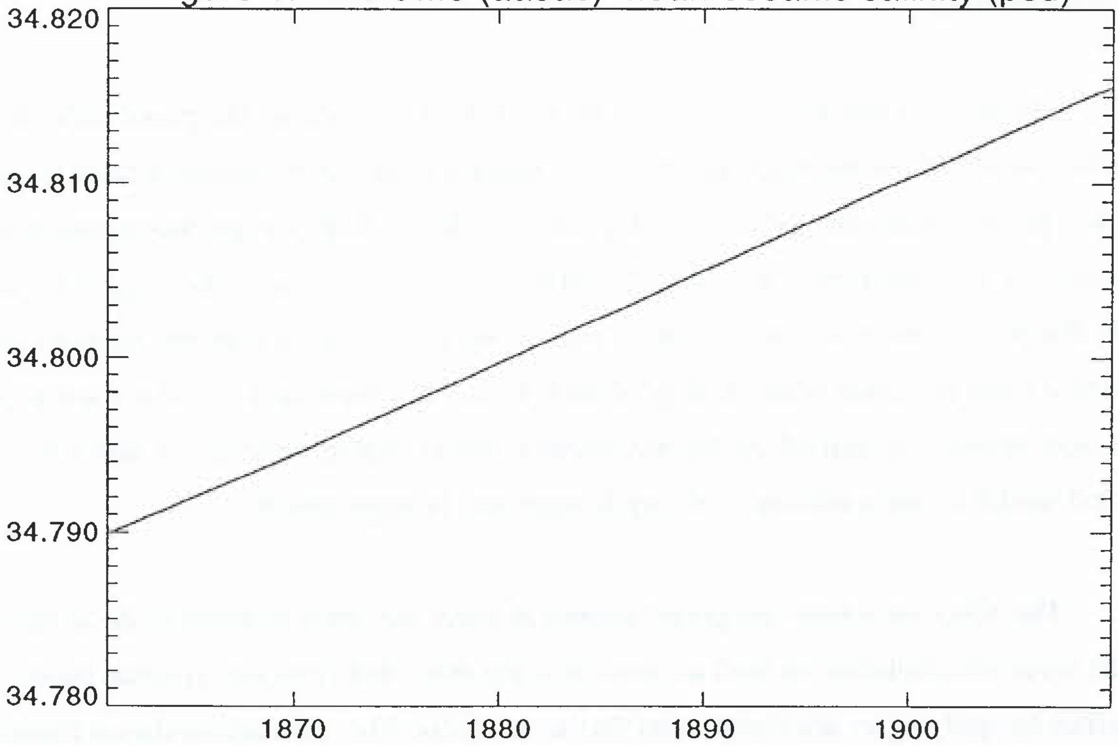
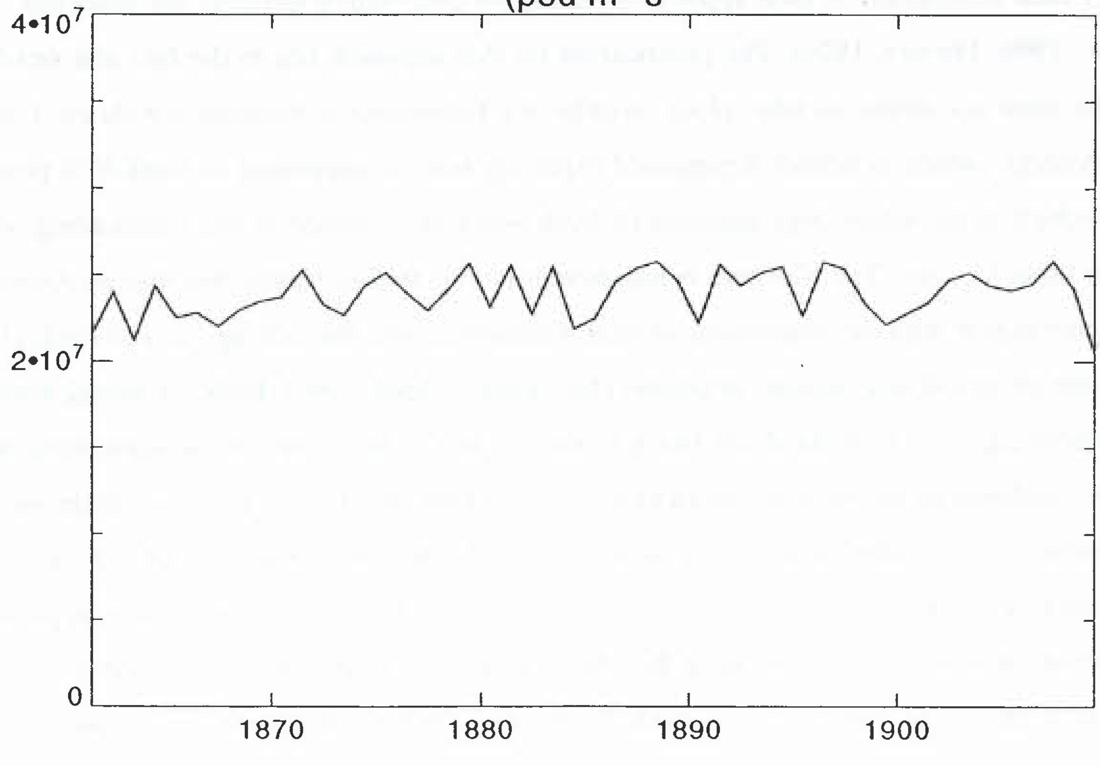


Figure 2: Rate of salinity drift in HADCM3 (aasac) (psu m^3 s^{-1})



4. Techniques for Eliminating the Volume Integrated Salinity Drift of HadCM3

4.1 Methodology

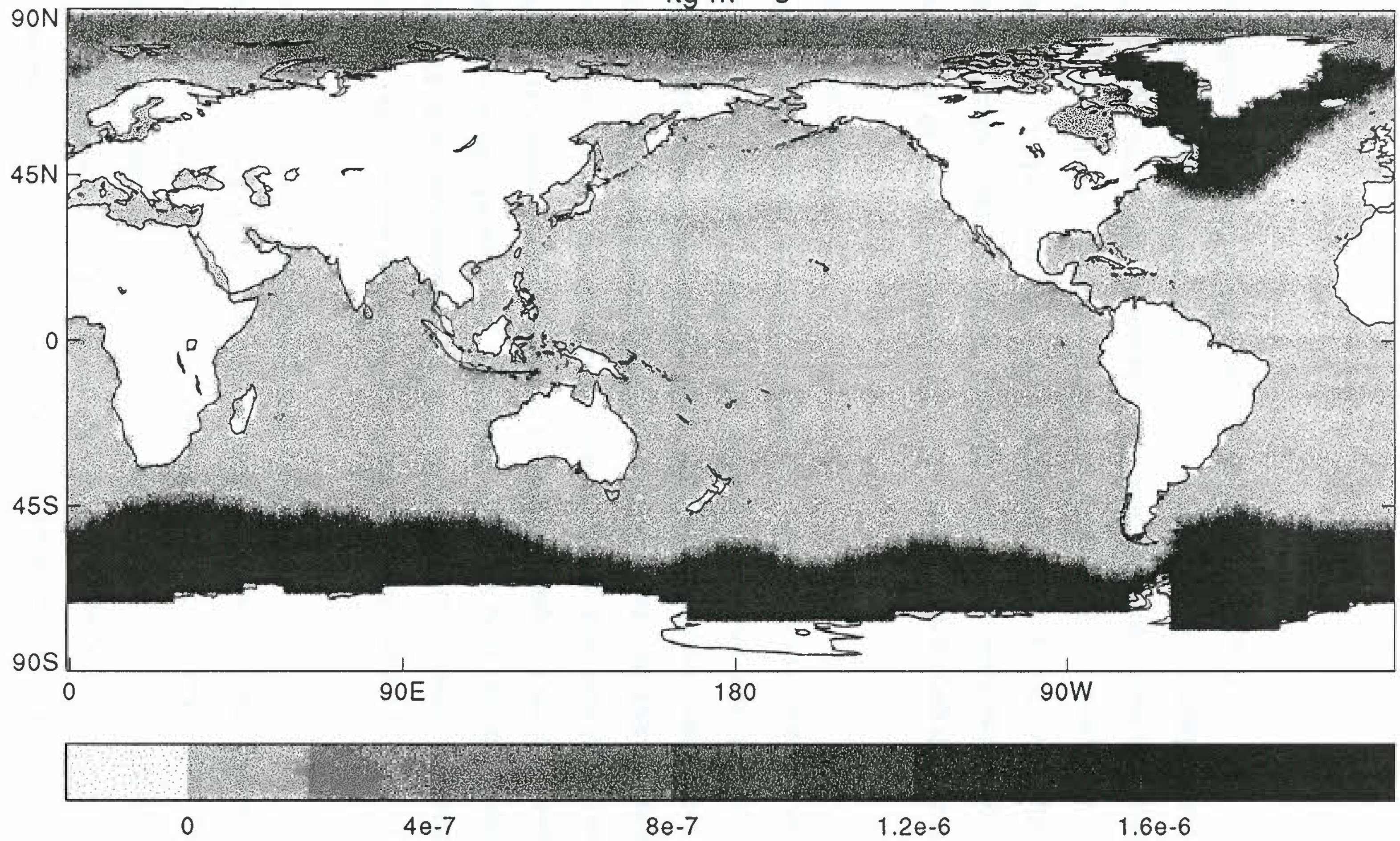
The water relocation (covariance) component of the volume integrated salinity drift can be removed from the HadCM3 model by using a fixed ocean reference salinity for all surface process that alter salinity via Equation 1, that is both ice processes and non-ice processes. This is equivalent to setting S_1 and S_2 equal to a reference value, S_{ref}, in Equation 3. In this work, a reference salinity of 35 psu has been used. As in previous versions of the model, a fixed reference ocean density of 1000 kg m^{-3} has been used for all surface process that alter salinity. A unified model modification set to enable version 4.3 and 4.4 of the unified model to use a reference salinity is described in Appendix A.

The effect on volume integrated salinity of water not being returned to the ocean, that is the water accumulating on land as snow or being discarded from non-riverine basins, can be offset by applying an additional water flux to the ocean. The snow accumulation correction term is separated into Northern hemisphere and Southern hemisphere components. The water from each hemisphere is then applied in the region over which icebergs are observed (Bigg *et al.*, 1996; Harvey, 1976). The justification for this approach lies in the fact that nearly all of the snow accumulation takes place on either the Greenland or Antarctic ice sheets. Calving of icebergs (which is neither represented explicitly nor parameterised in HadCM3) provides a mechanism by which large amounts of fresh water are returned to the surrounding waters from these ice caps. The Northern hemisphere region is further subdivided into an Arctic and Atlantic region with the proportions of melt determined from the calving rates of Reeh (1984) and the observed local ocean circulation (for instance Apel, 1986). Icebergs calved from the Northern edges of Greenland are likely to remain in the Arctic ocean for some time, while those produced on the more southern parts of Greenland are likely to travel south through the Greenland sea and/or Baffin bay into the north Atlantic. Within each of the snow accumulation correction regions the correction is constant over most of the area of application. However, to avoid discontinuities at the edges of these areas, it is linearly scaled to zero over a region of 5 grid boxes at its oceanic boundaries, in both the north-south and east-west directions. The discarded runoff correction term is applied as a globally average water flux. The combined snow accumulation and discarded runoff water flux field is illustrated in Figure 3. The method by which the water correction field was applied in versions 4.3 and 4.4 of the unified model is described in appendix B. The impact of using a more realistic iceberg melt field for the northern hemisphere, derived from an explicit iceberg model and scaled to the amount of water we must return to the ocean, has also been considered and is discussed in more detail in appendix C. It was found that this physically more realistic correction did not significantly alter the north Atlantic overturning stream function.

The water flux fields are not time varying. While this is acceptable for the control climate, the amount of snow accumulating on land and/or discarded runoff may alter in perturbed climate experiments. In turn, this introduces a salinity drift into the perturbed climate results. Part of this salinity drift may be due to realistic changes in the mass balance of the ice sheets, however, the use of a temporally constant calving term may lead to errors because the calving rate and iceberg melt distribution should depend on past and present climate state. Furthermore, even if calculated changes in the mass balance were correct, the change in volume integrated salinity caused by the changes in ice sheet mass balance may still be incorrect because changes in ocean volume are not taken into account in the calculation of the volume integral.

A more permanent fix for the discarded runoff term may involve placing the discarded runoff into a model reservoir such as canopy moisture. The excess moisture could then evaporate in a physically realistic manner, and, presumably, some would eventually be returned to the ocean. Alternatively, a scheme is being considered where all land points result in runoff to either the ocean or represented inland seas so that non of the runoff is discarded. An improvement to the scheme which accounts for the calving of icebergs would be to link the pattern and magnitude of the melt distribution to the state of the climate.

Figure 3: Water flux correction field
kg m^{-2} s^{-1}



4.2 Results and Discussion

The methods used to correct the drift of volume integrated salinity present in the pre-HadCM3 simulations, as described in section 4.1, have been applied in a test simulation (**aayfa**). The time mean volume integrated salinity drift of this corrected run has been reduced from $2.4\text{x}10^7$ psu m^3 s^{-1} to less than $1\text{x}10^6$ psu m^3 s^{-1}. The interannual standard deviation of the rate of integrated salinity drift was found to be approximately $1\text{x}10^6$ psu m^3 s^{-1} in the test run; this is similar in magnitude to that of the original pre-HadCM3 control run.

The salinity drift due to non-ice processes has been reduced from $2.94\text{x}10^7$ psu m^3 s^{-1} to $7.6\text{x}10^6$ psu m^3 s^{-1}. Of this, the time average of the water relocation drift and the drift due to water permanently lost from the ocean have been eliminated. The salinity drift due to water lost from the ocean but returned by ice processes has been increased by approximately $2\text{x}10^6$ psu m^3 s^{-1}. However, this increase is offset by a larger negative salinity drift associated with ice processes. The slight increase in magnitude of the drift associated with ice processes is due to the use of a reference salinity which is larger than the mean salinity in the cold high latitude waters where the melting of snow accumulated on sea-ice occurs.

Possible side effects of the changes to the model have been investigated by comparing the oceanic temperature and salinity in the upper layer of the ocean and the maximum northern hemisphere meridional overturning stream function for the trial run (**aayfa**) with an uncorrected control run (**abaea**). The differences between the spatial pattern of the upper layer ocean temperatures and salinities (averaged over a 10 year period at the end of the 60 year trial) and the observed Levitus climatology (Levitus *et al.*, 1994) are shown in Figures 4 and 5 for runs **abaea** and **aayfa**. In most regions of the ocean the Figures are similar, indicating that the changes introduced to correct the salinity drift have had little effect on the upper ocean temperatures or salinities over the trial period.

The time series of northern hemisphere meridional overturning stream functions is shown in Figure 6. During the first decade of the test runs the meridional overturning stream function is similar in both corrected and uncorrected runs. Later (approximately 15 years in to the trial) the stream functions begin to diverge with the corrected simulation having a weaker overturning stream function than the uncorrected run. This suggests that the method of removing the volume integrated salinity drift results in a slight reduction of the thermohaline circulation. Towards the end of the trial period there is some convergence of the results from the two simulations. The reduction in overturning stream function is a favourable change to the results, but still exceeds the value of -16 Sv estimated from observations.

A problem not corrected by the current scheme is the drift in salinity of individual ocean basins. While the global ocean now conserves volume integrated salinity, individual basins are still free to drift if the distribution of precipitation into and evaporation out of the ocean are not correct. This is largest for isolated basins, such as the Caspain sea. If, in the future, flux correction is not to be used to correct this, a better representation of the precipitation and evaporation fluxes will be required.

Figure 4a: Difference between non-corrected run (ABAEA) and Levitus upper layer temperature (^{o}C)

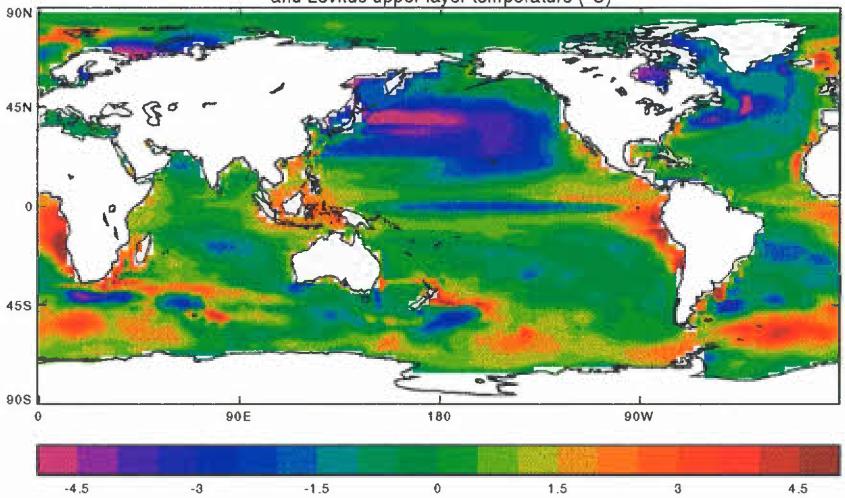


Figure 4b: Difference between corrected run (AAYFA) and Levitus upper layer temperature (^{o}C)

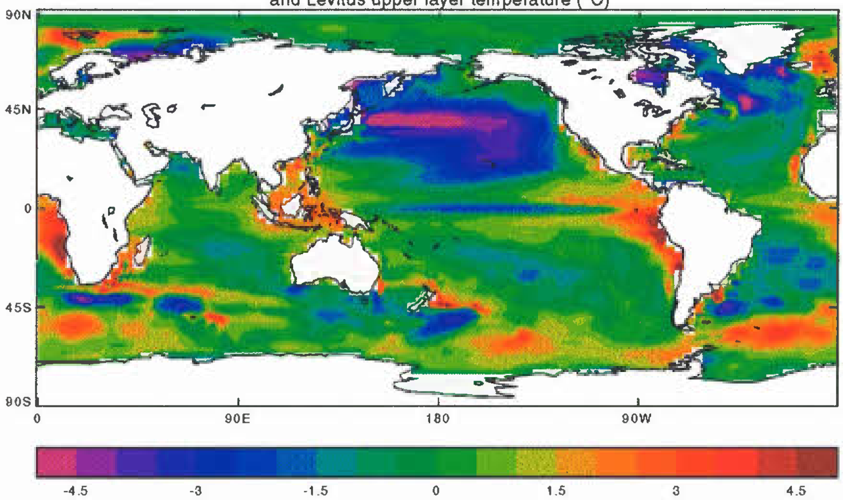


Figure 4c: ABAEA minus AAYFA temperature (^{o}C)

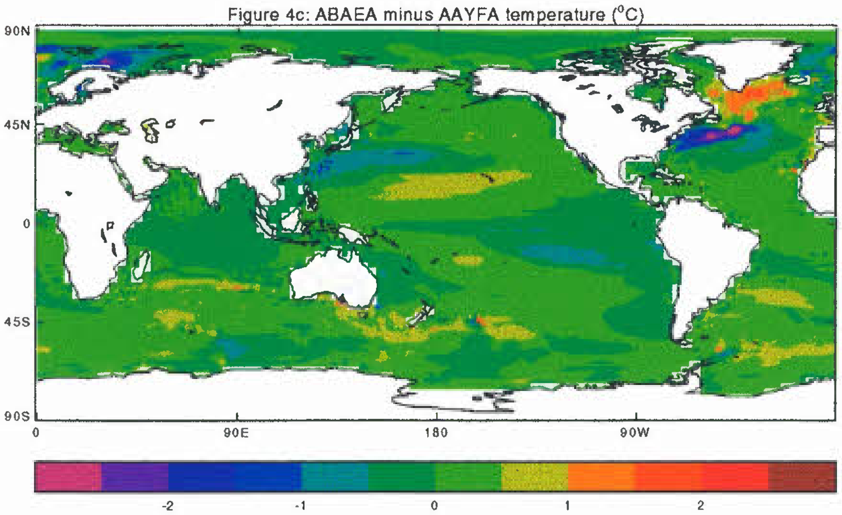


Figure 5a: Difference between non-corrected run (ABAEA) and Levitus upper layer Salinity (psu)

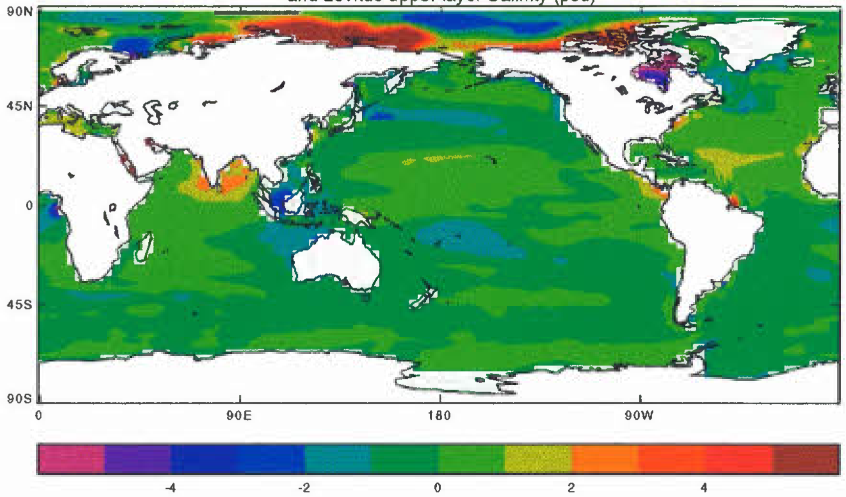


Figure 5b: Difference between corrected run (AAYFA) and Levitus upper layer salinity (psu)

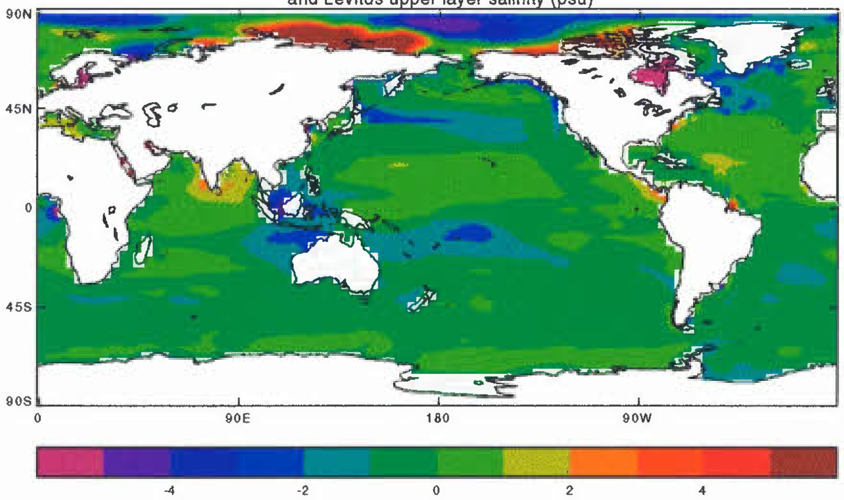


Figure 5c: ABAEA minus AAYFA salinity (psu)

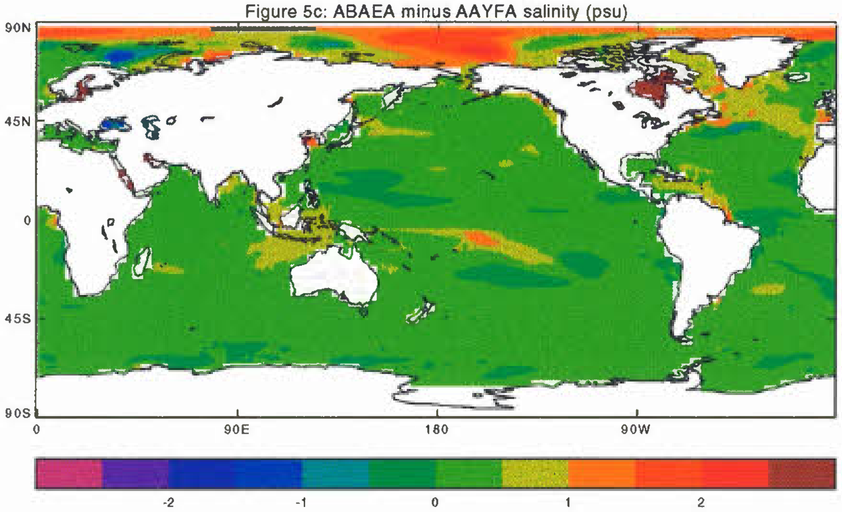
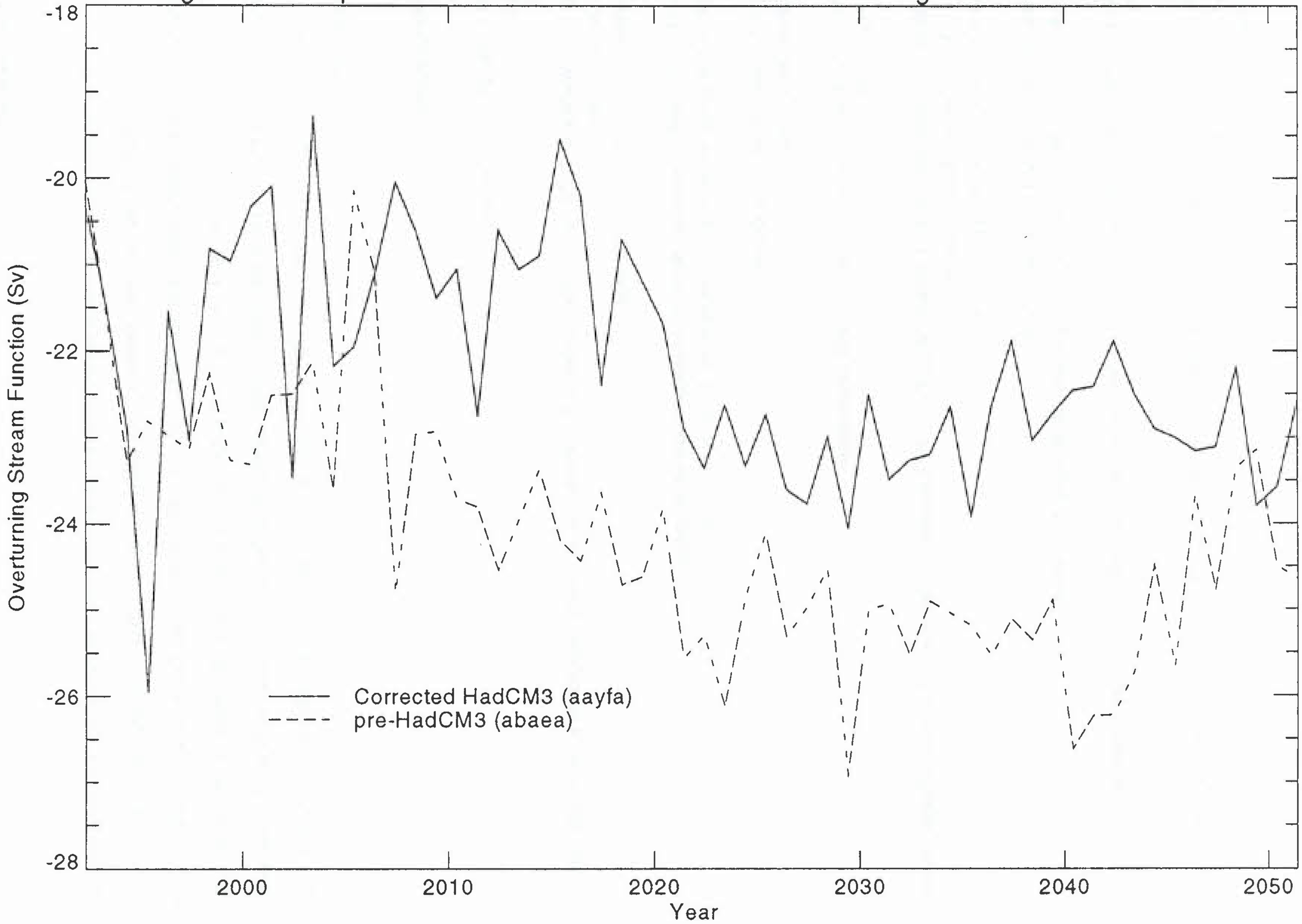


Figure 6: Comparison of maximum north Atlantic overturning stream function values



5. Conclusion

The sizeable drift in volume integrated ocean salinity seen in pre-HadCM3 simulations has been described. A method has been put forward to remove this drift and trial simulations conducted. These trials show that the mean rate of salinity drift has been reduced to less than 1% of its initial value, and that over a time scale of approximately 50 years detrimental side effects are not evident. Dealing with salinity drifts in individual basins, such as the Atlantic, is a longer term problem and is likely to remain a major issue in the absence of water flux corrections.

6. References

Apel J. The physics of oceans
Academic Press, 1986

Bigg G R. Wadley M R, Stevens D P and Johnson J A. Prediction of iceberg trajectories in the North Atlantic and Arctic Oceans.
Geophys Res Lett, 23, 3587-3590, 1996

Cox M D. A primitive equation, three dimensional model of the ocean.
GFDL ocean group Techn Rep 1, Princeton NJ, USA, pp132, 1984

Gill A E. Atmosphere-ocean dynamics.
Academic Press, 1982

Harvey J G. Atmosphere and ocean: our fluid environment.
Artemis, pp143, 1976

Houghton J T, Meira Filho L G, Callender B A, Harris N, Kattenberg A and Maskell K (eds): Climate change 1995, The science of climate change.
Cambridge University Press, pp572, 1996.

Levitus and Boyer. World ocean atlas 1994.
NOAA Prof Paper No 13. US Gov Printing office. Washington D C. 1994

Lowe J A and Gregory J M.Changes in Mean Sea-Level Under Several Future Climate Scenarios.
DETR report, Oct 1997

Reeh N. "Greenland ice-sheet mass balance and sea-level change" in glaciers, ice sheets and sea-level: effects of a CO2 induced climate change.
US DoE, 1984

Trenberth K. "Climate system modelling"
Cambridge University Press, 1993

Appendix A

The modification set used in version 4.3 and 4.4 of the unified model to ensure a reference salinity is used in surface processes is listed below

```
*IDENT fixsal
*/ Use constant reference salinity of 35 ppt for application of surface
*/ water fluxes
*DECLARE SFCADD
*D ORH1F305.421
      TA(i,1,2)=TA(i,1,2)+con_salt*SFLUX(i)*0.035
*D ORH1F305.428
      TA(i,1,2)=TA(i,1,2)+con_salt*fluxcorw(I)*0.035
*DECLARE TRACER
*D OJG2F403.22
      diagsw(i,j)=1e9*(diagsw(i,j)*0.035)/rho_water_si
*DECLARE ICEBNDS
*D ICEBNDS.277
   &    +CONST1*(0.035-SALICE )*DELH(I,J)
*D ICEBNDS.278
   &    +0.035*(CONST4*SUBLIM(I,J)-CONST2*SNOWMELT(I,J))
*DECLARE ICEUPDT
*D ICEUPDT.181
   &          +CONST1*(0.035-SALICE )*DELH(I,J)
*D ICEUPDT.182
   +          +CONST4*SUBLIM(I,J)*0.035
*D ICEUPDT.199
        SNOWMELT(I,J)=CONST2*SNOWMELT(I,J)*0.035
*D HNYCAL.168
       if (L_REFSAL) then
        ANOM_SALT(I)=ANOM_SALT(I)/0.035
       endif
```

A more sophisticated modification is being included in version 4.5 of the unified model which allows the changes to the model to be controlled using options in the UMUI. In addition, the newer modification will allow limits to be placed on the maximum and minimum values of salinity permitted in the model.

Appendix B

The water flux correction can be applied in the following manner

(1) Create a pp file of water flux corrections, this has stash code 186 and field code 672

(ii) Copy the file to the T3E

(iii) Use the routine pptoanc to produce an ancillary file of the flux correction

(iv) In the UMUI, activate flux corrections for the water flux and set it to updating. The time period for updating should be set to as long as possible, so it is never updated.The flux correction file should point to the ancillary file on the T3E

(v) Ensure stash 185 is unavailable, this can be done by creating a user stashmaster record for 185 using an option code or version of a model component that is not being used.

(vi) Since heat flux correction is not being supplied, the following modification should be used to ensure spurious heat fluxes are not applied.

```
*IDENT PJL3F403
*/ Make sure heat flux correction is not used
*DECLARE SFCADD
*D OJT0F304.72
          fluxtoth(I)=0.0
```

Appendix C

In section 4.1 a method for offsetting unrealistic water loss from the pre-HadCM3 to the frozen ice sheets was discussed. The method involved applying to the HadCM3 ocean the water lost to the northern hemisphere ice sheets over a region corresponding to estimates of the limits of iceberg travel. The limits of iceberg travel were deduced from observations and the results of an iceberg population model. Over the region of the observed icebergs, iceberg melt rate was taken as constant except close to the ocean boundaries it was scaled linearly to zero over 5 grid boxes. Here, the spatial pattern of iceberg melting, as predicted by an iceberg population model, is also taken in to account. The iceberg model used to produce the melt rate distribution was that of Bigg *et al* (1996) and includes a treatment of both iceberg dynamics and thermodynamics. In the iceberg model a spectrum of iceberg sizes is calved from each of a number of locations in the northern hemisphere as estimated from observations. The position and size of these icebergs is then tracked by the iceberg model. The spatial pattern of iceberg meltwater provided by the iceberg model was scaled by a multiplication factor to ensure the correct amount of water was added to the HadCM3 ocean to maintain the salinity balance. A test run was performed which included this revised northern hemisphere water flux treatment together with the discarded runoff and southern hemisphere ice sheet contribution as before. The new test simulation (**abcia**) also included the reference salinity fix described in section 4.1.

The simulation **abcia** was run for approximately 40 years. The maximum north Atlantic overturning stream function was compared with that of run **aaxzc** which used the simpler iceberg distribution for the northern hemisphere region described in section 4.1. The time series of north Atlantic overturning stream functions are shown in figure C1 for the two runs. The increased complexity of the iceberg distribution in run abcia appears to have resulted in no significant difference in strength of the thermohaline simulation during the 40 year test period.

Acknowledgement:- Dr Grant Bigg of the University of East Anglia kindly provided the iceberg melt distributions obtained from his iceberg model.

Figure C1: Comparison of maximum north Atlantic overturning stream function values