# OCEAN APPLICATIONS TECHNICAL NOTE 1

## FLUX CORRECTIONS IN COUPLED MODELS

by

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## ABSTRACT

It is shown that the magnitude of the flux corrections required in a coupled model simulation to prevent climate drift do not reduce when a higher resolution, and more realistic, ocean component is used. This is because the flux correction field is dominated by regions of high horizontal SST gradient where even small shifts in positioning lead to large SST errors and therefore large flux corrections. A simple scheme is used to remove the effects of these regions in the calculation of flux correction and this results in corrections considerably smaller than those usually obtained. The remaining peak values are generally identifiable with known systematic errors in the atmospheric model.

## 1. INTRODUCTION

When coupled ocean-atmosphere models of the climate system are integrated over extended periods of time, it is often the case that the simulated climate drifts away from that observed to such an extent that the basic model state is not a suitable control around which to investigate climate perturbations. This is the problem of climate drift.

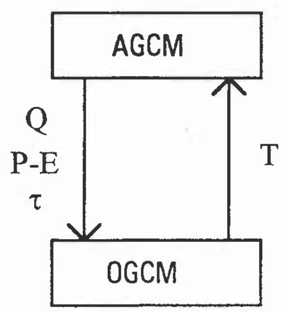
In order to reduce the systematic errors and keep the model climate closer to reality, artificial forcing terms are often added to the fluxes simulated at the ocean-atmosphere interface. The use of this flux correction technique compensates for a wide variety of different errors in both the atmosphere general circulation models (AGCMs) and ocean general circulation models (OGCMS) (Sausen et al, 1988).

These corrections, rather than being small, are typically of the same order as the net surface heating and the reduction of these corrections is of primary importance in the development of coupled models. In this paper we show that many of the dominant features in the flux correction field defined by a coupled model are a consequence of the models inability to correctly simulate the detailed thermal structure associated with the major ocean currents. When this contribution to the flux correction is removed, the remaining corrections are considerably smaller in magnitude and generally correspond to known systematic errors in the atmospheric and ocean models.

The flux correction technique is looked at in detail in the Section 2, the models are described in Section 3, the sensitivity of the calculated flux correction to ocean model resolution is explored in Section 4, and Section 5 discusses some alternative methods of flux correction. Finally, Section 6 contains a summary.

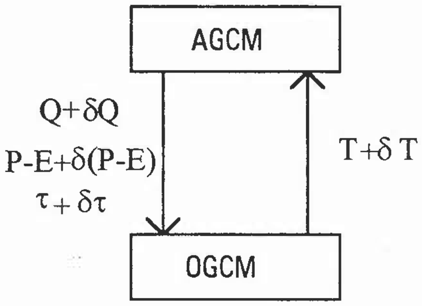
## 2. FLUX CORRECTION

It will be useful for what follows to look at the technique of flux correction in some detail. Schematically the coupled model may be represented as indicated in fig.(a) below more closely .




(a)

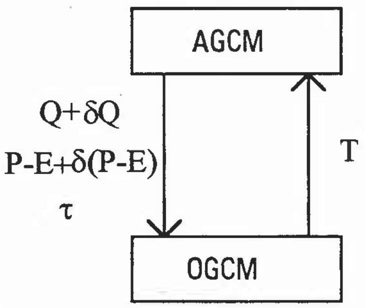
The surface fluxes of heat (Q), water (P-E) and wind stress ($\tau$) are calculated by the atmospheric model using the SST simulated by the ocean model. The ocean model then updates the sea surface temperature T and passes this back to the atmospheric model. When this system is integrated, the simulated T often moves away from the climatological SST leading to climate drift. The key notion of flux correction is illustrated in fig.(b) which shows correction terms added to the forcing fluxes and the predicted SST to 'correct' for the model biases. The correction terms on the fluxes are to correct for the systematic biases in the fluxes calculated by the atmospheric model and the SST correction is to correct for the biases in the calculation of surface temperature by the ocean model.



(b)

If the fluxes were known accurately then the correction terms could be calculated simply by differencing the model and the observed climatological fluxes. Similarly, the ocean model could be integrated with the known climatological fluxes and the SST correction calculated. This type of scheme has been discussed by Sausen et al (1988).

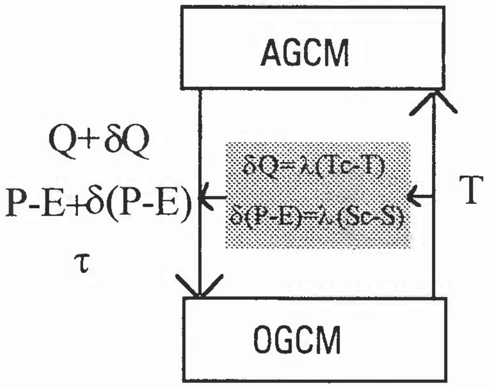
In fact the fluxes are not known accurately, which means it is not clear how to calculate the individual corrections to the different flux components. For example, there is no unambiguous way to distinguish between biases in SST caused by wind stress errors and those caused by heat flux errors. Most coupled models using flux correction have employed a simplification of this full correction scheme which is illustrated in fig.(c). (Manabe et al, 1991; Murphy, 1995).



(c)

In this scheme all corrections are lumped into the surface heat and water flux correction. In this case the correction terms correct explicitly for surface heat and water flux biases but also, implicitly, for wind stress biases and ocean model biases. The problem here is the determination of the correction terms. If the basic assumption is made that T should remain close to $T_C$ at all points on the globe, and the same for salinity (e.g. the modelled SST (or salinity) should stay close to the climatological SST (or salinity) etc.), then the correction terms can be easily obtained by relating the correction to the difference of the model fields from climatology. A scheme similar to that in fig.(d) is then used to define the flux corrections.

The requirement that T remains <u>everywhere</u> close to $T_C$ is only one of many possible conditions to define the correction terms. For example, global mean corrections could be made to ensure the global mean temperature does not drift but allow temperatures to be locally free or, alternatively, temperatures could be corrected more strongly in some regions than in others. Alternative schemes will be investigated later in this paper.



(d)

The integration of the system shown in fig.(d) eventually leads to stable values of the correction terms δQ and δ(P-E). As indicated, this correction is calculated as a simple restoring term (specific details are given later). Once these stable values are determined, they are then applied as geographically and seasonally varying 'constant' corrections to the surface heat and fresh water fluxes in climate experiments. There are differences in the details of how the correction terms are calculated, which is usually done by a combination of coupled and uncoupled integrations of the system illustrated in (d). (Manabe, 1991, Murphy,1995)

One of the major drawbacks of this widely used scheme is that biases arising from the ocean model simulation, which may have nothing to do with the local fluxes, are corrected by a local flux term. One of the purposes of this paper is to investigate a scheme in which some separation is made between the biases arising from local ocean - atmosphere fluxes and those arising from the ocean model simulation. One way to investigate this is to consider coupled simulations with the same atmospheric model but with different ocean models.

In particular, the focus in what follows is the sensitivity of the flux correction to the horizontal resolution in the ocean model. In models using the flux correction technique the corrections are of a similar size to the air-sea flux itself (Gates et al, 1993). Since these flux correction terms have no physical basis, it is clearly necessary to try and reduce their size and eventually the need for them altogether by improving the ocean and atmosphere models. The ocean models that are typically used in coupled climate experiments have a resolution that is very coarse compared to many ocean features of climatic importance (Gates, 1993). An obvious improvement in the representation of these features is to be expected by increasing the horizontal resolution of the ocean model.

## 3. MODEL DESCRIPTION

A number of coupled GCM experiments will be described that have been carried out to investigate the sensitivity of the flux correction to ocean model resolution. They fall into two categories: the first are straight coupled simulations without flux correction (fig.(a) above); the second set are simulations that are fully coupled but include the addition of SST and surface salinity relaxation terms (fig.(d) above). All the experiments use the same atmospheric model but two versions of the ocean model with different resolutions are used. The atmosphere model is the 19-level climate version of the Meteorological Office Unified Model used for both operational weather forecasting and climate research (Cullen, 1991). The model includes detailed physical parametrisations such as an interactive cloud scheme, a detailed representation of the planetary boundary layer and surface processes.

The ocean model is that of Cox (1984) with the inclusion of a number of additional physical parametrizations. Two different global versions of the ocean model were used in the coupled integrations described in this paper. Both versions have 20 levels in the vertical and realistic bottom topography. The vertical levels are distributed to give maximum resolution in the surface mixed layer with the first five levels having a spacing of around 10m. The higher resolution model has a horizontal 1.25°x1.25° grid with enhanced meridional resolution in the tropics down to a 1/2° spacing at the equator. This enhanced tropical resolution is included to better represent the relatively small scale features in the equatorial dynamics. The basic 1.25° grid is half of the meridional resolution and a third of the zonal resolution of the atmospheric model grid.

The second model is the coarse resolution ocean model that has been extensively used at the Hadley Centre in climate change integrations (Murphy, 1995). It is global in extent and has a uniform 2.5°x3.75° latitude-longitude grid.

Both ocean models have the same representation of vertical mixing near the ocean surface. In each case there is an embedded Kraus-Turner mixed layer model (Kraus and Turner, 1967) and, in addition, a Richardson number dependent vertical diffusion (Pacanowski and Philander, 1981). The parameters in these schemes are the same in each of the models. The horizontal diffusion is parametrized as a simple down gradient flux in the momentum and the tracer equations. The tracer diffusion employs the isopycnal mixing scheme of Redi (1982) and has the same parameters in both models. The momentum viscosity, however, is chosen in order to maintain numerical stability and is also a function of the local grid spacing. The value in the 2.5°x3.75° model is $3 \times 10^9$ $cm^2 s^{-1}$ and in the 1.25°x1.25° model it is dependent on the grid spacing (that is, the numerical viscosity is smaller in the higher resolution parts of the grid in the tropics) but has typical values of $5 \times 10^7 cm^2 s^{-1}$. A physically reasonable value is thought to be $O(10^7)$. The high value in the coarse resolution model is necessary to maintain numerical stability. Clearly this large diffusion in the 2.5°x3.75° model will have a significant effect on ocean features with a small spatial scale (Gordon, Wright and Roberts, 1995). Sea ice is not explicitly modelled in the experiments described below and polewards of 70°N and 70°S the modelled SSTs are constrained to be close to climatology. This effectively fixes the sea-ice extents to their climatological position.

The method used to define the flux correction is that of Manabe et al (1991) and Murphy (1995) in which the SST relaxation term is calculated as:

$$\delta Q = \lambda (T_C - T) \qquad (1)$$

where $T_C$ is the seasonally varying climatological SST, T the SST predicted in the ocean model and $\lambda$ the relaxation coefficient. (In what follows attention will be given to the corrections to the heat flux. Similar considerations also apply to the fresh water flux). This is equivalent to the simple linear feedback forcing introduced into ocean modelling by Haney (1971), which he formulated in terms of an effective air-temperature. The consequent relaxation terms $\lambda(T_C-T)$ are averaged over the last few years of the integration of each model in order to form flux correction fields for a subsequent coupled climate experiment. A value of $\lambda$ ($\lambda$=163$Wm^{-2}$), considerably larger than the 'physical' values suggested by the calculations of Haney (1971) and subsequent calculations, was used in the correction procedure. This is because, in the context of flux correction, the value of $\lambda$ is chosen so as to ensure the predicted SSTs remain 'close' to climatology, where the relaxation coefficient $\lambda$ determines how 'close'.

The basic spatial pattern of the flux correction field is established within a few years and, for the most part, does not alter very significantly after this time. Experiments conducted using a 2.5°x3.75° ocean model with accelerated physics show that over most of the globe the heat flux correction field has the same general pattern as that described here after 1200 years. The exception is the region off the southern hemisphere ice-edge where there are significant long time scale changes in the pattern associated with the spin-up of the deep circulation in this region of the model ocean (Wood, personal communication). It will be shown later that it is the regions of high horizontal SST gradient that largely dominate the flux correction pattern, and the positioning of these gradients remains essentially fixed in time, as these are associated with particular oceanographic features (e.g. boundary currents).

The initial conditions for the various runs described below were taken from previous 40 year integrations of the respective coupled models. In these 'spin-up' integrations the initial ocean conditions were climatological Levitus (1982) temperatures and salinities and zero currents.

## 4. THE SENSITIVITY TO MODEL RESOLUTION

In this section the results from coupled integrations with ocean models of different resolution and with and without SST relaxation are compared.

### (i) Models at low ocean resolution

Fig.1 shows the annual mean heat flux correction term for the final (fourth) year for a coupled run using atmosphere and ocean models on a 2.5°x3.75° latitude-longitude grid using standard SST relaxation. This should be compared with fig.2 which shows the annual mean net surface heat flux over the last year from the same model integration. In some areas the flux correction is of a similar magnitude to the total flux and this becomes very apparent in the zonal mean fluxes shown in fig.3. Why are the flux corrections so large? The peaks in the zonal mean correction are associated with particular oceanographic features that are poorly simulated in the coupled model. For example, as fig.1 shows, the large values around 35-45°N are associated with the Gulf Stream and Kuroshio current systems. (The peak at 70°N is dominated by the relatively few grid points in the Norwegian Sea).

For purposes of illustration one of the areas with large flux correction, the North Atlantic, is considered here in detail. Fig.4 compares the North Atlantic SSTs from this 2.5°x3.75° coupled model run with the $\lambda(T_C - T)$ relaxation term (fig.4a), the equivalent temperatures after four years of integration from a fully coupled simulation without any SST relaxation (fig.4b) and the Levitus (1982) climatological SSTs (fig.4c). Comparing the SSTs from the simulation with no SST relaxation (fig.4b) and climatology (fig.4c) shows the coupled model run without SST relaxation has the maximum gradient of SST further to the south and considerably less tight than the climatological gradient. It is not surprising that the ocean model cannot maintain high horizontal SST gradients with the coarse resolution and large lateral viscosity in the ocean model. As expected, when the relaxation term is included the SSTs are forced towards climatology (fig.4a). Although the SST field in this case looks close to climatology it is important to note that this is brought about almost entirely by the SST relaxation term. In other words, the relaxation term is replacing the heat that should be carried by advection in the modelled Gulf Stream system.

It is also of interest to look at the surface current fields in these two experiments. Figs.5a and b show the annual mean North Atlantic sea-surface currents for the 2.5° x3.75° model simulations with and without the relaxation of SST towards climatology. In both integrations the position of the maximum North Atlantic Current speed is too far south and the magnitude is very weak ($15cms^{-1}$ in the model compared to around $100cms^{-1}$ from ship drift current data. See fig.10c). This is a consequence of the low resolution of the model and the high value of the viscosity coefficient needed for numerical stability. In the run without relaxation to climatological SST (fig.5b), the position of the maximum temperature gradient and maximum current are reasonably coincident, whereas in the model with SST relaxation they are distinctly further apart (c.f. figs.4a and b). Using SST relaxation alters the location of the maximum SST gradient but not the location of the maximum in the surface currents. This is because the relaxation forcing only restrains the near surface temperatures, whereas it is the sub-surface thermal structure, integrated over depth, that determines the geostrophic part of the surface current. In the forced run the maximum SST gradient is considerably to the north of the maximum current, which is unrealistic. It is shown later that this inconsistency between the surface temperature and current fields in simulations where the SST is relaxed towards climatology becomes even more apparent as the ocean model resolution is increased and the strong currents are better resolved.

In this case the large flux corrections are necessary to compensate for the poor simulation of the Gulf Stream/North Atlantic Current in the coarse resolution model. Large corrections are therefore also expected in association with other narrow ocean features such as the Kuroshio, the Antarctic Circumpolar Current, coastal upwelling zone etc. These are all evident in the global flux correction field shown in fig.1.

(ii) <u>Runs at higher ocean resolution</u>

At first sight it might be expected that increasing the horizontal resolution of the ocean model would improve the ocean simulation and thereby reduce the correction needed to the heat flux. To investigate whether this is the case a coupled simulation, parallel to that described above, was performed with a higher resolution ocean component. As already discussed, this version of the ocean model has the same vertical resolution as the coarser 2.5°x3.75° model and the same physical parametrizations of mixing and convection.

Fig.6 shows the annual mean of the heat flux correction for the final fourth year of a coupled simulation with the same SST relaxation as before and with a horizontal ocean grid resolution of 1.25°x1.25°. This should be compared with fig.1 which shows the flux corrections obtained with the 2.5°x3.75° ocean model. The magnitude of the heat flux correction is not reduced at the higher resolution. Both models show a broadly similar pattern in the positioning of the major maxima and minima, though there are differences in their exact location and magnitude.

The zonal mean flux correction for this model, compared to the net surface heating, is shown in fig.7. It is apparent by comparing with fig.3 that the higher resolution in the ocean model has not reduced the flux correction values. In fact, the peak values have increased in size.

Many of the peaks in the flux correction occur close to areas with large horizontal SST gradients such as the western boundary currents and the Antarctic Circumpolar Current. The co-relation of the high corrections with the regions of sharp horizontal SST gradient is illustrated in fig.8 in which the magnitude of the flux correction from the higher resolution model (the 'grey' scale) is overlaid with contours of the magnitude of the horizontal SST gradient. The grey scale has been chosen to highlight the peaks in the flux correction field. It is clear from this figure that most of the maximum values of the flux correction are in regions of high SST gradient. Many of the high values of flux correction can be explained in terms of a small shift in the positioning of the SST gradient in these regions.

In places where the model simulates the position of high gradients of SST in the model to be shifted from their climatological locations, even by a small amount, the difference between T (the SST in the model) and $T_C$ (the climatological SST) at a particular point can be very large, hence leading to large values of $\lambda(Tc-T)$ and a peak in the flux correction field. The pattern of the major maxima and minima in the flux correction field is thus associated with the inability of the coupled model to simulate the exact climatological position of the regions of high SST gradient. In fact, the peaks in the flux correction field from the higher resolution model are generally greater than those in the coarser resolution model due to the sharper SST gradients that can be maintained in the former, which amplifies the effects of discrepancies in their simulated position. Note also the peaks that occur close to the equator in the higher resolution model in the Pacific. There is a better representation of equatorial upwelling than in the coarser resolution model. The SST climatology used in these experiments was specified on a 2.5°x3.75° grid (the atmospheric model grid) and therefore will not properly resolve the cold SSTs associated with equatorial upwelling. For this reason, the 1.25°x1.25° model simulation on the equator may be realistic in that it produces a narrow band of cold SSTs and therefore positive flux corrections.

Most ocean models have difficulty in simulating the location of narrow ocean currents such as the Gulf Stream and the natural position of the current in a model is often differently placed to climatology (e.g. see Boning et al, 1995). Increasing the horizontal grid resolution of the model does improve the representation of narrow currents in the model but it remains difficult for them to be correctly located in the simulation. For example, as discussed later, the simulated speed of the Gulf Stream is much improved by using the 1.25°x1.25° ocean model.

Fig.9 shows the annual mean North Atlantic SSTs as simulated in the coupled integration with the 1.25°x1.25° ocean component with standard SST relaxation (fig.9a) and without SST relaxation (fig.9b). The higher resolution model is clearly capable of maintaining higher horizontal gradients of SST associated with the Gulf Stream and the North Atlantic Current. However, comparison of the SSTs from the uncorrected run with the climatological SSTs (fig.4c) shows that the model does not reproduce the northward turning of the high gradient in the region of the Grand Banks. The modelled high SST gradient leaves the coast and moves zonally eastward across the Atlantic leading to significant SST errors in the central North Atlantic. This erroneous positioning of the North Atlantic Current is not only a feature of the coupled model but also occurs in ocean stand-alone experiments using climatological surface forcing functions. The reason for the absence of northward turning in the North Atlantic Current appears to be associated with the lack of model deep water formation in the Greenland-Iceland-Norway seas and the Labrador sea (Wright and Gordon, 1995).

In the forced experiment (fig 9a) the SST field is now a combination of the climatology (which is imposed by the correction restoring term) and the unforced model field in fig.9b. As a consequence, the high gradient SST region is very broad.

Figs.10a and b show the annual mean North Atlantic surface currents for the two integrations of the 1.25°x1.25° ocean version of the coupled model with and without SST relaxation. The magnitude and width of the maximum current is more realistic in these simulations using a high resolution ocean model than in those using the 2.5° x3.75° ocean model (cf. fig.5). In both cases fig.10 shows that east of 60°W the maximum current is further to the south than observations would indicate, and is close to the location of the maximum SST gradient in the run with no SST relaxation. As already discussed in the low resolution case, the surface current and temperature field are not consistent in the corrected case.

Fig.10c shows the annual mean ship drift currents in the north Atlantic region, and this provides a useful comparison for the model surface current simulation. The UKMO database of ship drift currents, which contains 5.4 million observations from 1850 to the present day, was used to produce this long term mean field of currents on a 1°x1° grid. The figure shows that the maximum current speeds (up to 100cms$^{-1}$) in the Gulf Stream are much larger than those simulated by the model. The position of the maximum currents is also shown to be further north, less wide and less zonal than in the model, with a slight northerly diversion east of 45°W.

In modelling terms it may well be better to allow the coupled model to place these high gradient regions, associated with the strong narrow ocean currents, slightly offset from their climatological positions, rather than falsely forcing the SSTs using the relaxation term. The higher resolution ocean model can maintain the high SST gradients to some degree, and the use of the standard flux correction overwrites the positive benefits obtained by using the higher resolution in the ocean model.

In summary, the results discussed so far indicate that in the coupled GCM experiments:

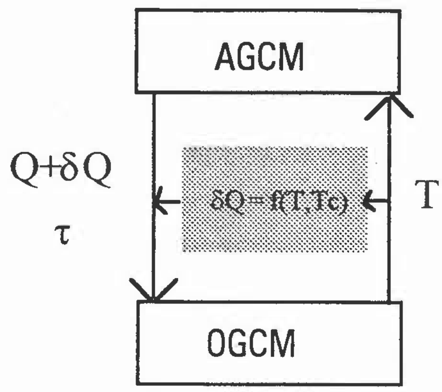
i) Increasing the resolution of the ocean model will generally not reduce the size of the peaks in the flux correction.

ii) Many of the peaks in the flux correction are associated with regions of high horizontal gradient in SST.

iii) When the models are flux corrected at every point the SST and surface current fields are not dynamically consistent.

## 5. ALTERNATIVE FORMULATIONS OF FLUX CORRECTION

In Section 2 fig.d illustrated the simple scheme of flux correction used in the coupled experiments described in the previous section. Fig.e is a generalised version of this scheme (the water flux and salinity terms are not shown explicitly). The scheme in fig.e is generalised in that the method used to determine the flux correction terms is not necessarily a local relaxation to the climatological SST. There are numerous possibilities as to how the function $f(T,T_C)$ can be defined. For example:




(e)

i) <u>A global mean constraint:</u>

$$\delta Q = \lambda(\overline{T_c} - \overline{T}) = \lambda\overline{(T_c - T)} \qquad (2)$$

where the overbar denotes the global mean. This scheme would ensure there is no drift in the global mean SST although large local drifts may be allowed.

ii) <u>A zonal mean constraint</u>. In this case the overbar in (2) would refer to the zonal mean.

iii) <u>An area mean constraint.</u> In which the overbar in (2) represents a running area mean over a predetermined sized area. The obvious way to introduce a horizontal scale into the flux correction is to apply the relaxation to area averages rather than to point values. The idea of a scale dependent flux correction is to constrain the SST on the 'large scale' but to allow the model to more freely determine the SST on the 'smaller scale'. The aim is to remove the effect of large flux correction values in regions with a high SST gradient, and thereby allow the coupled model more freedom to determine its own positioning of these gradients, but keep the general overall pattern of flux correction in regions where it is compensating for poor heat fluxes and large scale problems in the atmosphere and ocean models.

Such a scheme was initially implemented using a simple spatial smoothing. However, the effect of a few large values associated with the shift in SST gradient is simply spread out, leaving a smoother field but with the flux correction still peaking in the high gradient regions. If the field is smoothed still further, so that these peak values are considerably reduced, the flux correction is reduced too much in regions where large values are necessary to counteract surface flux problems in the coupled model (e.g. regions of too little or too much cloud).

iv) Constraining the SST only in certain regions. Rather than to attempt to define the flux correction in terms of the local or mean SST field, an alternative approach is to only constrain the SST to be close to climatology in certain areas. The scheme discussed below constrains T to be close to Tc except in regions where the horizontal gradient of SST is high. It has already been argued that to locally constrain the SST in high gradient regions is not desirable.

Various methods have been tested, and in the one discussed here the high resolution ocean coupled simulation described in the previous section was repeated but with a revised SST relaxation in these high SST gradient regions. The basic approach was to replace the relaxation term in these regions with typical values from the surrounding open ocean. What this means is that the high gradient regions essentially play no part in determining the flux correction field.

This was achieved by first identifying the regions of high SST gradient using a simple criterion determined by experimentation. The criterion applied was $|\nabla \mathbf{T}| > 1.5\text{x}10^{-5}$ $\text{Km}^{-1}$ (c.f. fig.8) and the following procedure was used.

The SST relaxation contribution to the surface heat flux was calculated in the usual way (i.e. equation (1)) and then large values were removed from regions where the gradient criterion was satisfied. The position of the major ocean currents, and their corresponding temperature fields, shift only slightly from their climatological positions in a model integration that uses SST relaxation. Consequently the maxima of $|\nabla \mathbf{T}|$ and $|\nabla \mathbf{T}_c|$ are in slightly different locations and large values of the SST relaxation term (that is $\boldsymbol{\delta}\mathbf{Q} = \boldsymbol{\lambda}(\mathbf{T}_C - \mathbf{T})$) will both occur in regions of high $|\nabla \mathbf{T}|$ and high $|\nabla \mathbf{T}_c|$. A simple approach was taken (to ensure that regions of high $|\nabla \mathbf{T}_c|$ were also eliminated) in which a small area surrounding the high $|\nabla \mathbf{T}|$ regions was also removed in the calculation of the relaxation term. Finally the removed values were replaced by values interpolated from the surrounding ocean.

When using a scheme such as this in long integrations, an additional constraint may also be necessary to ensure there is no drift in global mean SST (in the runs described here the global mean flux correction is less than $1\text{Wm}^{-2}$ and no additional global constraint was applied ).

The above procedure is illustrated in fig.11 using an annual mean field of the calculated SST relaxation term. For clarity, only the North Atlantic region is shown. Fig.11a shows the original field as calculated in the high resolution ocean coupled simulation described in the previous section using the relaxation in equation (1). The regions removed on the basis of the $|\nabla T|$ criteria are illustrated in fig.11b and the additional surrounding points are as also included in fig.11c. Finally, fig.11d shows the SST relaxation field applied to the ocean model, in which the removed values have been replaced by interpolated open ocean values.

The North Atlantic SSTs simulated in an integration of the high resolution coupled model using this technique are shown in fig.12. The scheme does not just smooth the correction field because the resulting SSTs from the ocean model feed back on the atmosphere model. The SSTs in fig.12 should be compared with the equivalent field from the unforced and standard SST relaxation cases presented earlier in figs.9a and b. The SST pattern has essentially the same character as in the un-forced integration, and so this scheme based on the elimination of high gradients essentially allows the SSTs associated with the North Atlantic Current to find their natural position in the model.

The resulting global annual mean flux correction field is shown in fig.13 for the final year of the simulation. Comparison with the standard flux correction field in fig.6 shows that, with this new scheme, the peak flux correction values are now considerably smaller in many areas but the overall pattern of the flux correction on the large scale has been generally maintained. The field in fig.13 is not simply a smoothed version of the usual flux correction because the high SST gradients are located in a different position in the simulation and the basic model state is therefore different. Other regions which have fairly large flux corrections are now given greater prominence. Many of these reflect known systematic errors in the atmosphere and ocean models. For example, large values in the flux correction off the west coasts of South America and Africa can be attributed to a lack of marine stratocumulus, which is a well known problem in the atmospheric model, and also to the underestimation of coastal upwelling in the ocean model.

## 6. Summary

The large values of the flux correction associated with high SST gradient regions dominate the standard flux correction fields. It may be desirable in a climate simulation to allow the model more freedom to re-position these regions slightly shifted from climatology. Correct simulation of the positioning of strong boundary currents is a well known problem in ocean modelling, even in models at high resolution.

There is a difference in the impact of flux correction between the coupled simulations with coarse and fine resolution ocean models. In the coarse model, the flux correction effectively 'inserts' ocean features which, because of the poor ocean resolution, the ocean model cannot simulate (such as the tight gradients associated with the major currents). The finer resolution models can simulate these features significantly better and, in this case, the flux correction attempts to repositioning these features in the SST field. With the higher resolution models it clearly makes sense to develop them so that these features are simulated with the correct positioning.

The magnitude of the total heat fluxes between the ocean and atmosphere are not known to more than an accuracy of about $20Wm^{-2}$. Hence, once the errors in the ocean and atmosphere models can be improved so that the magnitude of the flux correction is of this order, it makes sense to 'tune' the models by altering values such as exchange coefficients, vertical mixing parameters and so on, in an attempt to remove the flux correction completely. If point-by-point flux correction is used, high values can always be expected in the strong current regions until such a time as ocean models, within a coupled system, can realistically simulate the positioning of these currents.

The use of the alternative correction scheme discussed in this paper is aimed at removing the effects of the high gradient regions, and to reduce substantially the magnitude of the maxima and minima in the flux correction field, and the constraints in the model simulation. The intention is to allow the model to respond with more freedom in climate simulations. The peaks that remain in the flux correction can generally be identified with known systematic errors in the atmosphere and ocean models.

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## Figure captions

**Fig. 1** Annual mean of the final year of the heat flux correction term for the coupled integration with 3.75°x2.5° atmosphere and ocean components using standard SST relaxation with a coefficient (which has the same value in all experiments) of $163 Wm^{-2}K^{-1}$. The contour interval is $50Wm^{-2}$. Dotted shading is used to show values > $50Wm^{-2}$, dash dot dash shading values < $-50Wm^{-2}$.

**Fig. 2** Annual mean of the final year of the net surface heat flux for the coupled integration with 3.75°x2.5° atmosphere and ocean components using standard SST relaxation. The contour interval is $50Wm^{-2}$. Dotted shading is used to show values > $50Wm^{-2}$, dash dot dash shading values < $-50Wm^{-2}$.

**Fig. 3** Zonal mean of the net surface heat flux and the heat flux correction term in $Wm^{-2}$ for the coupled integration with 3.75°x2.5° atmosphere and ocean components using standard SST relaxation.

**Fig. 4** Annual mean North Atlantic sea surface temperatures (°C) for the final year of the coupled integration with 3.75°x2.5° atmosphere and ocean components (a) using SST relaxation forcing (b) with no SST relaxtaion applied. Fig 2c shows the annual mean SST from the Levitus (1982) climatology.

**Fig. 5** Annual mean North Atlantic sea surface currents in $cms^{-1}$ for the final year of the coupled integration with 3.75°x2.5° atmosphere and ocean components (a) using SST relaxation forcing (b) with no SST relaxation applied.

**Fig. 6** Annual mean of the final year of the heat flux correction term for the coupled integration with 1.25°x1.25° ocean component and 3.75°x2.5° atmosphere component using standard SST relaxation. The contour interval is $50Wm^{-2}$. Dotted shading is used to show values > $50Wm^{-2}$, dash dot dash shading values < $-50Wm^{-2}$.

**Fig. 7** Zonal mean of the net surface heat flux ($Wm^{-2}$) and the heat flux correction term for the coupled integration with 1.25°x1.25° ocean component and 3.75°x2.5° atmosphere component using standard SST relaxation.

**Fig. 8** Magnitude of the heat flux correction term shown in shades of grey with contours of the magnitude of the sea surface temperature gradient ($°Cm^{-1}$) overlaid for the annual mean of the final year of the coupled integration with 1.25°x1.25° ocean component and 3.75°x2.5° atmosphere component using SST relaxation forcing. The contour interval is $0.4 \times 10^{-5}$ °$Cm^{-1}$ and the grey scale shading (in $Wm^{-2}$) is shown on the bar below the chart.

**Fig. 9** Annual mean North Atlantic sea surface temperatures (°C) for the final year of the coupled integration with 1.25°x1.25° ocean component and 3.75°x2.5° atmosphere component (a) using standard SST relaxation, (b) with no SST relaxation applied.

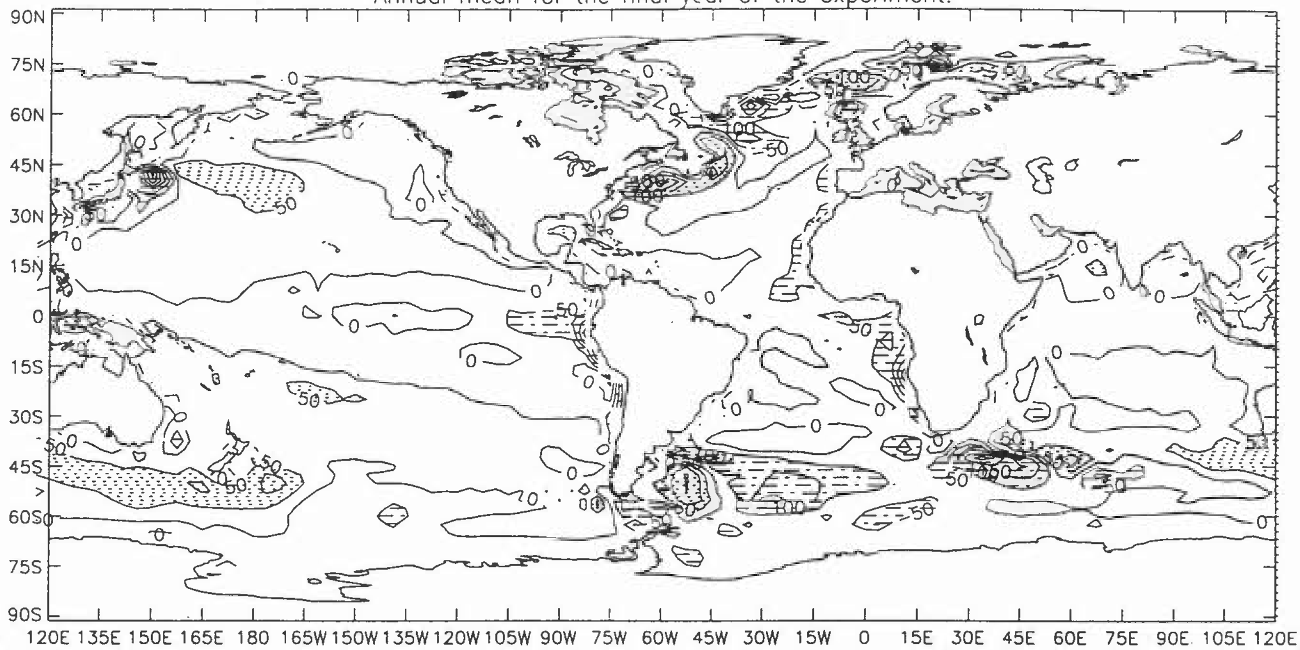
**Fig. 10** Annual mean North Atlantic sea surface currents in $cms^{-1}$ for the final year of the coupled integration with 1.25°x1.25° ocean component and 3.75°x2.5° atmosphere component (a) using standard SST relaxation, (b) with no SST relaxation applied.

**Fig. 11** Illustration of the procedure used to remove regions of high gradient in the SST relaxation contribution to the surface heat flux. The North Atlantic region only is shown for clarity. Fig 11a shows the initial field from the coupled integration with 1.25°x1.25° ocean component and 3.75°x2.5° atmosphere component. Fig 11b shows the field after the removal of regions where both the flux correction and temperature gradient are high (in this case $|\nabla T| > 1.5 \times 10^{-5}$°$Cm^{-1}$ and flux correction > $50Wm^{-1}$). Fig 11c shows the field after the removal of additional surrounding points, where the flux correction is large close to high temperature gradient regions. Fig 11d shows the final field where the removed values are replaced by values typical of the surrounding open ocean. This final field is then applied in the ocean model.

**Fig 12** Annual mean North Atlantic sea surface temperature (°C) for the final year of the coupled integration with 1.25°x1.25° ocean component and 3.75°x2.5°atmosphere component using the revised SST relaxation scheme.

**Fig 13** Annual mean of the final year of the heat flux correction term for the coupled integration with 1.25°x1.25° ocean component and 3.75°x2.5° atmosphere component using the revised SST relaxation scheme. The contour interval is $50Wm^{-2}$. Dotted shading is used to show values $>50Wm^{-2}$, dash dot dash shading values $< -50Wm^{-2}$.

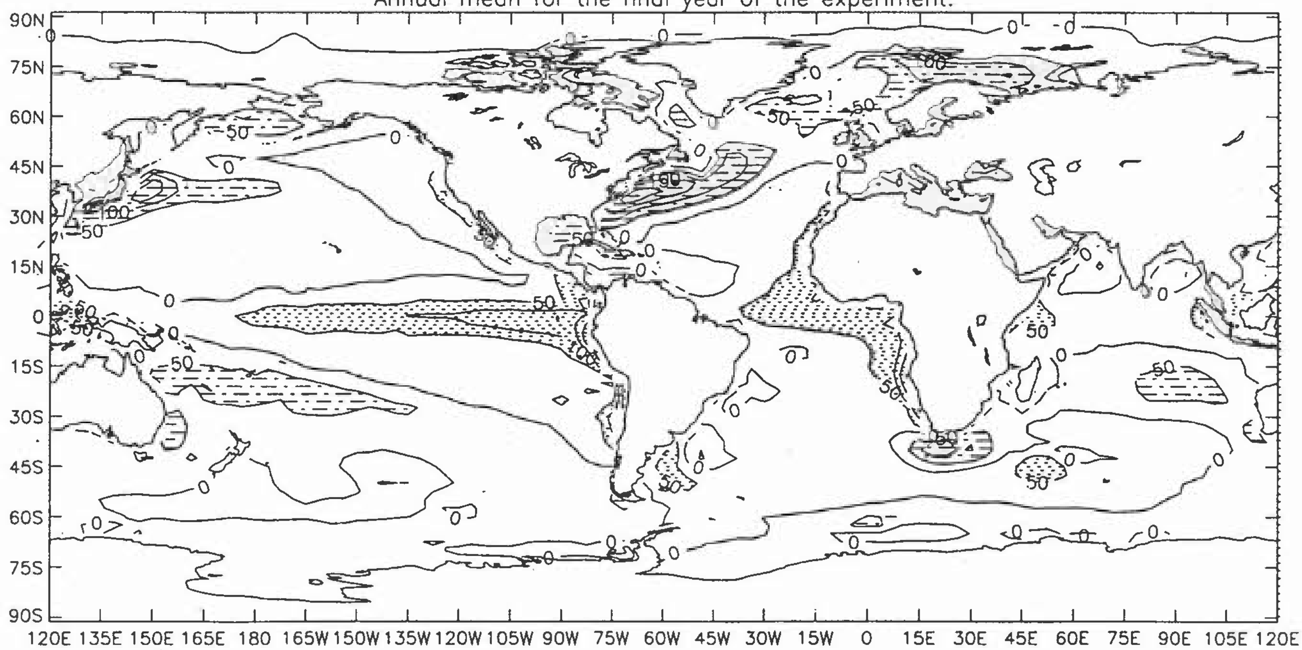
Fig. 1. Heat flux correction term for coupled integration with standard SST relaxation forcing using 3.75 by 2.5 deg ocean and atmosphere models. Annual mean for the final year of the experiment.




Contour intervals of lines = 50 W/m**2

Values greater than 50 W/m**2 : dotted shading
Values less than −50 W/m**2 : dash dot dash shading

Fig. 2. Net surface heat flux for coupled integration with standard SST relaxation forcing using 3.75 by 2.5 deg ocean and atmosphere models. Annual mean for the final year of the experiment.




Contour intervals of lines = 50 W/m**2

Values greater than 50 W/m**2 : dotted shading
Values less than −50 W/m**2 : dash dot dash shading

Fig. 3. Zonal mean of the net surface heat flux ($Wm^{-2}$) and the heat flux correction term for coupled integration with 3.75° by 2.5° atmosphere and ocean components using standard SST relaxation.

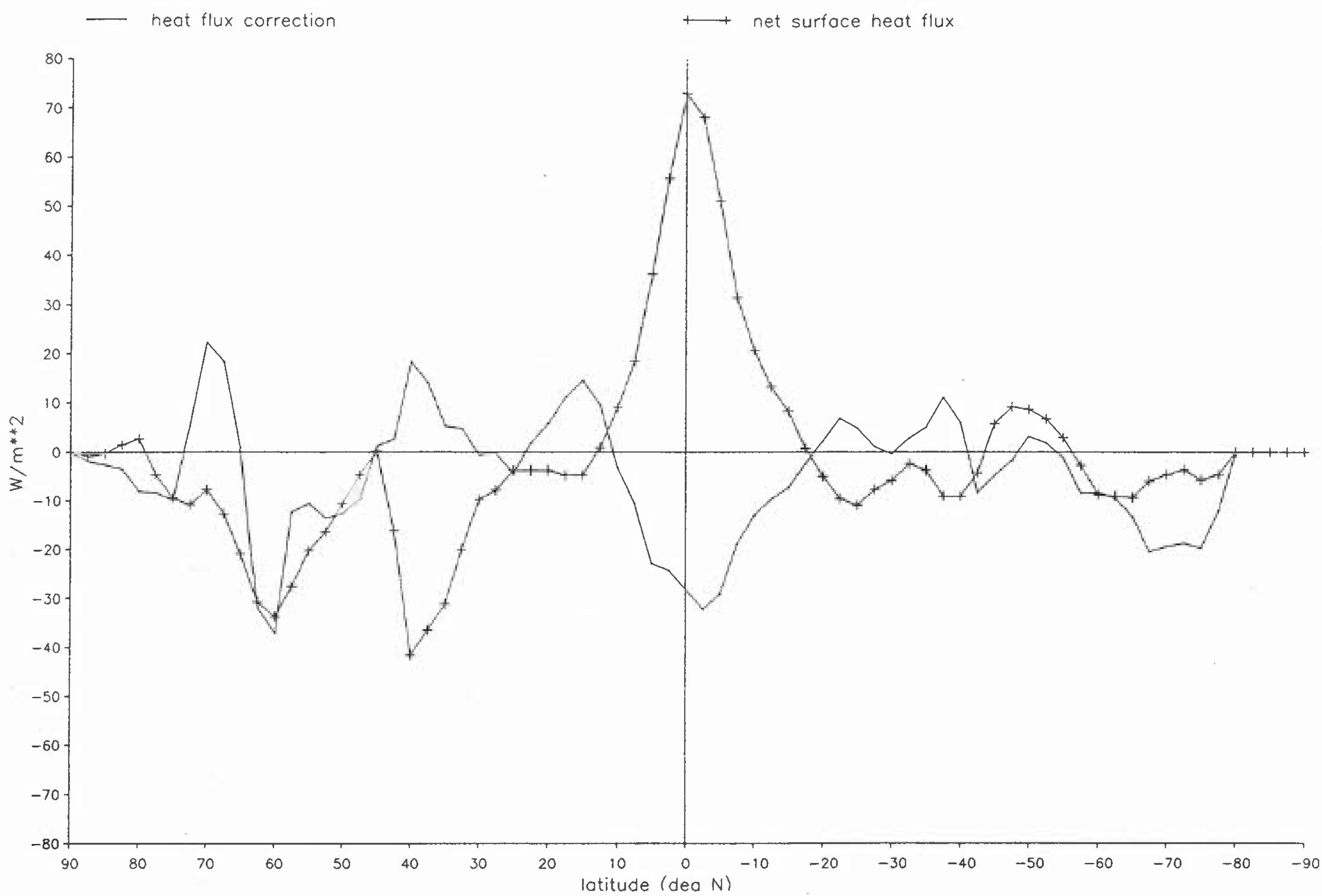


Fig. 4a. North Atlantic sea surface temperature for coupled integration with standard SST relaxation forcing using 3.75 by 2.5 deg ocean and atmosphere models. Annual mean for the final year of the experiment.

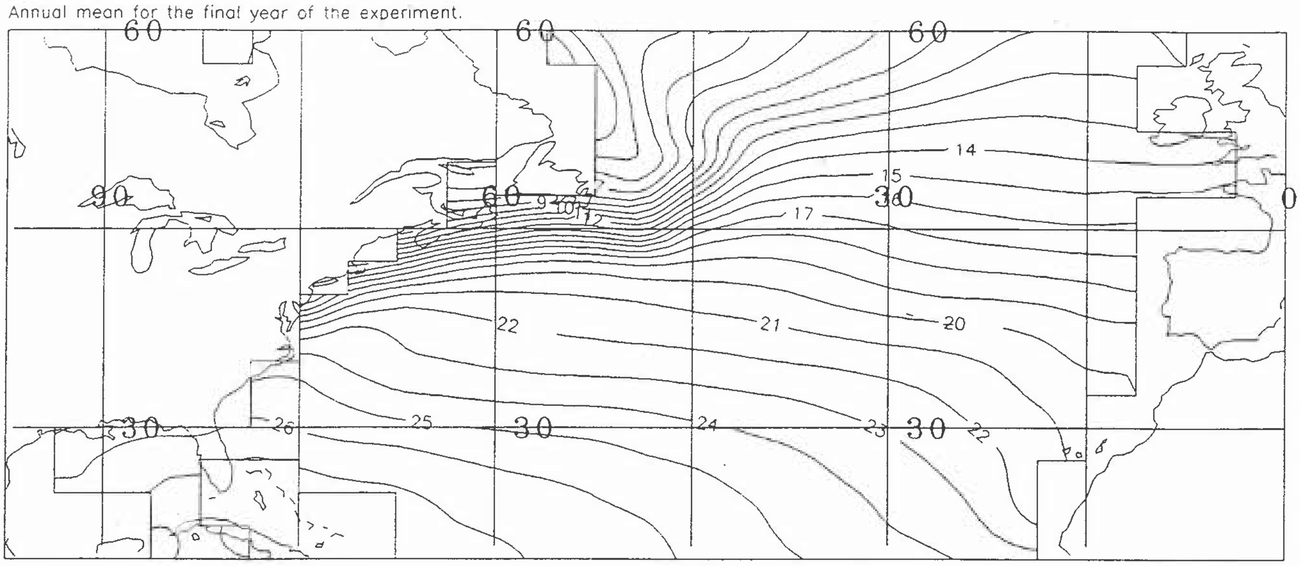



Fig. 4b. North Atlantic sea surface temperature for coupled integration with no SST relaxation forcing using 3.75 by 2.5 deg ocean and atmosphere models. Annual mean for the final year of the experiment.

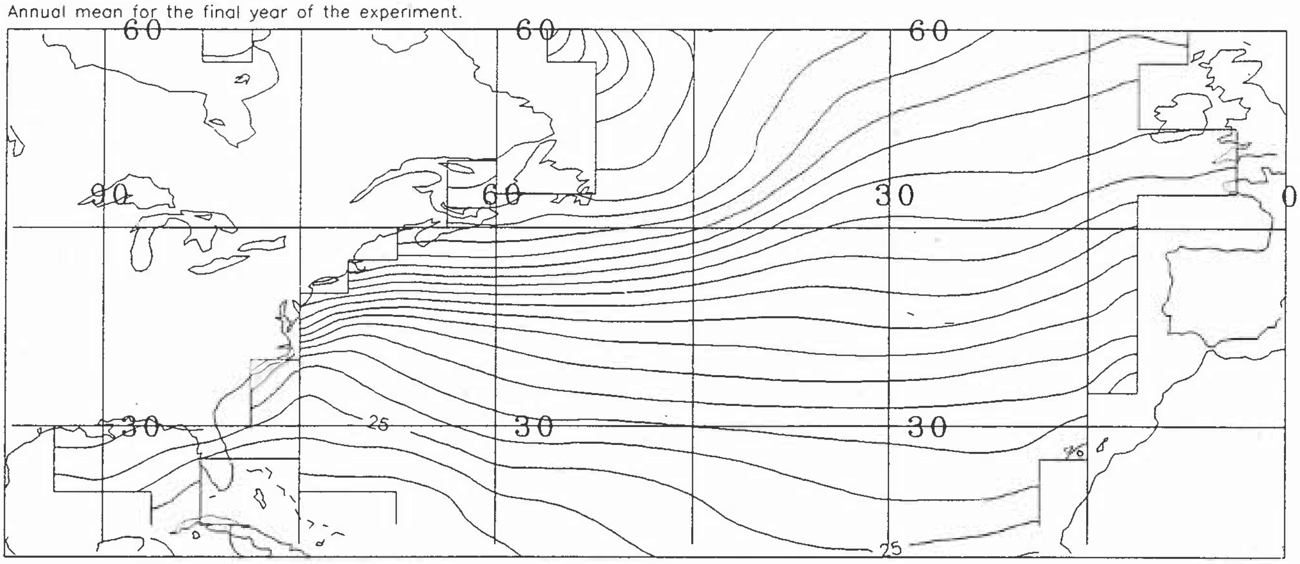



Fig. 4c. North Atlantic Levitus climatological SST. Annual mean.

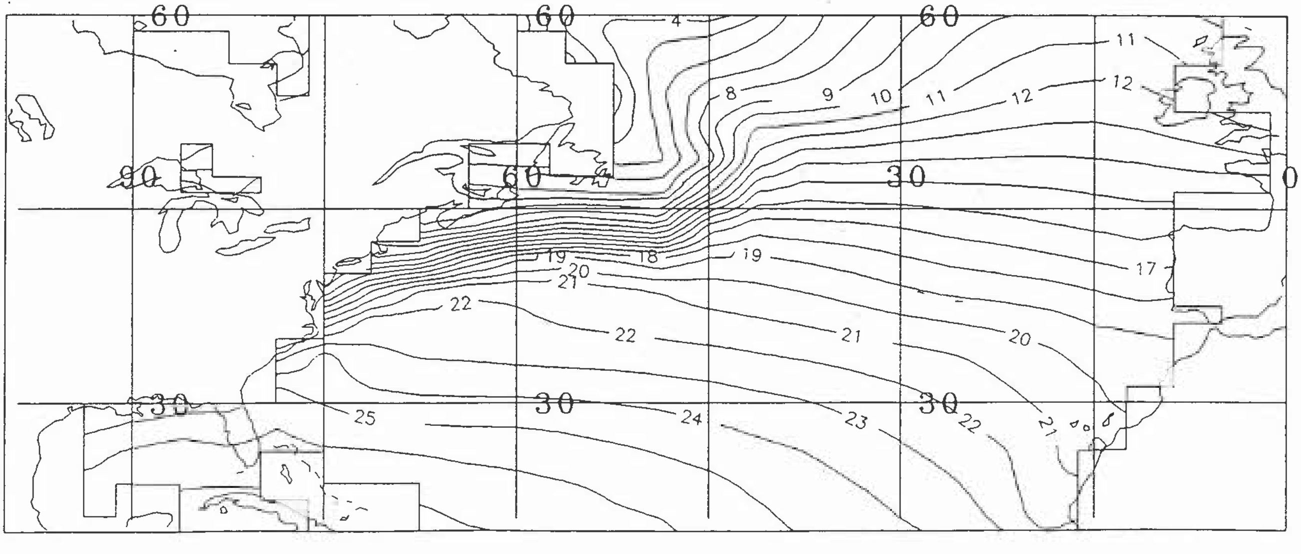
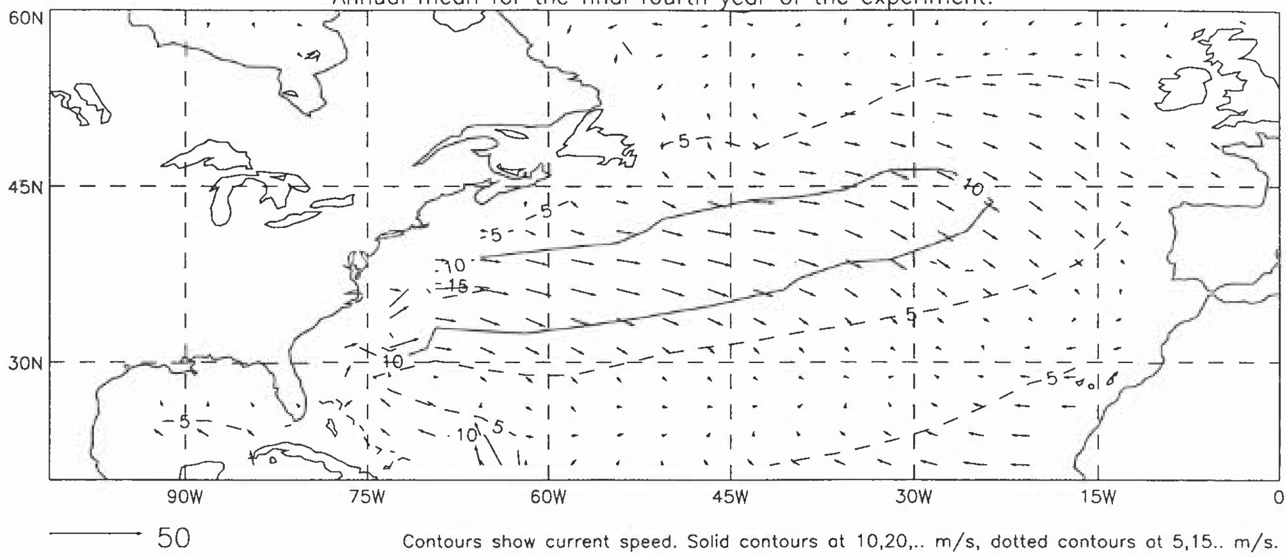
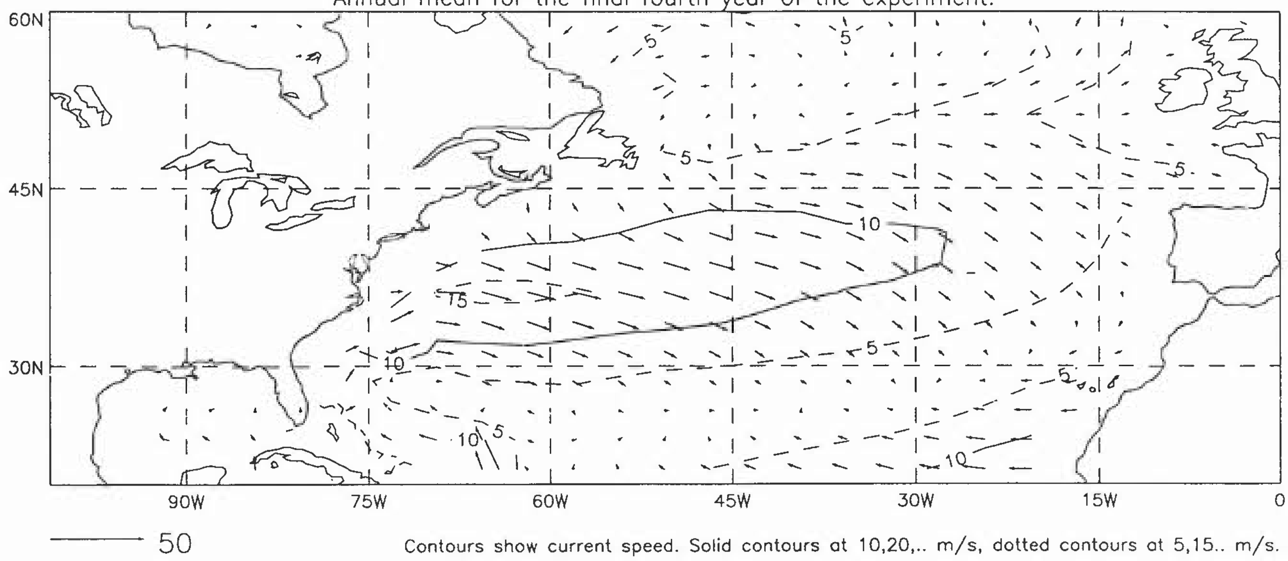


Fig. 5a. North Atlantic sea surface currents for coupled integration with standard SST relaxation forcing using 3.75 by 2.5 deg ocean and atmosphere models.
Annual mean for the final fourth year of the experiment.



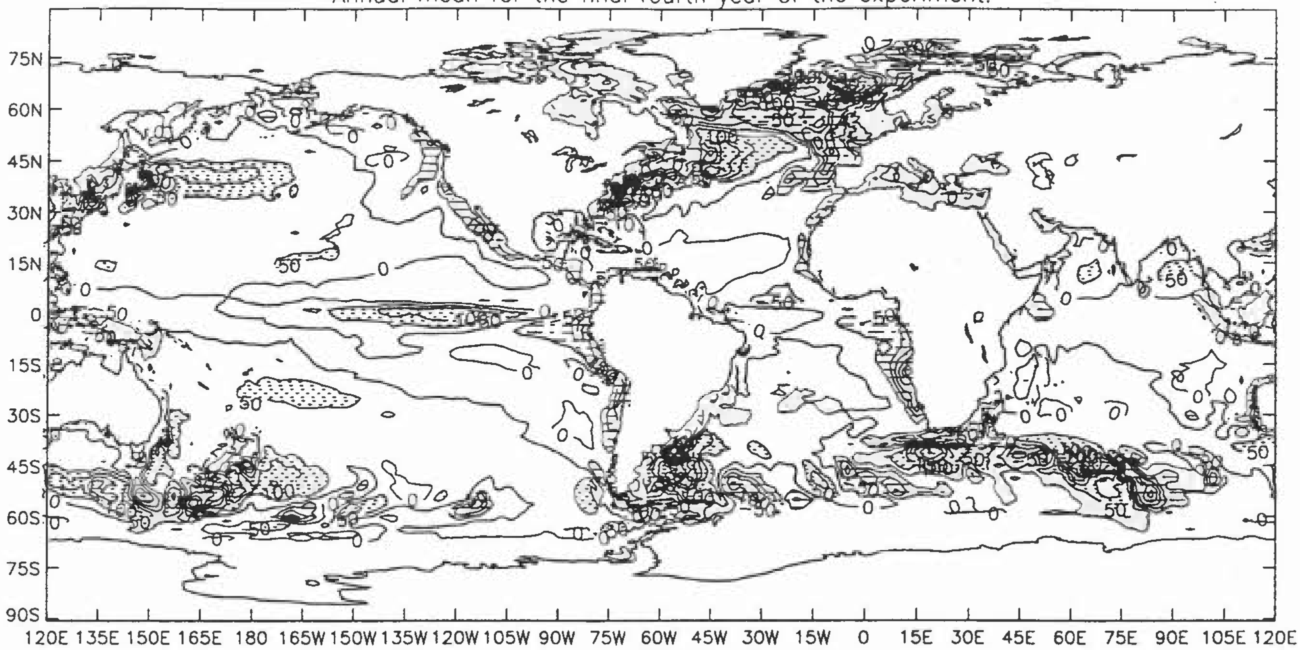

Contours show current speed. Solid contours at 10,20,.. m/s, dotted contours at 5,15.. m/s.

Fig. 5b. North Atlantic sea surface currents for coupled integration with no SST relaxation forcing using 3.75 by 2.5 deg ocean and atmosphere models.
Annual mean for the final fourth year of the experiment.




Contours show current speed. Solid contours at 10,20,.. m/s, dotted contours at 5,15.. m/s.

Fig. 6. Heat flux correction term for coupled integration with standard SST relaxation forcing using 1.25 by 1.25 deg ocean and 3.75 by 2.5 deg atmosphere models. Annual mean for the final fourth year of the experiment.




Contour intervals of lines = 50 W/m**2

Values greater than 50 W/m**2 : dotted shading
Values less than −50 W/m**2 : dash dot dash shading

Fig. 7. Zonal mean of the net surface heat flux ($Wm^{-2}$) and the heat flux correction term for coupled integration with 1.25° by 1.25° ocean and 3.75° by 2.5° atmosphere components using standard SST relaxation.

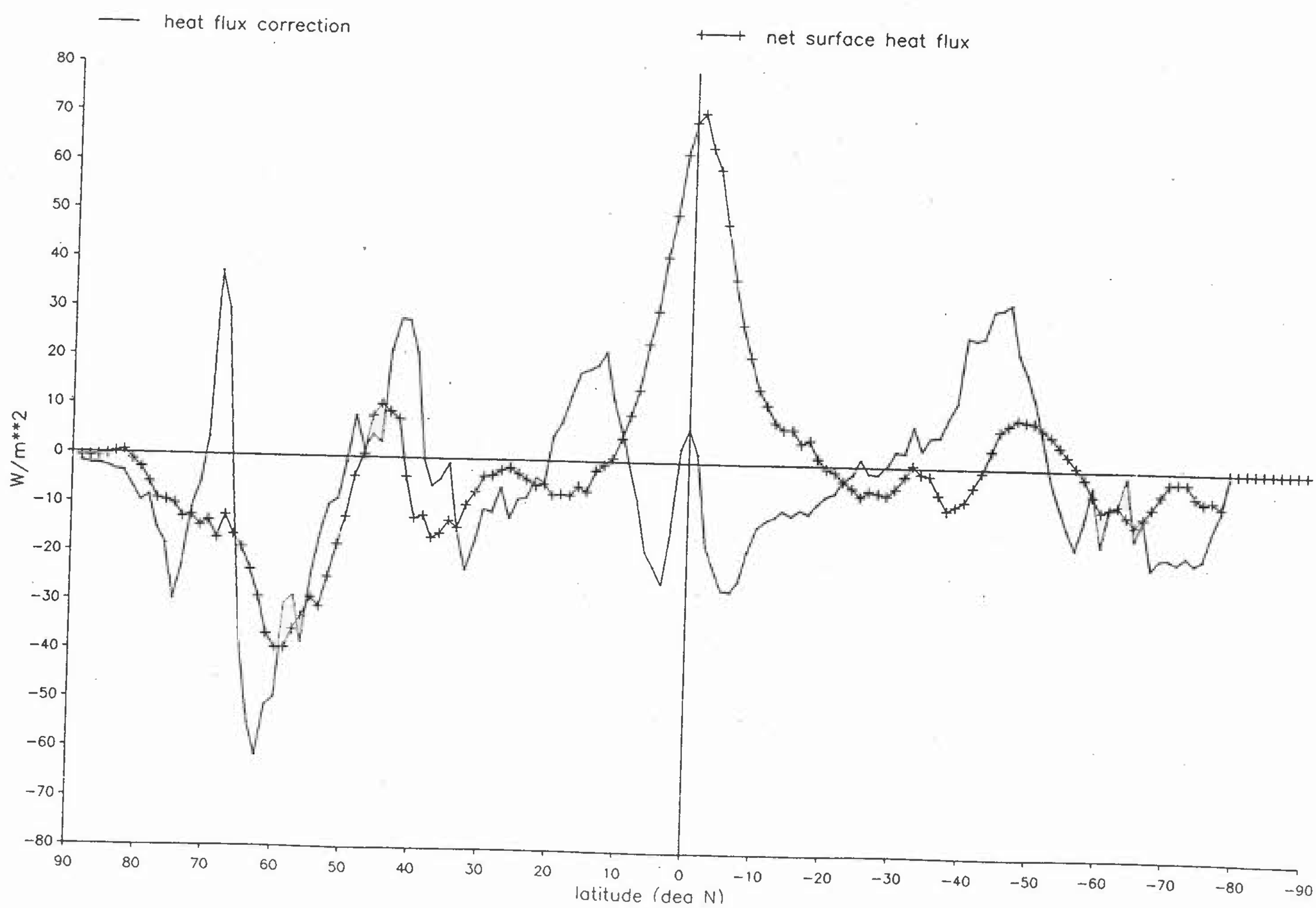


Fig. 8. Magnitude of heat flux correction term (shades of grey) and magnitude of SST gradient (contours) for coupled integration with standard SST relaxation forcing using 1.25 by 1.25 deg ocean and 3.75 by 2.5 deg atmosphere models. Annual mean for the final year of the experiment.

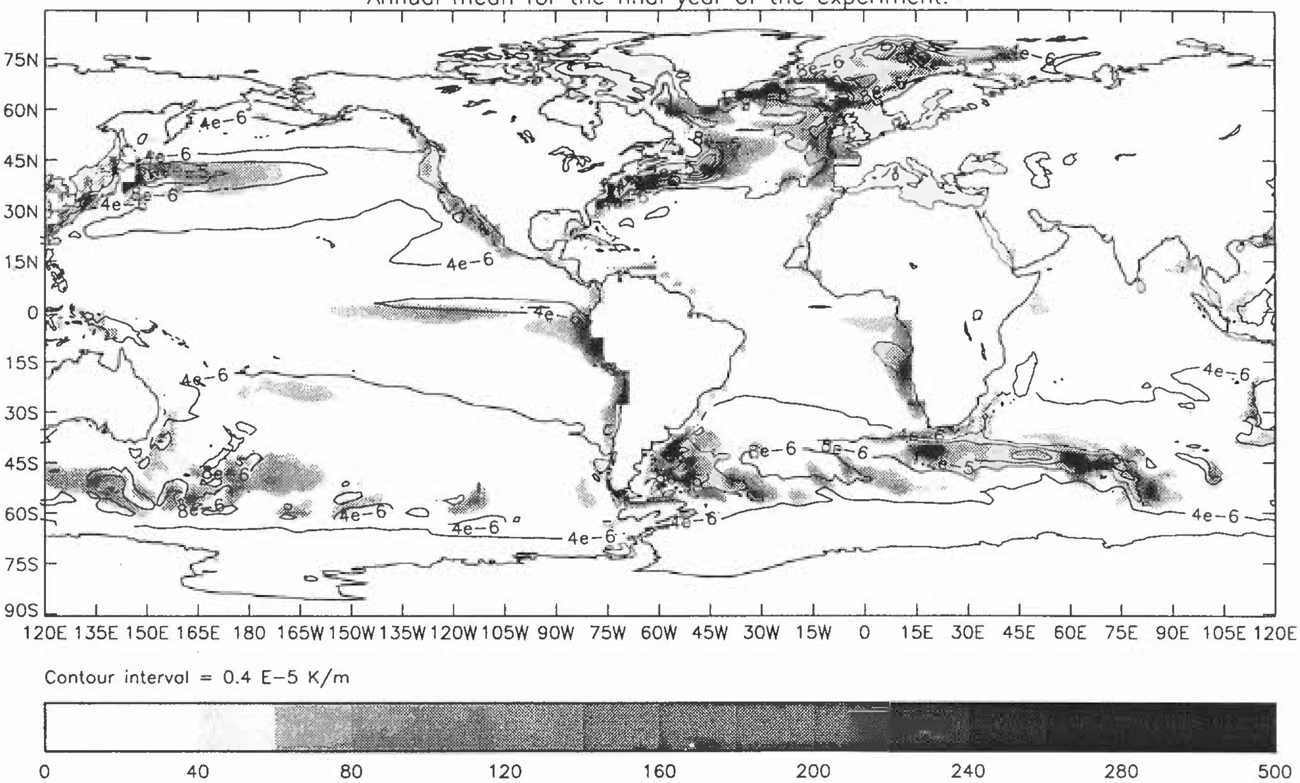


Fig. 9a. North Atlantic sea surface temperature for
coupled integration with standard SST relaxation forcing using
1.25 by 1.25 deg ocean and 3.75 by 2.5 deg atmosphere models.
Annual mean for the final year of the experiment.

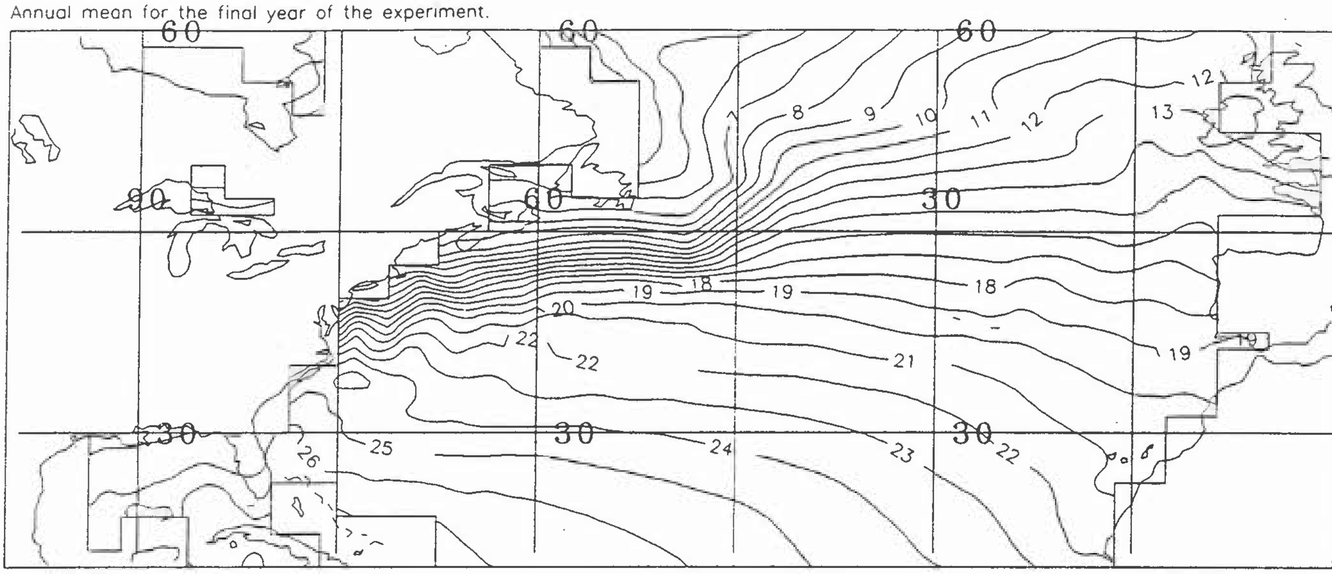



Fig. 9b. North Atlantic sea surface temperature for
coupled integration with no SST relaxation forcing using
1.25 by 1.25 deg ocean and 3.75 by 2.5 deg atmosphere models.
Annual mean for the final year of the experiment.

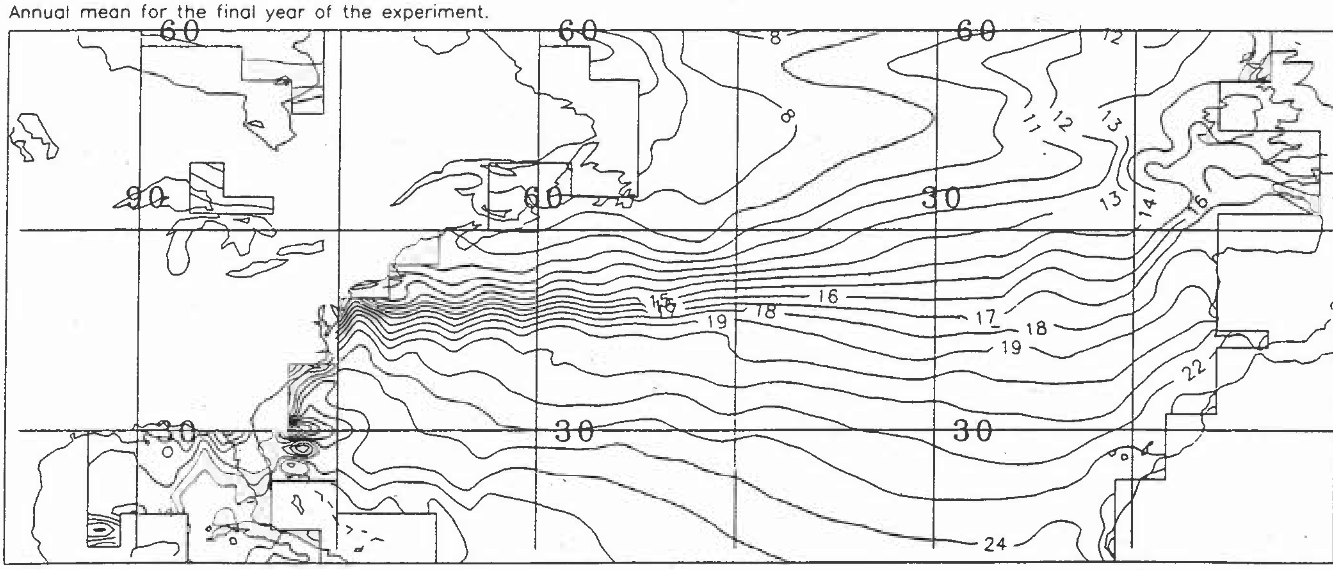
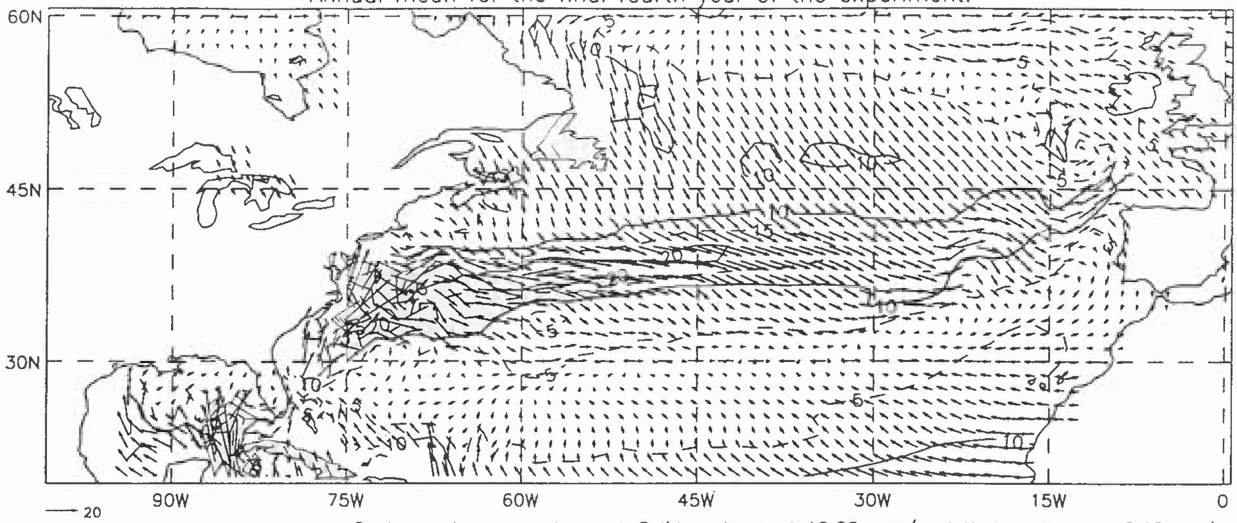
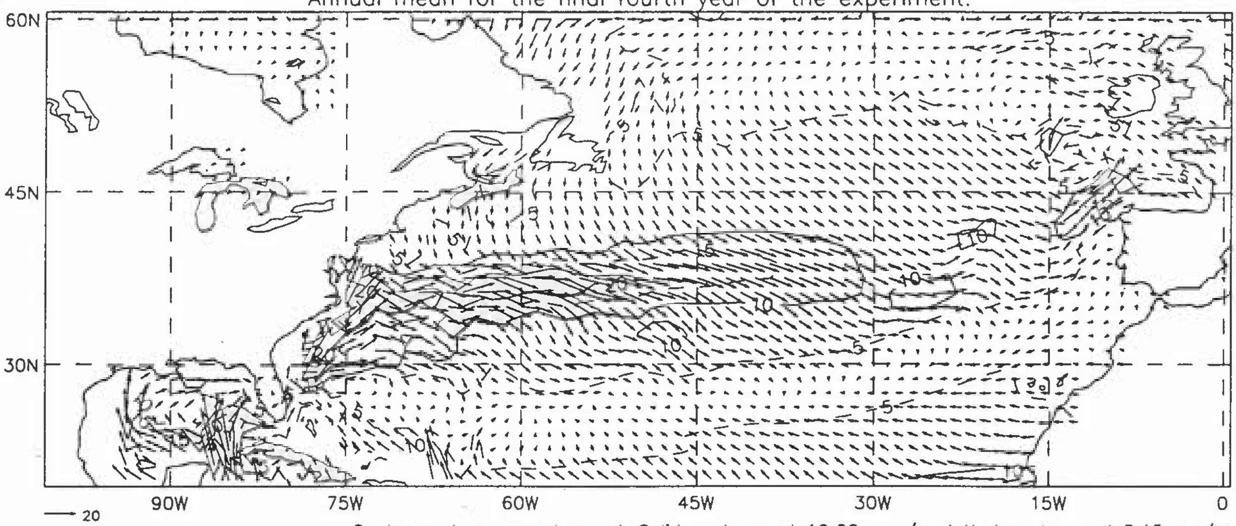


Fig. 10a. North Atlantic sea surface currents for coupled integration with standard SST relaxation forcing using 1.25 by 1.25 deg ocean and 3.75 by 2.5 deg atmosphere models.
Annual mean for the final fourth year of the experiment.



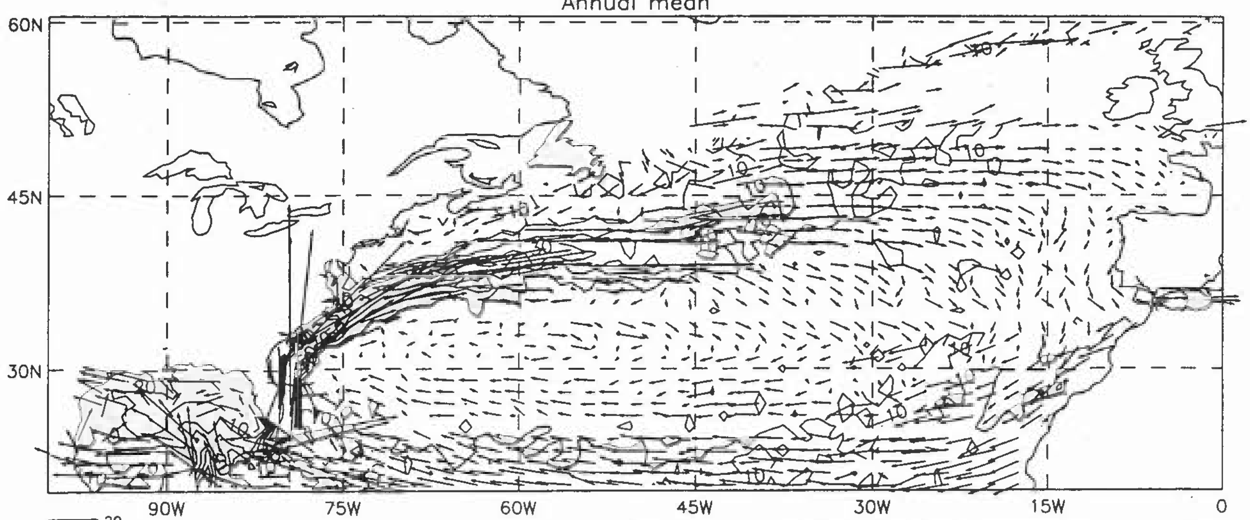

Contours show current speed. Solid contours at 10,20,.. m/s, dotted contours at 5,15.. m/s.

Fig. 10b. North Atlantic sea surface currents for coupled integration with no SST relaxation forcing using 1.25 by 1.25 deg ocean and 3.75 by 2.5 deg ocean atmosphere models.
Annual mean for the final fourth year of the experiment.




Contours show current speed. Solid contours at 10,20,.. m/s, dotted contours at 5,15.. m/s.

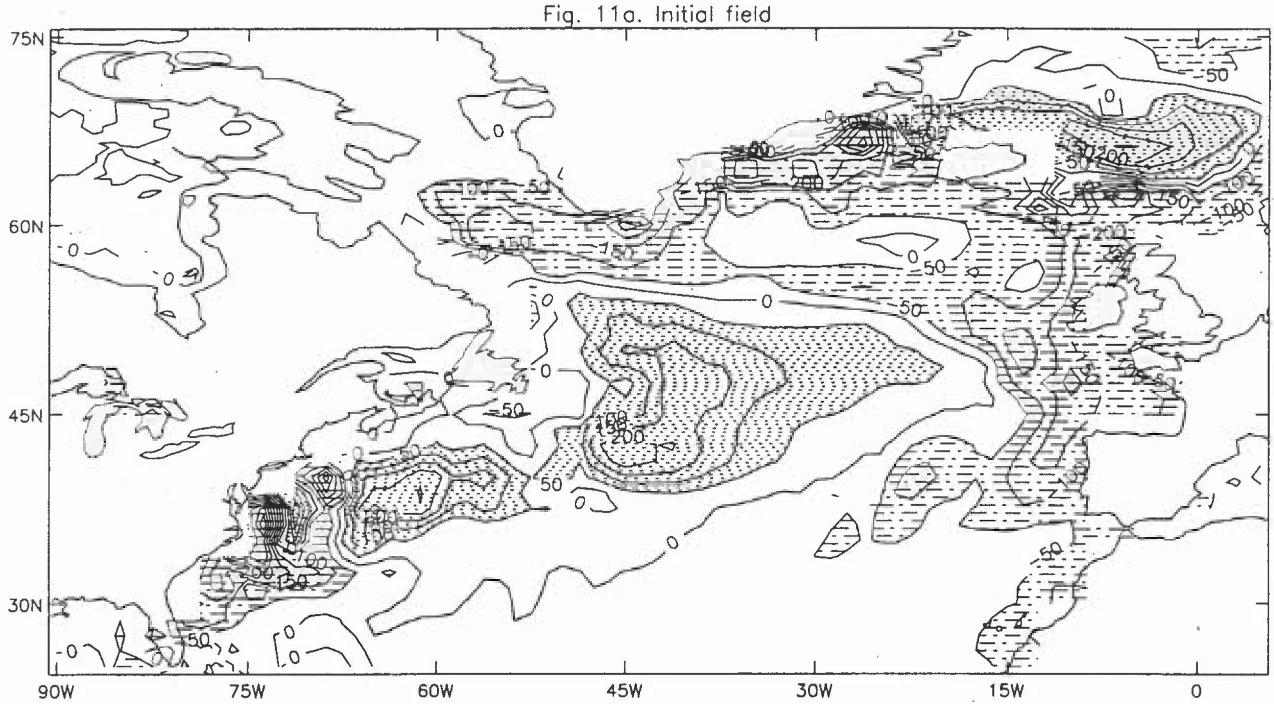
Fig. 10c. North Atlantic shipdrift currents on 1 by 1 deg grid
Annual mean




Contours show current speed. Solid contours at every 10 m/s
Every second arrow is shown for clarity

Fig. 11 Illustration of the procedure used to remove regions of high gradient in a heat flux correction field. The North Atlantic region only is shown for clarity. Fig 11a shows the initial field from the coupled integration with 1.25° by 1.25° ocean component and 3.75° by 2.5° atmosphere component. Fig 11b shows the field after the removal of regions where both the flux correction and temperature gradient are high (in this case grad T > $1.5 \times 10^{-5}$ °C$m^{-1}$ and flux correction > 50W$m^{-1}$). Fig 11c shows the field after the removal of additional surrounding points, where the flux correction is large close to high temperature gradient regions. Fig 11d shows the final field where the removed values are replaced by values typical of the surrounding open ocean. This final field is then applied in the ocean model.

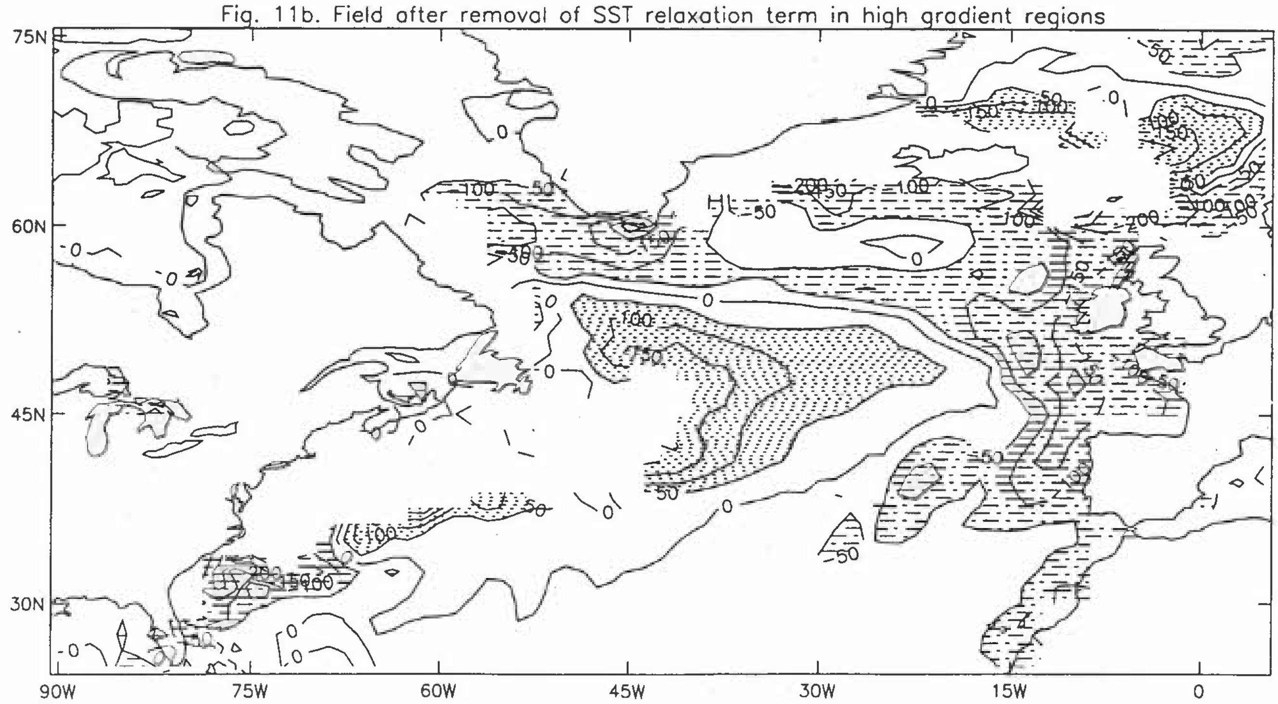
Fig. 11a. Initial field




Contour intervals of lines = 50 W/m**2

Values greater than 50 W/m**2 : dotted shading
Values less than −50 W/m**2 : dash dot dash shading

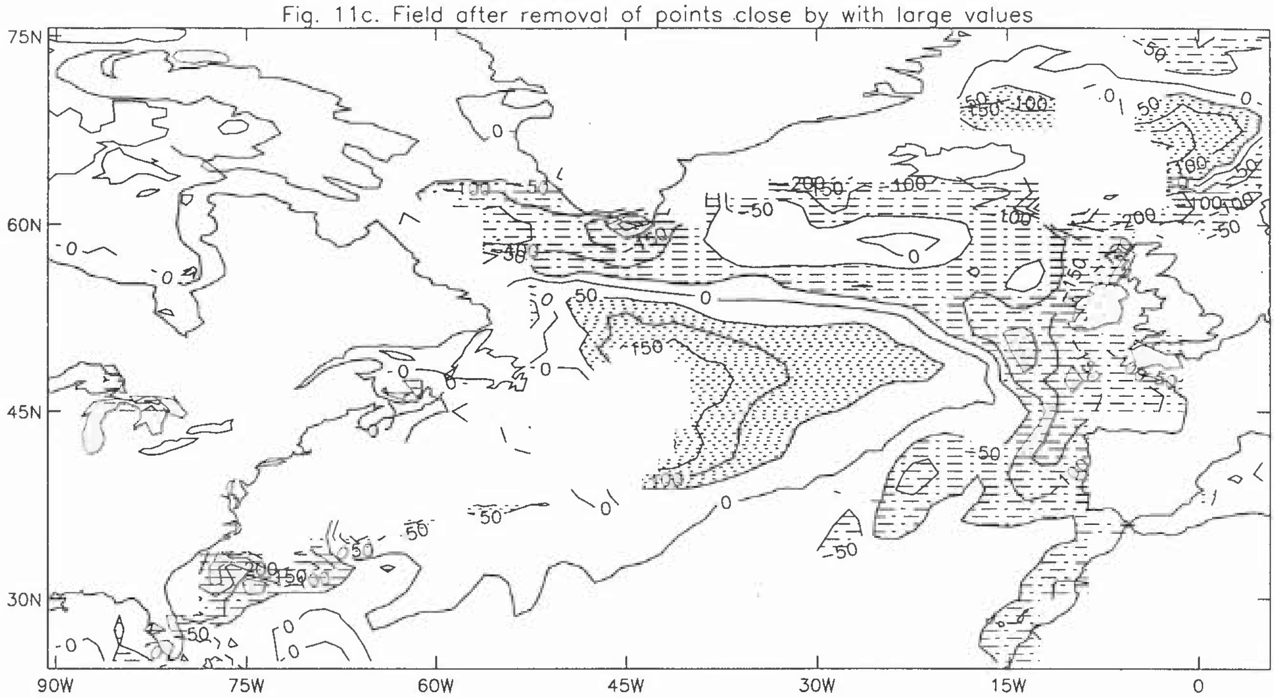
Fig. 11b. Field after removal of SST relaxation term in high gradient regions




Contour intervals of lines = 50 W/m**2

Values greater than 50 W/m**2 : dotted shading
Values less than −50 W/m**2 : dash dot dash shading

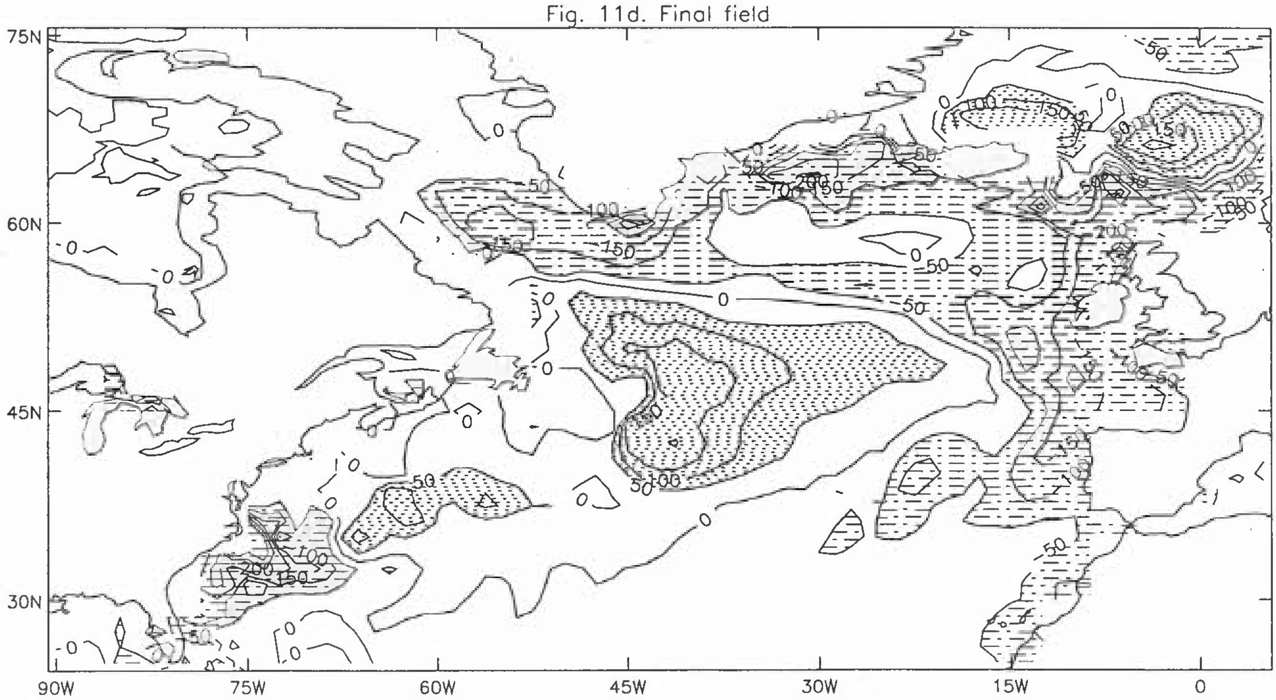
Fig. 11c. Field after removal of points close by with large values




Contour intervals of lines = 50 W/m**2

Values greater than 50 W/m**2 : dotted shading
Values less than −50 W/m**2 : dash dot dash shading

Fig. 11d. Final field




Contour intervals of lines = 50 W/m**2

Values greater than 50 W/m**2 : dotted shading
Values less than −50 W/m**2 : dash dot dash shading

Fig. 12. North Atlantic sea surface temperature for coupled integration with replaced SST relaxation forcing values in high gradient regions using 1.25 by 1.25 deg ocean and 3.75 by 2.5 deg atmosphere models. Annual mean for the final year of the experiment.

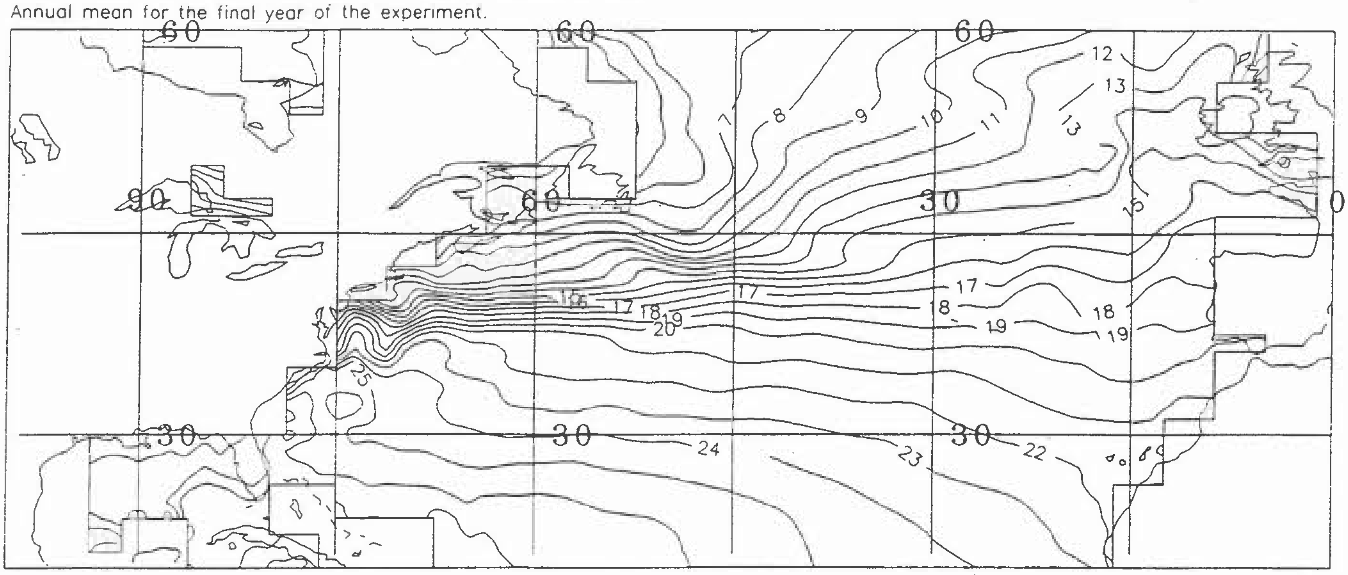
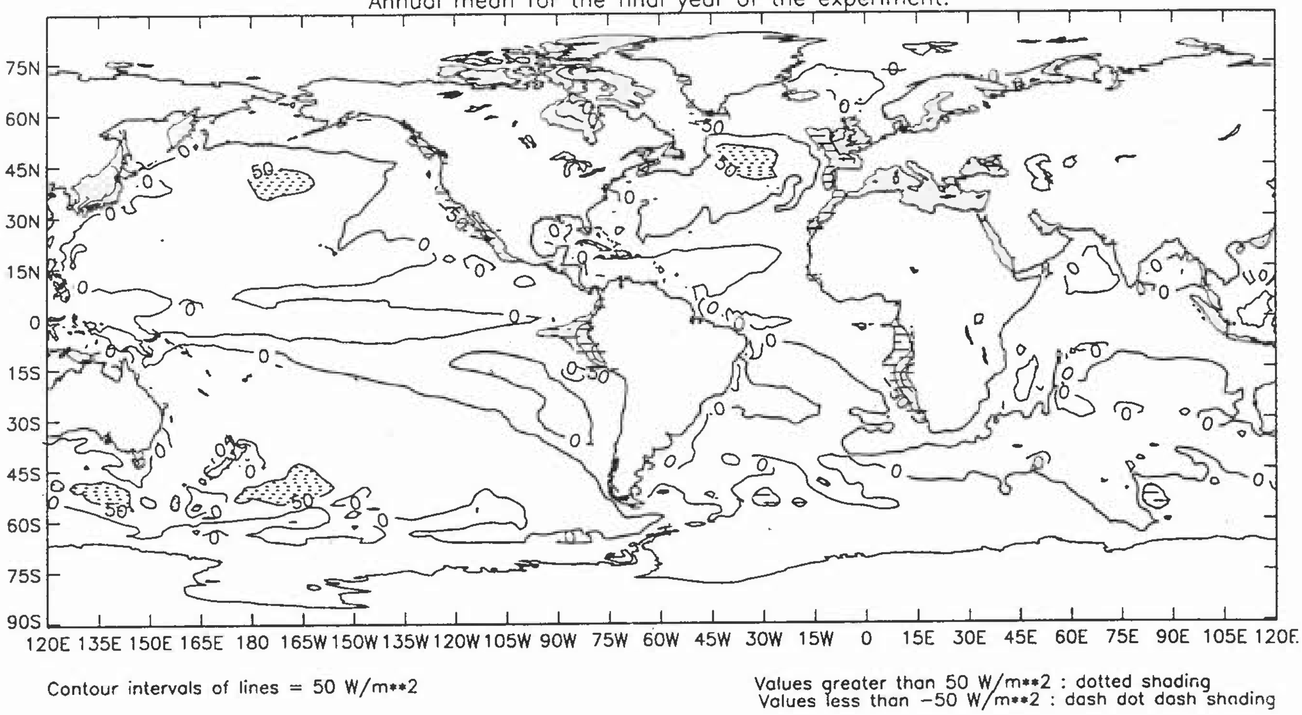



Fig. 13. Heat flux correction term for coupled integration with replaced SST relaxation forcing using 1.25 by 1.25 deg ocean and 3.75 by 2.5 deg atmosphere models. Annual mean for the final year of the experiment.




Contour intervals of lines = 50 W/m**2

Values greater than 50 W/m**2 : dotted shading
Values less than −50 W/m**2 : dash dot dash shading