Causes of Atlantic Ocean climate variability between 1958 and 1998.[1][2]

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Abstract

Numerical experiments are performed to examine the causes of variability of Atlantic Ocean SST during the period covered by the NCEP reanalysis (1958-1998). We use three ocean models. Two are mixed layer models, one with a 75m deep mixed layer and the other with a variable depth mixed layer. For both mixed layer models the ocean heat transports are assumed to remain at their diagnosed climatological values. The third model is a full dynamical ocean general circulation model (GCM). All models are coupled to a model of the subcloud atmospheric mixed layer (AML). The AML model computes the air temperature and humidity by balancing surface fluxes, radiative cooling, entrainment at cloud base, advection and eddy heat and moisture transports. The models are forced with NCEP monthly mean winds from 1958 to 1998.

The ocean mixed layer models adequately reproduce the dominant pattern of Atlantic Ocean climate variability in both its spatial pattern and time dependence. This pattern is the familiar tripole of alternating zonal bands of SST anomalies stretching between the subpolar gyre and the subtropics. This SST pattern goes along with a wind pattern that corresponds to the North Atlantic Oscillation. Analysis of the results reveal that changes in wind speed create the subtropical SST anomalies while at higher latitudes changes in advection of temperature and humidity and changes in atmospheric eddy fluxes are important.

An observational analysis of the boundary layer energy balance is also performed. Anomalous atmospheric eddy heat fluxes are very closely tied to the SST anomalies. Anomalous horizontal eddy fluxes damp the SST anomalies while anomalous vertical eddy fluxes tend to cool the entire mid-latitude North Atlantic during the NAO's high index phase with the maximum cooling exactly where the SST gradient is strengthened the most.

We compare the SSTs simulated by the ocean mixed layer model with those simulated by the dynamic ocean GCM. In the far North Atlantic Ocean anomalous ocean heat transports are equally important as surface fluxes in generating SST anomalies and they act constructively. The anomalous heat transports are associated with anomalous Ekman drifts and are consequently in phase with the changing surface fluxes. Elsewhere changes in surface fluxes dominate over changes in ocean heat transport. These results suggest that almost all of the variability of the North Atlantic SST in the last four decades can be explained as a response to changes in surface fluxes caused by changes in the atmospheric circulation. Changes in the mean atmospheric circulation force the SST while atmospheric eddy fluxes damp the SST. Both the interannual variability and the longer timescale changes can be explained in this way. While we were unable to find evidence for changes in ocean heat transport systematically leading or lagging development of SST anomalies this leaves open the problem of explaining the causes of the low frequency variability. Possible causes are discussed with reference to the modeling results.

1. Introduction

Climate in and around the Atlantic Ocean has been observed to vary in broad spatial patterns and on a variety of timescales during the Twentieth Century. In the northern regions the dominant mode of variability is the North Atlantic Oscillation (NAO, e.g. Hurrell and Van Loon 1997). The NAO involves an oscillation in atmospheric mass between the subtropics and the high latitudes. In the high index phase the Icelandic Low is anomalously low, the Azores high is anomalously high, the mid-latitude surface westerlies are strong and there is a strong storm track that trends from the United States coast towards the British Isles and Scandinavia. In the low index phase both the Icelandic low and the Azores high are weaker, the westerlies are weaker, and storms tend to move from the United States into the Labrador Sea region while those that do make it across the Atlantic move into Southern Europe and the Mediterranean. The NAO has a coherent signal in sea surface temperature (SST) involving a tripole pattern of almost zonally oriented anomalies with subtropical and high-latitude SSTs varying in phase and mid-latitude SSTs varying out of phase (e.g. Kushnir 1994, Cullen and DeMenocal 1999). Spectral analyses of NAO time series reveal enhanced variance at periodicities of around two years and at some decadal periods (Hurrell and Van Loon 1997). Further, the NAO has revealed some long term trends, most recently in the form of the tendency towards a deeper Icelandic low and stronger Azores high from 1960 to the 1990s (Hurrell 1995).

In the tropical regions the dominant mode of variability involves variations in the cross-equatorial SST gradient and SST anomalies that are off-equatorial and encompass the entire subtropical oceans (Nobre and Shukla 1996). When one hemisphere warms anomalous winds tend to blow across the equator into the warmer hemisphere. There has been some debate about whether the SSTs of the subtropical oceans vary out of phase (Houghton and Tourre 1992) with, most recently, Rajagopolan et al. (1998) concluding that the SSTs of the two hemispheres are not related to each other. The remote effects of the El Niño-Southern Oscillation (ENSO) creates off-equatorial SST anomalies in the Atlantic Ocean (e.g. Saravanan and Chang 1999, Giannini et al. 1999). In addition, the subtropical SST anomalies vary strongly on decadal timescales in a way that seems independent of ENSO. The equatorial Atlantic also contains a weak equatorial pattern of variability that is akin to the El Niño-Southern Oscillation phenomena in the Pacific Ocean but which is not self-sustained (Zebiak 1993).

Dividing Atlantic Ocean climate variability into tropical and mid-latitude modes may be useful but is not necessarily valid. For example, the NAO is associated with variations of winds and SST in the northern subtropical Atlantic Ocean. Further, Ragajopolan et al. (1999) present statistical evidence that SSTs in the subtropical South Atlantic are associated with variations in the NAO. This connection might work via the impact of South Atlantic SSTs on Amazon rainfall; the latter influencing the NAO via atmospheric teleconnections or changes in the Hadley Cell (Robertson et al. 1999).

Recently, several investigations have concluded that interannual variations of Atlantic SSTs are primarily driven by the atmosphere via changes in surface fluxes. The concept of flux driven SST variability was first suggested on the basis of analyses of SSTs and marine meteorological data by Cayan (1992a,1992b) and has been supported by modeling studies (Battisti et al. 1995, Delworth and Mehta 1998, Luksch 1996, Halliwell 1998). Barsugli and Battisti (1998) and Bladé (1999) have shown that coupling to an ocean mixed layer, and hence coupling between the atmosphere, surface fluxes, and the SST, is an important process that enhances the variance of low level thermal fields in the atmosphere, and can lead to modest persistence.

In contrast to the dominant role of the atmosphere on interannual timescales, it has been suggested that the longer timescale variations might involve a more active role for the ocean including changes in ocean heat transport (e.g. Deser and Blackmon 1993, Kushnir 1994, Grötzner et al. 1998). Appealing to an active role for the ocean is attractive in that the long timescales associated with ocean dynamics make it easy to explain decadal fluctuations and long periods of persistent oceanic anomalies. Others have suggested that the ocean's role is largely restricted to the ability of near surface mixing to sequester heat content anomalies from one winter to another below the summer mixed layer (e.g. Battisti et al. 1995, Bhatt et al. 1998).

Explanations for low frequency variations which invoke atmosphere-ocean coupling require that the mid-latitude atmosphere be responsive to underlying SST anomalies. The latter has proven elusive to demonstrate. Some models do show a coherent response to North Atlantic SST anomalies (e.g. Grötzner et al. 1998, Ferranti et al. 1994, Rodwell et al. 1999) while others do not (e.g. Pitcher et al. 1988, Lau and Nath 1994; see also Peng et al. 1995 and the review by Kushnir and Held 1996). To date it has not been resolved whether different models respond differently to the same SST anomalies or whether the apparently inconsistent results are explained by differences in the imposed SST anomalies, experimental design, length of integration and so on.

In this paper we will report on efforts to understand the variability of Atlantic Ocean climate from 1958 to 1998, which is the period for which reliable atmospheric data are available from the National Centers for Environmental Prediction (NCEP) Reanalysis (Kalnay et al. 1996). We attempt to model the SST over this period using ocean mixed layer models, in which the ocean heat transport is held at its climatological value, and also with a fully dynamical ocean general circulation model (GCM). All models are coupled to a simple thermodynamic model of the well mixed atmospheric mixed layer (AML) that forms the lower component of the marine convecting boundary layer (Seager et al. 1995). In this manner, the models are forced only by the time-varying wind speed and direction, while the SST and the boundary layer temperature and humidity are computed according to balances between the surface fluxes, ocean heat transport (if allowed to vary), advection and eddy transports of heat and moisture in the atmospheric mixed layer, entrainment across the top of the atmospheric mixed layer and radiative cooling. Since, in nature, the atmospheric temperature and humidity and SST equilibrate to each other on timescales of a day or so, imposing the atmospheric thermodynamic state in the heat flux boundary conditions of an ocean model informs the model what the SST was. While this is commonly done in ocean modeling studies it ensures that the simulated SST will track that observed while making interpretation of that result confusing. In the ocean modeling work reported here we instead attempt to properly model the coupling between the ocean and the atmospheric boundary layers. This is very clearly an improved experimental set up that allows the SST full freedom to evolve (Seager et al. 1988). We will examine the extent to which surface fluxes and ocean heat transport determine the SST variability and, in turn, why these components of the ocean surface heat budget vary. To what extent can variations in Atlantic Ocean climate be understood as the atmosphere forcing the ocean or vice versa?

There have been two previous attempts to look at North Atlantic SST variability in forced ocean GCM simulations. Luksch (1996) and Halliwell (1998) both concluded that surface flux anomalies were responsible for most of the variability but that anomalous Ekman advection was also important in the region of mean surface westerlies. In addition, Halliwell (1998) suggested that changes in ocean heat transport, other than those associated with Ekman dynamics, were important in the Gulf Steam region (as also suggested by Deser and Blackmon (1993)). The current work expands on these previous studies by simulating a longer period with the recent NCEP reanalysed forcing. We also use a more complete treatment of surface fluxes than either Halliwell (1998), who assumed the latent heat flux to be merely damping, or Luksch (1996) who assumed the surface relative humidity did not alter. Both previous studies, therefore, cannot properly account for how changes in moisture advection might impact SSTs. We also provide a direct comparison between the results of a full ocean GCM and simpler ocean models that allows for an easy assessment of the relative roles of different ocean processes.

Before reporting on the ocean model simulations we begin by using NCEP reanalyses to examine the terms in the thermodynamic energy budget of the lowest level of the atmosphere. This allows us to assess the different roles that changes in wind speed, advection, subsidence and atmospheric eddy transports have in generating the flux anomalies that influence the SST. In Section 3 we present some preliminary calculations in which we use the atmospheric mixed layer model to simulate the observed changes in surface latent and sensible heat flux given the observed SST. The modeled fluxes are in good agreement with those observed so, in Section 4, we couple the AML model to a uniform depth ocean mixed layer in which the ocean heat transport is assumed to remain at its climatological, seasonally varying, values. The coupled AML-OML model is used to simulate the SST from 1958 to 1998 forced by the time-varying NCEP wind speed and direction. This experiment is analyzed and demonstrates that much of the observed SST variability can be explained in terms of surface fluxes without the need to invoke changes in ocean heat transport. We then repeat this calculation using a variable depth ocean mixed layer in order to assess the role of ocean mixing. In Section 5 we model the SST from 1958 to 1998 using the full ocean GCM which allows the mixed layer depths and ocean heat transport to vary. This run is analyzed, in comparison to the AML-OML experiments, to isolate the role of ocean heat transport. Conclusions are offered in section 6.

2. Observational analyses of atmospheric boundary layer thermodynamic budgets and SST forcing

Before attempting to model Atlantic SSTs we perform an analysis of the thermodynamic budget of the lowest part of the atmosphere. We wish to examine which terms in the budgets are responsible for the changes in surface fluxes that force changes in SST. We use the NCEP reanalyses for the period 1958-1998. We consider a layer, assumed vertically uniform, that extends from 925mb to 1000mb which we take to be representative of the atmospheric mixed layer that forms the lower portion of the boundary layer. We assume there is a well-mixed layer extending from 1000mb to 925mb which can be characterized by the 1000mb values. This is a well-justified assumption (e.g. Norris 1998). We assume a steady state because the mixed layer adjusts to the underlying SST on timescales of less than a day (Boers and Betts 1988). Integrating from 1000 to 925 mb, the moist static energy equation is:

$$\frac{1}{g}\left(P\underline{u}\cdot\nabla h-\omega_B(h_B-h)+P\nabla\cdot(\underline{u}'h')-(\omega'h')_B\right)=\frac{\omega_0}{g}(h_0-h)+\frac{1}{g}(\omega''h'')_B+\frac{1}{g}c_pPR. \quad (1)$$

Here h is the moist static energy, ω_B is the pressure velocity at 925mb, $\omega_0=\rho C_D V$ where V is the surface wind speed and C_D is the drag coefficient, h_0 is the moist static energy of the ocean surface and h_B is that at 925mb, P is the pressure thickness of the layer (7500Pa) and R is the radiative cooling rate in Ks^{-1}. In this equation every term is a monthly anomaly (and includes both linear terms and nonlinear cross terms). For example the term $\underline{u}\cdot\nabla h$ equals the advection evaluated for a month, using total values of u and h, minus the climatological monthly mean of that term. $\nabla \cdot (\underline{u}'h')$ is the anomalous horizontal convergence of moist static energy by eddies with timescales less than one month, $(\omega'h')_B$ is the anomalous vertical eddy moist static energy flux at 925mb. $(\omega''h'')_B$ is the anomalous vertical turbulent flux of moist static energy at 925mb. The first term on the right is the anomalous ocean to atmosphere moist static energy flux.

The turbulent flux and the radiative cooling rate are unknown and cannot be calculated. In the case of the radiative cooling rate this is because we are at a loss to know what the details of the cloud field were. We do not expect anomalies in the radiative cooling rate to be large but anomalies in the turbulent flux are expected to be significant. Nevertheless we are particularly interested in how changes in advection and surface and eddy fluxes impact the boundary layer temperature and humidity and, therefore, the SST. To do this we first computed Empirical Orthogonal Functions (EOFs) of the NCEP observed SSTs. In this EOF analysis, and the subsequent Singular Value Decomposition (SVD) analyses, we use area-weighting of the analysed fields. We then regressed the individual terms of the boundary layer thermodynamic budget onto the time series of the EOF expansion of SST. We only present results for the first SST mode. This mode is the familiar tripole pattern of SST anomalies that accompanies the NAO, which is virtually identical to the SST pattern emerging from the SVD analysis discussed later, and explains 25% of the domain integrated SST variance. The figures shown are for the positive phase of the NAO when there is an anomalously strong anticyclone over the subtropical North Atlantic, a strong Icelandic low and strong mid-latitude westerlies between them. The energy budget terms, as written in Equation 1, are in Wm^{-2}. The values plotted correspond to the flux anomaly that accompanies one standard deviation of the normalized SST anomaly time series. In Figure 1 the flux anomalies are contoured over the one standard deviation SST anomalies plotted in color.

The anomalous advection presents a simple pattern (Fig. 1a). When anomalous $\underline{u} \cdot \nabla h$ is positive this represents a cooling of the boundary layer and, hence, the SST. We see that anomalous advection matches the SST pattern quite well with cool water present where there is equatorward advection in the southeastern North Atlantic and warm water present where there is poleward advection off the North American coast. Further north, anomalous flow off the cold Canadian coast and Labrador Sea area leads to cold waters offshore. The signal in the South Atlantic is weak.

Figures 1c and 1d show the horizontal and vertical eddy flux convergence. The total eddy flux convergence primarily acts to cool the warm waters east of North America. Reduced eddy fluxes warm the northern subtropics, indicative of a poleward shift of the region of maximum eddy heat and moisture fluxes. The individual eddy terms show that the horizontal fluxes almost perfectly damp the SSTs, while the vertical fluxes have maximum cooling at around $45°N$ where the SST gradient is strengthened the most. Therefore, when summed, the cooling over, and slightly to the north of, the warm water is the dominant signal. Even if the eddy activity did not depart from its climatological nature, it would be expected that the eddy flux would have this effect on the SSTs. We found that the low level air temperature and humidity anomalies closely track the SST anomalies. Since the eddy heat and moisture transports are always down gradient, they strengthen and weaken as the SST gradient does and therefore damp the SST anomalies. What is more interesting is that the eddy momentum transports also vary in a systematic way with the SST (not shown) which will be the subject of future work. Generally the match between the eddy fields and the SST anomalies is quite remarkable.

The term $\omega_B(h_B - h)$, shown in Fig. 1b, represents anomalous subsidence warming and drying and is typically the same sign as ω itself. The anomalously strong anticyclone over the North Atlantic, which is also poleward of its usual position, leads to anomalous subsidence at around $35°N$ and weaker subsidence to the south. Increased subsidence cools the SST, primarily via increased latent heat flux, by bringing down air of lower moist static energy. Changes in subsidence primarily damp the SST fluctuations.

We broke the surface flux term into two terms, the anomalous wind speed working on the mean vertical gradient of moist static energy, and the mean wind working on the anomalous vertical gradient of moist static energy. The latter term includes how the changing SST influences the moist static energy budget. In the case of the atmosphere forcing the ocean via changes in wind speed, this will be a negative feedback term that opposes the wind-induced flux change. It was assumed that the nonlinear cross term was small. Figs. 1e and 1f show the regression of these terms on the SST. The flux anomaly derived from the anomalous wind speed working on the mean thermodynamic gradients perfectly matches the SST change and increases in size from pole to equator. The effects of anomalous thermodynamic gradients are more complex. In the subtropics, this gives a flux anomaly that damps the SST: increased wind speed cools the SST, and h_0 is reduced by more than h, implying that, as increased surface heat loss reduces the SST, the air-sea thermodynamic disequilibrium is also reduced, which tends to reduce the surface heat loss in an attempt to restore balance. Here we are seeing the component of the surface flux anomaly that involves the ocean's SST response to the increase in wind speed. North of $30°N$ this damping effect is less obvious. In these regions anomalous advection causes anomalies in h that force SST changes.

All of these terms are of significant magnitude somewhere. Nonetheless, it is possible to draw some simple conclusions. In the subtropics wind speed changes drive changes in SST. The altered SST then creates changes in the surface fluxes that largely offset those created by the wind speed changes. North of about $25°N$ advection is important with, for the positive phase of the NAO, advection cooling the eastern Atlantic and warming the western Atlantic. Wind speed changes tend to warm the whole strip between $25°N$ and $45°N$. Subsidence drying also tends to cool the east. Summing these effects explains why the west warms but the SST anomalies in the east are small. Over the entire mid-latitude zonal strip atmospheric eddies primarily, and strongly, damp the SST anomalies. North of $45°N$, anomalous advection off North America cools the SST with the reduced SST feeding back by restricting the surface heat loss. Therefore, the SST anomalies are forced by changes in the mean flow and are dissipated by transient eddy fluxes of heat and moisture. This does not exclude the possibility that the changes in the mean flow are forced by changes in eddy momentum fluxes, but that will have to await further investigation.

3. Simulation of surface heat flux variability between 1958 and 1998

The results of the previous section suggest that it may be possible to model the observed fluxes with a simple atmospheric model that balances surface fluxes, advection, subsidence, eddy transports and radiation. The results also suggests that a model that parameterized changes in eddy fluxes in terms of changes in SST gradients would capture much of the observed eddy effects. Here we describe efforts to simulate the variability of surface latent and sensible heat fluxes over the 1958 to 1998 period using a model of the AML forced by the observed SSTs. The model is described in detail in Seager et al. (1995). It represents either a dry convective layer or the subcloud layer that underlies marine clouds. Within this layer it determines the virtual potential temperature and specific humidity by balancing advection, eddy transports, the fluxes at the surface and the atmospheric mixed layer top and, for temperature, radiative cooling. The model assumes a steady state because of the rapid adjustment time of the mixed layer as previously discussed.

We use the usual bulk formulae to compute the surface fluxes. The closure for the flux of virtual potential temperature at the mixed layer top sets the downward flux to be a fixed proportion of the surface flux. This has been justified on the basis of data analysis (Nicholls and LeMone 1980), modeling (Betts 1976) and theory (Tennekes 1973) and has been used extensively in models of marine boundary layers (e.g. Bretherton 1993, Betts and Ridgway 1989, Albrecht et al. 1979, Clement and Seager 1999). The radiative cooling is assumed to be a constant $2Kday^{-1}$. The closure for the moisture flux is more empirical and simply ensures that, in the absence of advection, the mixed layer relative humidity will be close to 80% as observed.

With these assumptions the model equations for the total fields of virtual potential temperature and specific humidity are (see Seager et al. 1995 for a complete derivation):

$$P\left(\underline{u}+\underline{u}^{*}\right)\cdot\nabla\theta_{v}=(1+\beta_{V})C_{0}\omega_{0}(\theta_{V0}-\theta_{V})+PR^{'}, \tag{2}$$

$$P\left(\underline{u}+\underline{u}^{*}\right)\cdot\nabla q=C_{0}\omega_{0}q_{0}-C_{0}\omega_{0}(1+\mu)q, \tag{3}$$

$$\theta = \theta_V/(1 + .61q). \tag{4}$$

Here P is the fixed mixed layer pressure thickness (a typical value for the subcloud layer of 6000Pa is assumed), θ_V is the virtual potential temperature and θ_{V0} is its surface value, q is the specific humidity and q_0 is the saturation specific humidity at the surface temperature. θ is the potential temperature, R' is $(1 + .61q)$ times the radiative cooling, C_0 is the surface exchange coefficient and $\omega_0 = \rho g C_D V_0$, where C_D is an exchange coefficient and V_0 is the surface wind speed. β_V is the closure parameter that determines the virtual potential temperature flux at the mixed layer top (see Betts 1976). μ is a parameter related to the closure on the moisture flux at the mixed layer top and is set so that, in local equilibrium ($q = q_0/(1 + \mu)$), the modeled relative humidity is close to the observed value of 80%. The advecting velocity includes the NCEP analyzed monthly mean 1000mb wind $\underline{u}$ and an eddy advecting velocity $\underline{u}^*$. The latter is assumed to be proportional to the surface temperature gradient (over land and sea) averaged over a distance 10° north and south of the grid point and extending 60° west. This is designed to mimic the effects of unresolved sub-monthly transient eddies which advect cold dry air equatorward and down and move warm moist air poleward and up (see also Kushner and Held 1998).

We first use the model to compute the turbulent fluxes using NCEP reanalyzed monthly averaged 1000mb wind speed and direction prescribing the monthly averaged observed NCEP SSTs. The model also uses NCEP 1000mb air temperature and specific humidity over land. These are needed where the winds blow offshore, in which case observed values are advected out over the ocean. The model computes the total surface fluxes given the total SSTs and anomalous fluxes are computed by subtracting the climatological mean modeled fluxes. We assume the cloud cover does not depart from its climatological seasonal cycle. The longwave cooling of the ocean surface is estimated using bulk formula and varies as the SST and air temperature and humidity vary but its variations are much smaller than the variations in the total turbulent flux.

To examine the agreement between model and observed fluxes we look at two winter averages of January through March for 1969 and 1989 that exhibited opposite extreme states of the NAO. (We chose to check the model's ability to simulate the fluxes by examining specific winters, rather than an EOF for example, to emphasize that the climate variability we are talking about is clearly apparent in the raw data.) Fig. 2 shows the anomalies of SST, surface winds and upward latent plus sensible heat fluxes as derived from the NCEP reanalyses for these two winters. The latent and sensible flux anomalies are almost always the same sign. In the tropics the latent heat flux anomalies dominate, but at high latitudes they can be close to the same magnitude. Generally the anomalous turbulent fluxes are of the sign that would create the SST anomalies, i.e. anomalously positive fluxes above cold water. This relationship was noted by Cayan (1992a, 1992b) and is indicative of the atmosphere forcing the ocean. Figure 3 shows the modeled latent plus sensible heat flux when forced with observed SST. It is clear that the AML model does a reasonable job of reproducing the amplitude and spatial pattern of the observed flux anomalies.

The flux anomalies could be produced by changes in wind speed or direction. Next we ran the AML model holding the wind speed and direction fixed in the advection terms but allowing the wind speed to vary in the surface flux formulation. Then we allowed the advecting winds to vary but held the wind speed in the surface flux formulation fixed. The modeled latent plus sensible fluxes for these cases are shown in Figures 4a-d. It is clear that changes in wind speed are the dominant effect south of $40°N$, but that at higher latitudes changes in advection of temperature and moisture become important. These results are broadly consistent with the observational analyses of Cayan (1992a,1992b).

4. Simulation of SST anomalies with a uniform depth ocean mixed layer coupled to the AML model

Before we examine the SST anomalies simulated using the AML model we will consider the results derived by forcing an ocean mixed layer model with surface fluxes evaluated using bulk formulae and *observed* air temperature and specific humidity. It is quite common to use observed air temperature and humidity in ocean model boundary conditions and a simulation of the Atlantic Ocean that uses this design has been presented by Battisti et al.

(1995). We simulated the global SST anomalies from 1958 to 1998 forcing a 75m deep ocean mixed layer with surface fluxes computed with bulk formulas and the modeled SSTs and the NCEP observed air temperature and humidity. Figure 5 shows the global map of correlation coefficients between the time series of observed and modeled SST anomalies. The correlation is good to excellent almost everywhere. The amplitude of the modeled SST anomalies, as estimated by regression (not shown), are also reasonable. This might be taken to indicate that almost all the SST variability is flux-driven. However, we also see that the correlation is good in the tropical Pacific Ocean where we know that the SSTs are actually driven by changes in ocean heat transport.

This result is inevitable. The air-sea temperature difference is constrained to be whatever is needed to balance the radiative cooling of the subcloud layer (e.g. Betts and Ridgway 1989). The radiative flux divergence across the subcloud layer varies by very little and is typically about $10Wm^{-2}$. The sensible heat flux at cloud base is downward and typically small (e.g. Betts 1976). Since condensation and evaporation of falling rain are small terms in the subcloud layer, the SST and air temperatures must adjust to provide a surface sensible heat flux that balances the radiative flux divergence. Therefore the surface flux must also be about $10Wm^{-2}$ which requires an air-sea temperature difference on the order of 1K. Similarly the air-sea humidity difference always adjusts such that the surface relative humidity remains at about 80%, as the evaporation and entrainment of air at cloud base come into balance. The reasons for this are less clear than for the case of the air-sea temperature difference but this uniformity of surface relative humidity is nonetheless an undisputed fact of life for the marine boundary layer.

Consequently, the air temperature and air humidity are imprinted with the SST. Specifying them in the ocean model boundary conditions guarantees that the SST will approach its observed value and ensures that even ENSO related SST changes are simulated despite the ocean heat transport remaining fixed. Of course in this model the ENSO related SST changes are caused by changes in the surface fluxes and this is not so in the real world. Therefore, it may be possible to sort out the roles of surface fluxes and ocean heat transport in this experimental arrangement but only by simultaneously comparing modeled and observed SSTs and surface fluxes. This is what Battisti et al. (1995) attempted to do. Nonetheless, this remains a methodology that is prone to be ambiguous and misleading. Since the air temperature and humidity and SST equilibrate to each other in all circumstances this argument is valid in cases where the atmosphere is forcing the ocean as well as in cases where the ocean is forcing the atmosphere (e.g. ENSO). Clearly, when we are interested in simulating and understanding SST variability, it makes sense to explicitly model the coupling of the atmospheric and oceanic boundary layers and to avoid specification of the atmospheric thermodynamic state.

a. Simulation with a uniform depth ocean mixed layer

The success of the AML model's surface flux simulation suggests that it may be possible to simulate SST anomalies by coupling the AML model to a model of the ocean mixed layer. We assume the simplest of ocean mixed layers: a uniform, well mixed, 75m deep layer that is decoupled from the water below. Knowing the climatological mean fluxes from the runs with imposed SSTs we integrate an equation for the SST anomaly that is forced by flux anomalies. The anomalies are computed by subtracting the modeled climatological mean fluxes from the total flux computed by the AML model using the observed climatological mean SST plus the modeled SST anomaly. Here, and in all the experiments to be described, we assume that the surface solar flux and the cloud cover remain at their climatological seasonal cycle. It is well known that in mid-latitudes variations in the surface radiative fluxes are much smaller than those in the turbulent fluxes (Cayan 1992b) and our own estimation of these quantities using NCEP data confirmed this. The equation for the SST anomaly is:

$$\frac{\partial T'}{\partial t} = \frac{1}{\rho c_p H}\left[Q(T' + \bar{T}_{obs}) - Q(\bar{T}_{obs})\right]. \tag{5}$$

Here T' is the SST anomaly, H is the mixed layer depth, $\bar{T}_{obs}$ is the observed climatological mean SST, $Q(T' + \bar{T}_{obs})$ is the total heat flux and $Q(\bar{T}_{obs})$ is the modeled climatological heat flux. It should be understood that the total fluxes are computed using, not only the total (modeled anomaly plus specified mean) SST, but also the total wind speed and direction from the NCEP reanalyses for the appropriate time in the 1958-1998 period. This procedure is equivalent to holding the ocean heat transports fixed at their seasonally varying climatological values. The model is initialized with the SST anomaly in January 1958 and advanced forward through to December 1998. Time varying temperature and humidity over land are used as boundary conditions and get used when the flow is offshore. In this and the subsequent experiments the model spans the Atlantic Ocean from $30°S$ to $73°N$ with a resolution of $2°$ by $2°$. The model is integrated through all the months and years but we will only examine the results for the January to March winter season which is when North Atlantic climate variability is most apparent.

In order to evaluate the observed and modeled variability we perform analyses by singular value decomposition (SVD; Bretherton et al. 1992) between SST and vector wind. The first observed mode of variability, which explains 25% of the variance of SST and 25% and 20% of the variance of zonal and meridional winds respectively during the January to March season, is shown in Figure 6a and the SST time series in Figure 6c. The corresponding modeled patterns and SST time series are shown in Figures 6b and 6c. The observed first mode of variability shows the familiar tripole pattern of SSTs. When the waters are anomalously cold in the high latitude North Atlantic and subtropical North Atlantic they are anomalously warm in the mid-latitudes. The wind pattern that accompanies the SST field shows an anomalous anti-cyclonic circulation whose influence spans the entire North Atlantic. This circulation strengthens the northeast trades inducing ocean cooling. At higher latitudes the mid-latitude westerlies are shifted north compared to climatology and intensify, inducing cooling in the high latitude North Atlantic and warming in the mid-latitude Atlantic. As mentioned in Section 2 advection is also important in some areas. For example, advection of warm moist air poleward aids the warming of waters immediately off the United States coast. The pattern is the North Atlantic Oscillation.

The first mode of the modeled variability during winter is strikingly similar to that observed. The modeled first mode explains 23% of the variance of modeled seasonal SST anomalies. However, a close inspection reveals some differences: the warm anomaly in the mid-latitude Atlantic extends further south in the observations and also has an axis of maximum values in the location of the North Atlantic Current, the cooling of the subpolar gyre is too intense and the model also poorly represents SST anomalies in the Gulf of Guinea. However, the time series of the observed and modeled modes are in good agreement with the model capturing, not only the interannual variability of this pattern, but also the trend from the 1960s to the 1990s.

b. Simulations with a variable depth ocean mixed layer

In this section we simulate the SST anomalies using a variable depth ocean mixed layer while still holding ocean heat transports fixed at their climatological values. The mixed layer depth was diagnosed from the computed values in the full GCM experiment described below. We then compute a climatological seasonal cycle of GCM mixed layer depths at each model grid point and impose these in the SST calculation. Therefore, as with the case of the uniform depth layer, the SSTs are still decoupled from the water below. The SST anomalies evolve according to the schematic equation:

$$\frac{\partial T'}{\partial t} = \frac{1}{\rho c_p H(x,y,t)} \left[Q(T' + \bar{T}_{obs}) - Q(\bar{T}_{obs}) \right]. \tag{6}$$

Where now $H(x,y,t)$ is the specified, spatially and temporally varying, ocean mixed layer depth.

The spatial patterns of the first mode of variability during winter which explains 23% of the variance of SST, and its time series, as derived by an SVD analysis of modeled SSTs and observed winds are shown in Figures 7a and 7c. This mode is very similar to that derived with a uniform depth mixed layer but there are important differences. In the North Atlantic the amplitude of the modeled SST anomalies generally decreases. This is because the modeled mixed layer depth is much deeper than 75m so that the same flux anomalies generate smaller changes in SST. The deeper modeled mixed layer depths are more realistic. The SST anomalies simulated in the far North Atlantic with the uniform depth layer were too large but, using the variable depth mixed layer, they are now too small, for example, in the region south and west of Iceland. However, south of Greenland they remain too large. We compared the modeled mixed layer depths with those derived from the Levitus (1982) data and found that the model underestimates the depths over much of the Atlantic north of $40°N$. This difference explains why the model SST anomalies are too large south of Greenland but the underestimated SST anomalies south and west of Iceland suggest that other processes, such as ocean advection, must be contributing to the SST anomalies there.

In the subtropical South Atlantic, use of the variable depth mixed layer increases the size of the SST anomalies because the mixed layers in this region, where it is local summer, are shallower than 75m. The modeled SST anomalies would, however, appear to be too large. Comparing the modeled mixed layer depths with those derived from Levitus (1982) data reveals that this is because the model depths are somewhat too small. Spatial and temporal variations in the mean mixed layer depth, which are hard to model correctly, nonetheless have an important effect on the the SST variability.

5. Simulation of SST variability with an ocean GCM

In order to see how changes in ocean heat transport impact the evolution of SST anomalies we integrated the Lamont ocean GCM (Visbeck et al. 1998), coupled to the AML model, for the 1958-1998 period forced by the NCEP reanalyzed wind stresses as well as the wind speed and direction that are used within the AML. The GCM spans the Atlantic Ocean from $30°S$ to $73°N$ with a resolution of 2° by 2°, and 30 fixed vertical levels, 13 of which are in the upper 1000m. The model includes basin geometry and bathymetry consistent with the resolution. Model temperatures, at all depths, are relaxed towards seasonally varying climatological values within 5° of the northern and southern ends of the domain only. Salinity is restored to observed values at all grid points so the influence of salinity variability is ignored. The model includes a simple one and a half layer thermodynamic sea ice model, a bulk wind-driven mixed layer model, convective adjustment and isopycnal thickness diffusion. The mixed layer depth is now computed and can deviate from its climatological values.

When the model is forced with the total (climatology plus anomaly) forcing fields from the NCEP reanalyses it produces quite substantial errors in the annual mean SST which is in contrast to the model simulations using other forcing products (e.g. European Center for

Medium Range Weather Forecasts analyses). While the seasonal and lower frequency anomalies around this incorrect mean are quite realistic, we attempt to avoid potential problems by applying instead a correction in the form of a diagnosed seasonally varying mean surface flux that ensures the model reproduces a reasonable mean seasonal cycle of SST. The equation for the GCM's SST, T, can be written schematically as:

$$\frac{\partial T}{\partial t} + OHT = \frac{1}{\rho c_p H} Q. \tag{7}$$

Where OHT is the dynamical contributions, including mixing, to the SST tendency and Q is the surface heat flux. First we diagnose the surface flux, Q_{corr}, for which the model, forced by observed winds, reproduces the observed SST, $\bar{T}_{obs}$:

$$\frac{\partial \bar{T}_{obs}}{\partial t} + \overline{OHT} = \frac{1}{\rho c_p H} Q_{corr}. \tag{8}$$

$\overline{OHT}$ is the ocean heat transport from this run. The quantities in this equation were derived from a run using the monthly data for the entire 1958-1998 period and then averaging to derive monthly climatological means. The equation the model then integrates in order to derive SST anomalies relative to the observed climatological means is:

$$\frac{\partial T}{\partial t} + \overline{OHT} + OHT' = \frac{1}{\rho c_p H} \left(Q_{corr} + \left[Q(\bar{T}_{obs} + T') - Q(\bar{T}_{obs}) \right] \right). \tag{9}$$

Here OHT' is the anomalous ocean heat transport and mixing. Subtracting the last equation from the previous one we see that the SST anomaly evolves as:

$$\frac{\partial T'}{\partial t} + OHT' = \frac{1}{\rho c_p H} \left[Q(\bar{T}_{obs} + T') - Q(\bar{T}_{obs}) \right], \tag{10}$$

which is the same as for the mixed layer models except for inclusion of anomalous ocean heat transport in addition to vertical mixing.

Figure 7b shows the winds and modeled SST anomaly corresponding to the first mode of modeled SST variability in the GCM. Figure 7c shows the corresponding time series. It is immediately apparent that the pattern is very similar to both the pattern derived by the ocean mixed layer models and to the observations. There are however a few differences. The GCM now more faithfully reproduces the magnitude of the SST anomalies in large areas of the far North Atlantic which were overestimated by the uniform depth ocean mixed layer model and underestimated by the variable depth model. It is also apparent that the GCM faithfully reproduces the warm SST anomalies that occur south of the North Atlantic Current between $25°N$ and $30°N$. However, the GCM has problems simulating the variability north of the Gulf Steam and off the coast of New England and Canada and, generally, it produces too much variability in this region.

To compare the roles of surface fluxes and ocean heat transport in determining the SST variability we performed two regressions of the modeled surface heat fluxes and anomalous ocean heat transports, integrated down to the base of the modeled ocean mixed layer, each against the time series of the SST anomalies of the first SVD mode. These are shown in Fig 9a and 9b. The anomalous ocean heat transport is defined positive if it warms the SST whereas the anomalous surface heat flux is positive if it cools the SST. The broad scale features of the surface heat fluxes perfectly match the SST anomalies in the sense of the atmosphere forcing the ocean. Ocean heat transport is important in the northern North Atlantic where it is the same magnitude as the surface fluxes. We broke the anomalous heat transport into two terms, advection of the mean SSTs by the anomalous currents and advection of the anomalous SSTs by the mean currents. Advection of the mean temperature by the anomalous currents was the most important term outside the tropics. During a high index NAO year, stronger westerlies drive a southward Ekman drift over the high-latitude ocean that cools the SSTs. This amplifies the cooling due to enhanced surface fluxes. Luksch (1996) noted the same effect in her ocean model simulations of the 1950 to 1979 period. This anomalous Ekman drift increases the simulated SST anomalies relative to the case with a variable depth mixed layer. Clearly, the reasonably sized SST anomalies simulated with a uniform mixed layer depth, and no anomalous ocean heat transport, were obtained for the wrong reason. In reality, both ocean heat transport and surface fluxes are important in this area, and it is necessary to account for the fact that the surface fluxes impact the temperature of a deep wintertime mixed layer. We also looked at the role of anomalous entrainment and found that it damps the SST changes because of the altered difference in temperatures between the mixed layer and below. This confirms the same result seen by Halliwell (1998) in a model integration.

Further south, stronger trades drive a northward Ekman drift that warms the subtropics and southern mid-latitudes. This weakly opposes the cooling of the subtropics by surface fluxes but enhances the warming to the south of the North Atlantic Drift. In the latter region this causes a warming in the GCM which is realistic, but was missed by the ocean mixed layer models.

The strong anomalies seen in the data in the region of the North Atlantic Current region are not reproduced by the mixed layer model or the GCM. The GCM's Gulf Stream and North Atlantic Current are located too far north and lead to increased SST variability in the region east of Canada. It is probable that ocean heat transport is responsible for some of the observed variability of SST in the North Atlantic Current region but this low resolution GCM cannot capture this. The GCM also has an improved SST simulation in the Gulf of Guinea which may indicate a role for equatorial dynamics.

To further examine the role of ocean heat transport we looked at the heat budget of the ocean mixed layer averaged over different areas. An area average in the subpolar gyre shows a strong relationship between SST changes, surface fluxes, ocean heat transport and the wind forcing. Increased westerlies cause dynamical cooling of the ocean that is in phase with the cooling by surface fluxes. This is further evidence for our claim that changes in ocean heat transport are primarily associated with anomalous Ekman drifts that establish themselves instantaneously once the wind changes. We also computed the time series of northward heat flux by the Gulf Stream off Cape Hatteras. This showed no decadal variability or trend in contrast with the coupled model runs of Grötzner et al. (1998) where heat transport in this area precedes the development of SST anomalies in the subpolar gyre. We were unable to find any evidence for any lead or lag relationship involving ocean heat transports, confirming the earlier model result of Luksch (1996). While this is not a comprehensive examination of the possible roles for ocean heat transport it is in contrast with model simulations of the tropical Pacific Ocean where it easy to identify changes in ocean heat transport leading the development of SST anomalies (e.g. Seager 1989).

As a final assessment of the model's ability to reproduce the dominant mode of observed climate variability we show in Fig 9 the modeled SST anomalies for January-February-March averages of 1969 and 1989. These can be compared with the observed SST anomalies for that period shown in Figure 2. The NAO associated SST patterns of these individual winters are broadly the same as those derived by SVD or EOF analysis but show some interesting differences. For example, during 1989 cold water did not stretch all the way across the North Atlantic from Newfoundland to the British Isles but, instead, warm waters lay west of Europe. In 1969 the tropical SST anomalies were the same sign north and south of the equator. Neither winter showed a pattern of strong SST anomalies in the North Atlantic Current region. These differences give some idea of how individual winters can depart from the more typical patterns derived by SVD analysis. Looking at Figure 9, it is quite clear that, with modest differences in position and amplitude, the GCM accurately reproduces the observed variability of these two winters. The peculiarities of the SST patterns, in comparison with the SVD patterns, are also reproduced by the model.

We also examined the higher modes of observed and modeled variability. The second and third modes together explain less variance of SST than the first mode alone. Both higher modes are high latitude features dominated by anomalous circulations at around $55°N$, that associated with the third mode being located much further east than that associated with the second mode. Both modes are dominated by interannual variability without any noticeable trend. The ocean GCM reasonably reproduces the patterns and time evolution of these modes. We regressed the ocean heat transport and surface fluxes onto the time series of the pattern of modeled SST revealed by the SVD analyses. For the second mode anomalous surface heat fluxes are the dominant forcing for SST variability with changes in ocean heat transport contributing in the North Atlantic Current region at about $40°N$. Patterns of surface fluxes, ocean heat transport and SST are not coherently linked for the third mode which makes us wonder about its realism and we do not consider it further.

6. Conclusions

In this study we first examined why surface heat fluxes have varied over the Atlantic Ocean during the last four decades. We analyzed the different terms in the lowest level thermodynamic energy budget using NCEP reanalyzed data. In agreement with the results of others (e.g. Cayan 1992a, 1992b), we have shown that changes in wind speed cause the changes in surface fluxes over the subtropical North Atlantic but that further north anomalous advection is also important, especially advection of cold and dry air off North America. Changes in wind speed and direction cause changes in surface fluxes that force SST changes. We also found that anomalous subsidence can create changes in surface fluxes that damp SST anomalies. Changes in atmospheric eddy fluxes also primarily damp SST anomalies. Therefore, as far as the SST is concerned, it is changes in the mean atmospheric flow that create the SST anomalies while the eddies damp them.

Next we were able to show that a simple model of the atmospheric mixed layer (AML) that balances surface fluxes, radiation, subsidence, advection and eddy transports, was quite capable of reproducing the observed surface flux variability when forced by observed SSTs. This suggests that it would be possible to simulate the SST variability with an ocean model coupled to the AML model. We used three different ocean models: two in which the ocean heat transports were held fixed at their seasonally varying climatological values, the first with a uniform 75m depth and the second with a mixed layer model that allows the depth to vary and, third, a full ocean GCM in which ocean heat transports varied.

The SST variations simulated by the uniform depth mixed layer model were surprisingly similar to those observed. The model reproduces the familiar tripole banded structure of SST anomalies associated with the NAO and also reproduces the long term trend in that pattern towards the high index state of the NAO (Hurrell 1995). This result makes it clear that, to first order, the variations of Atlantic Ocean SSTs since 1958 can be explained as the response to variations in atmospheric circulation. This is true at all time scales. By comparing this result with the SSTs simulated using a variable depth ocean mixed layer we were able to assess the role of mixing. The deep winter mixed layers of the far North Atlantic greatly restricted the amplitude of SST anomalies forced by surface fluxes and, in fact, they were too small. In the South Atlantic the shallow summer mixed layers increase the SST anomalies.

The full ocean GCM also includes the variable depth ocean mixed layer model and, in addition, allows the ocean heat transport to vary. Changes in ocean heat transport are important in the far North Atlantic. Here, when anomalous westerlies cool the SSTs by surface fluxes, they also create an anomalous equatorward Ekman drift that enhances the cooling. The SST anomalies in this simulation were realistic suggesting that here surface fluxes, mixing of the influence of surface fluxes down to considerable depths and changes in ocean heat transport are all important. Anomalous entrainment at the base of the mixed layer damps SST anomalies. In the region to the south of the North Atlantic Current anomalous easterly winds drive an anomalous poleward Ekman drift which warms the SST and greatly improves the realism of the SST simulation relative to the mixed layer models. We were only able to identify a role for anomalous Ekman drifts. These are generated almost instantaneously and cannot provide any long term memory that could lead to oscillatory behavior (e.g. decadal variability). Analyses of the heat budgets in various regions did not uncover any evidence that ocean heat transports systematically lead or lag the SSTs. Instead, where there was a clear signal in changes in ocean heat transport, (e.g. the far North Atlantic) it was in phase with the SST changes forced by surface fluxes.

The results of an ocean modeling study alone cannot be used to fully explain climate variability in the Atlantic sector. We have demonstrated that changes in the surface fluxes forced by a changed atmospheric circulation and, to a much lesser extent, changes in ocean heat transport, can be successfully invoked to explain the variations of Atlantic SST. However, we cannot explain why the atmospheric circulation changed in the first place. The current results are consistent with the atmosphere forcing the ocean at all timescales, including decadal, but this raises a particularly difficult question: where does the persistence from one winter to another, including the long term trends, come from? Atmospheric timescales appear to be too short to explain such low frequency behavior while they may easily explain persistence during a winter. Assuming that the low frequency behavior is not simply the result of the ocean's ability to integrate the atmosphere's high frequency forcing, then there are several possible explanations for the low frequency behavior which we consider in turn.

1. In nature, the ocean heat transport does in fact play the dominant role and the atmosphere responds constructively such that the surface flux anomalies reinforce the SST anomalies generated by ocean dynamics. In our model, the reasoning would follow, we do not see the importance of the changes in ocean heat transport because, by fortuitous tuning, the surface fluxes account for almost all the SST change. If this scenario was correct then the modeled flux anomalies in our models would be systematically too large. There is no evidence for this. Also, while anomalous Ekman drift does contribute to SST variability in the far North Atlantic, elsewhere the ocean's dynamical role is limited. Further, we do not find that the ocean heat transport significantly leads or lags the SST or surface fluxes. This makes it hard to argue for changes in ocean heat transport driving Atlantic climate variability.

   This conclusion appears at face value to contradict the recent atmosphere modeling results of Rodwell et al. (1999). They forced an atmospheric GCM with observed SSTs and, in an ensemble mean, reproduced much of the observed behavior of the NAO since 1947, though with greatly reduced amplitude. This might be taken to suggest that aspects of the NAO's behavior were forced by the ocean. However, we note that their surface fluxes damp the SST anomalies rather than force the SST anomalies as observed. Bretherton and Battisti (1999, manuscript submitted to *Geophsical Research Letters*) argue that these features are the expected result of taking the mean of an ensemble of experiments in which an atmospheric GCM is forced by the time history of SSTs that were, in fact, created by atmospheric forcing. The results, therefore, in their interpretation, do not indicate that the NAO behavior was in any way forced by the changes in SST.

2. In nature, changes in atmospheric circulation create SST anomalies but the atmospheric response to those SST anomalies is such as to reinforce the changes in circulation and fluxes that created the SST anomalies in the first place. Persistence from one winter to

another would be aided if variations in ocean mixing could sequester thermal anomalies below the summer mixed layer to be re-entrained the following winter (Battisti et al. 1995, Bhatt et al. 1998). Both this explanation and the previous one flounder in that they rely on a coherent atmospheric response to mid-latitude SST anomalies, in the sense of high pressure downstream of warm water, that has been difficult to demonstrate.

3. Another explanation of North Atlantic variability is that it is driven from elsewhere, perhaps from the South Atlantic (Robertson et al. 1999). In this scenario changes in South Atlantic SSTs would influence the strength and location of convection over the Amazon and in the ITCZ. Atmospheric teleconnections, or changes in the Hadley Cell, would then communicate this change to the North Atlantic circulation. But why do the South Atlantic SSTs change? This explanation substitutes the problem of explaining the persistence of South Atlantic SSTs for the problem of explaining persistence in the North Atlantic. However it is easier to demonstrate a constructive response of the tropical atmosphere to SST anomalies (e.g. Chang et al. 1999a) so this idea is not entirely implausible.

4. The causes of low frequency variability in the Atlantic sector lie outside of the Atlantic basin. In particular the Pacific Ocean has strong decadal variability whose origin is unknown (Zhang et al. 1997) and, perhaps, this also impacts the Atlantic Ocean via teleconnections. However, the NAO and Pacific variability have not been demonstrated to be well correlated. More generally, the influence of the Pacific on the high and mid-latitude Atlantic, which occurs via the Pacific-North American teleconnection pattern, is weak. On the other hand ENSO has a powerful and coherent impact on the tropical Atlantic (Giannini et al. 1999), suggesting that a combination of this mechanism with the previous, South Atlantic explanation, is a contender for explaining decadal variability in the Atlantic sector.

5. The final contender is greenhouse warming. Shindell et al. (1999) have shown that

rising greenhouse gases in an atmospheric GCM can create a trend in the Arctic Oscillation, which is closely related to the NAO, as vertical wave propagation between the stratosphere and troposphere is altered by a strengthening stratospheric polar vortex. In this scenario surface winds over the Atlantic will be altered as the Arctic Oscillation shifts to a high index phase. This will then cause the SSTs to vary. Our modeling results are entirely consistent with this explanation but, obviously, cannot prove that it is correct.

Our ocean modeling experiments indicate that over the last four decades Atlantic Ocean climate variability can be adequately explained in terms of the ocean being forced by changes in atmospheric circulation. Progress therefore requires understanding why the atmospheric circulation changed. We need to discover what can excite trends in the circulation and what can cause persistence from one winter to another. Changes in the distribution of atmospheric convection in the tropical Atlantic sector are one possibility, greenhouse warming is another and there are probably others. In terms of the persistence within a winter our observational analysis of the thermodynamic budget of the lower part of the atmosphere is revealing. Clearly the mean flow creates SST anomalies that the atmospheric eddies damp. This may not be a fortuitous arrangement. It is possible that the atmospheric eddies force changes in the mean flow via changes in eddy momentum fluxes. These changes in the mean flow create surface flux and SST changes that the eddy heat fluxes then try to damp. This three way coupling between the eddies, the mean flow and the SST may arrange itself in such a way as to allow persistence and will generally redden the spectrum of variability. This will be the topic of future work.

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Figure Captions

Figure 1. The relationship of various terms in the moist static energy budget of the lowest level of the atmosphere to the underlying SST variability. The first EOF of NCEP observed SST is shown in color. Values represent one standard deviation of the corresponding time series. The regressions of the energy budget terms onto the time series of the EOF of SST are contoured. Shown are anomalies of (a) advection, (b) subsidence, (c) the horizontal eddy flux convergence, (d) the vertical eddy flux convergence, (e) the surface flux given by anomalous winds working on the mean vertical gradient of moist static energy and (f) the surface flux given by the mean wind working on the anomalous vertical gradient of moist static energy. The energy budget terms are in Wm^{-2}. Positive contours are solid, negative dashed, the zero contour is in bold and the contour interval is $2Wm^{-2}$

Figure 2. NCEP reanalyzed anomalies for January-February-March seasonal means of 1969 and 1989 of (a) and (b) SST, (c) and (d) surface wind and (e) and (f) the latent plus sensible heat flux.

Figure 3. The latent plus sensible heat flux anomalies in Wm^{-2} simulated by the AML model when forced by NCEP reanalyzed SSTs and winds for January-February-March seasonal means for (a) 1969 and (b) 1989.

Figure 4. Same as for Figure 2 but for the cases where (a) and (b) the wind vectors are held fixed and only wind speed varies and (c) and (d) the wind vectors change but the wind speed is held fixed.

Figure 5. The correlation coefficient between observed and modeled SST anomalies as computed with a 75m deep ocean mixed layer forced by surface flux anomalies computed using standard bulk formulae and observed surface air temperature and humidity. Ocean heat transports do not vary except for an imposed seasonal cycle. The correlation is good to excellent everywhere, even in the tropical Pacific, indicating that using the observed thermodynamic properties of the atmospheric boundary layer ensures a good SST simulation but frequently for the wrong reason.

Figure 6. (a) Results of an SVD analysis between NCEP wind vectors and SST anomalies during the January to March season. This first mode explains 23% of the variance in SST and 25% and 20%

of the monthly zonal and meridional winds respectively and is associated with the North Atlantic Oscillation. In (b) we show the same analysis performed with the SST anomalies computed by a 75m deep ocean mixed layer coupled to the atmospheric mixed layer model. The patterns in (b) explain 23%, 27% and 29% of the seasonal mean SST, zonal and meridional wind anomalies, respectively. The time series of the observed and modeled SST modes are shown in (c).

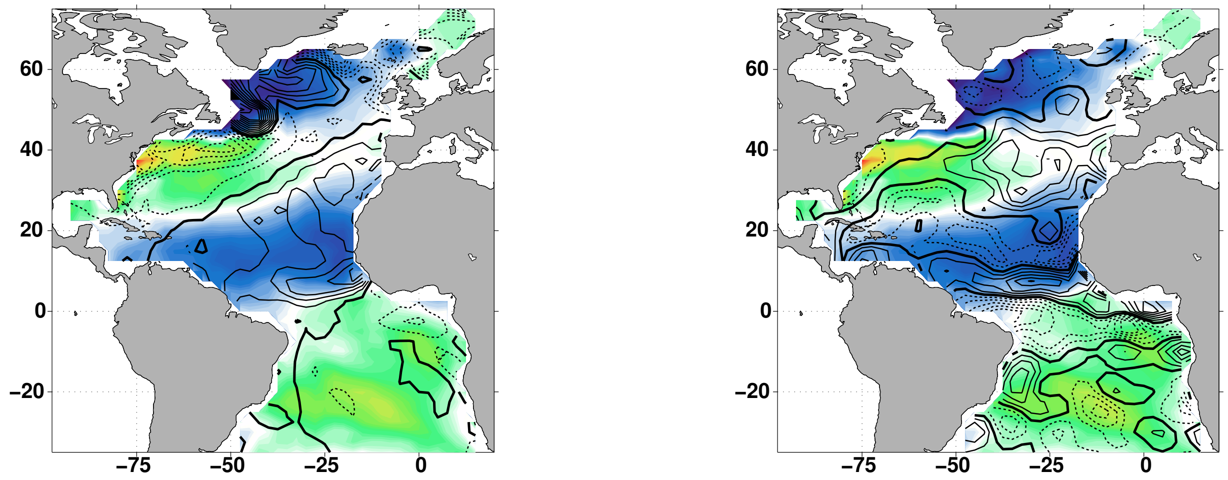
Figure 7. Same as for Figure 6 but for the cases where the SST anomalies were computed with a variable depth ocean mixed layer and (b) with the full dynamical ocean GCM both coupled to the AML model. The time series are shown in (c).

Figure 8. The regression of the anomalies of the modeled surface heat flux (positive if it cools the ocean) and ocean heat transport (positive if it warms the ocean) onto the time series of the first mode of SST variability shown in Figure 7b. The surface and dynamical heat fluxes are contoured in Wm^{-2} and represent the variations associated with a one standard deviation fluctuation in the principal component. Positive contours are solid, negative dashed, the zero contour is in bold and the contour interval is $3Wm^{-2}$

Figure 9. The SST anomalies for January-February-March seasonal means as simulated by the ocean GCM for (a) 1969 and (b) 1989. These can be compared to the corresponding figures for the observed anomalies in Figure 3.

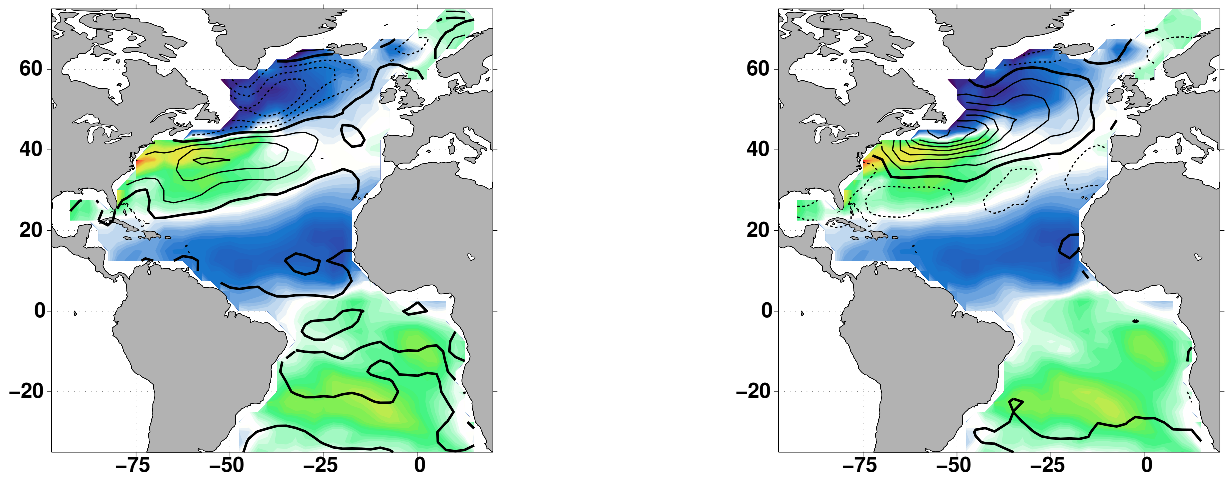
(a) Advection

(b) Subsidence



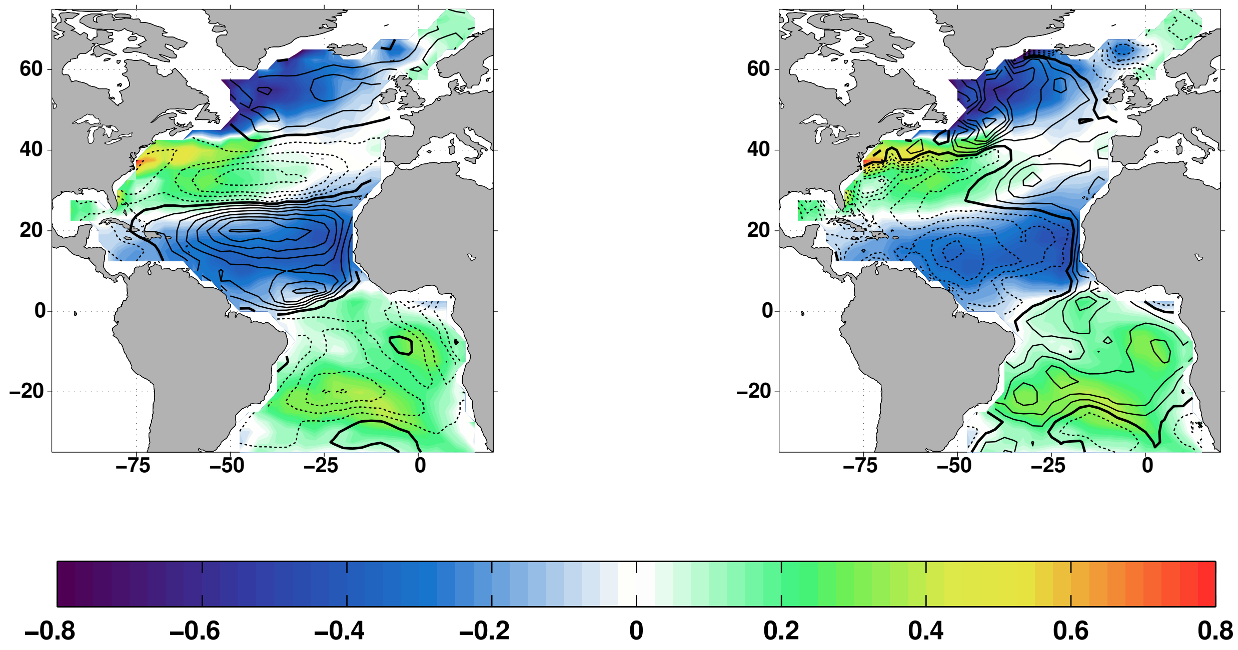
(c) Horizontal Eddy Flux

(d) Vertical Eddy Flux



(e) Wind Induced Surface Flux

(f) Gradient Induced Surface Flux



1969 **1989**

Sea Surface Temperature

(a) (b)

Surface Winds

(c) (d)

Latent and Sensible Heat Fluxes

(e) (f)

1969 **1989**

AML Latent and Sensible Heat Fluxes

(a) (b)

1969 **1989**

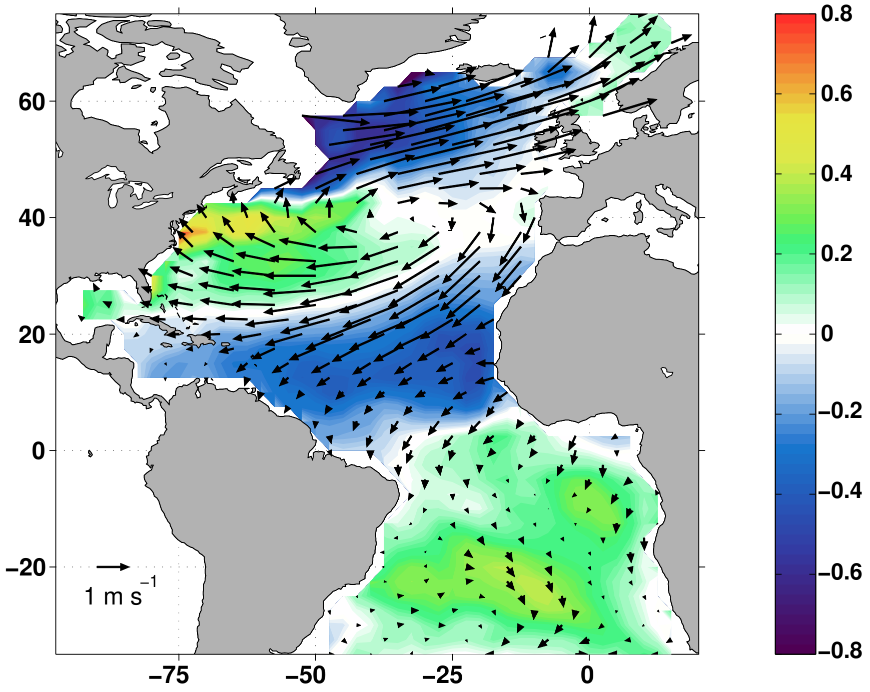
Wind Vectors Fixed

(a) **(b)**

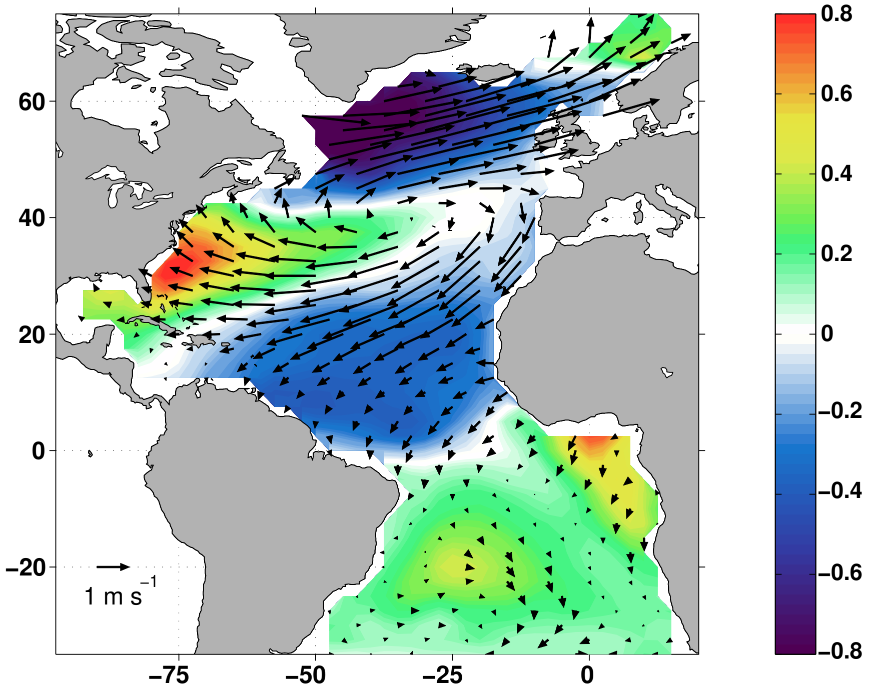
Wind Speed Fixed

(c) **(d)**

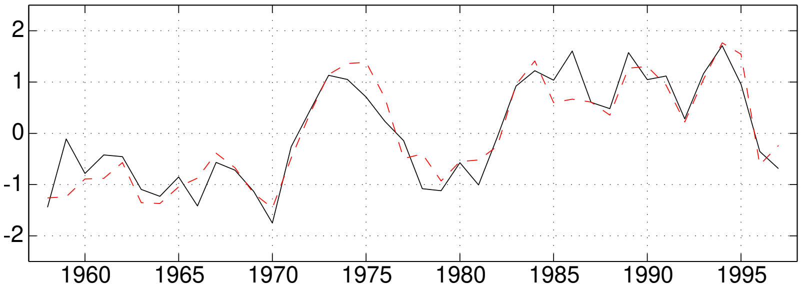
(a) Observed First Mode of Variability



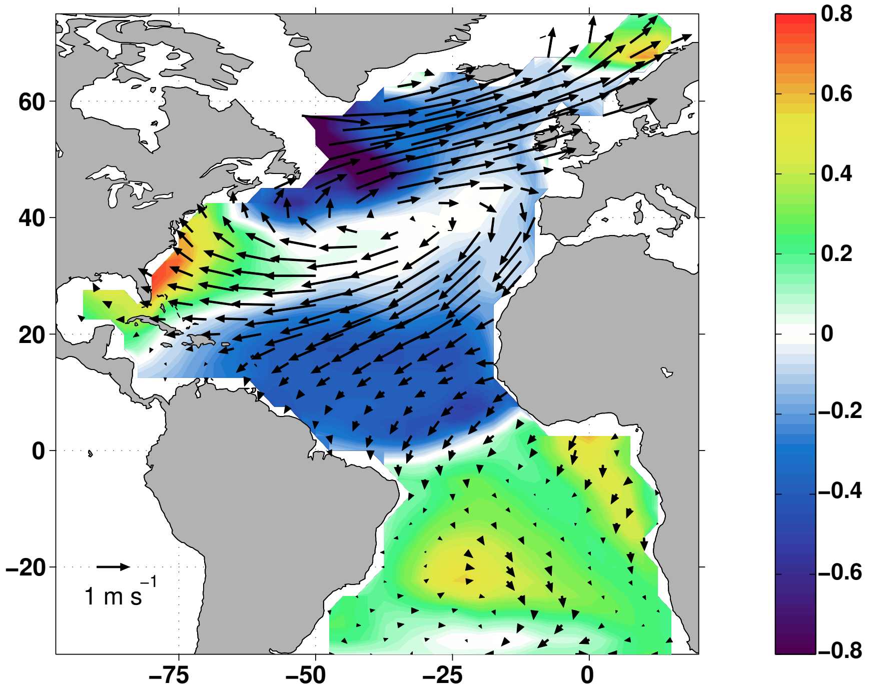
(b) Uniform Depth OML First Mode of Variability



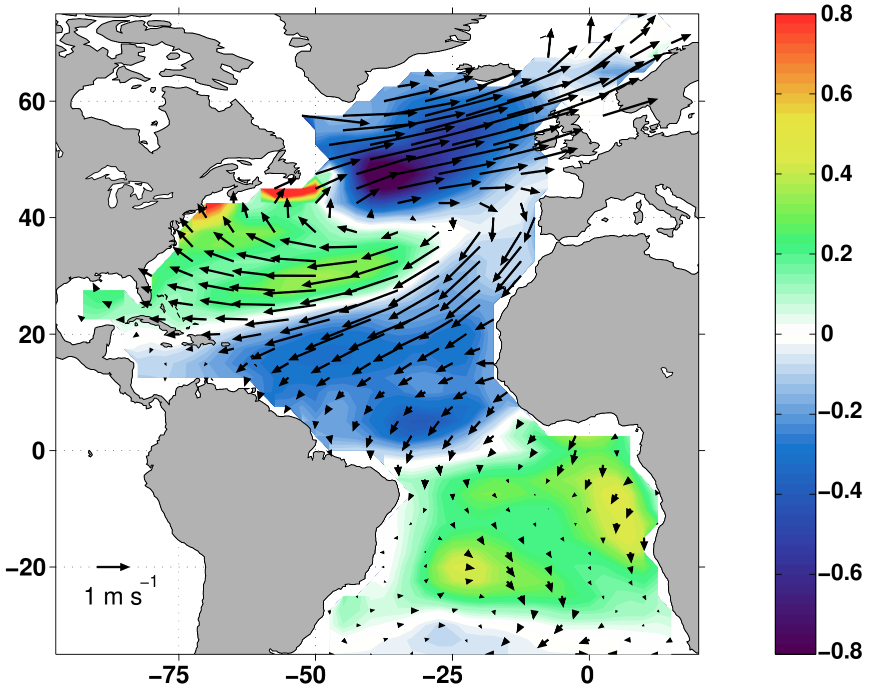
(c) Observed Time Series (Black), Uniform Depth OML (Red)



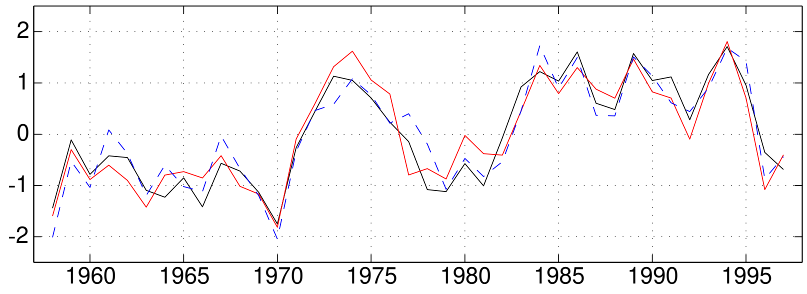
(a) Variable Depth OML First Mode of Variability



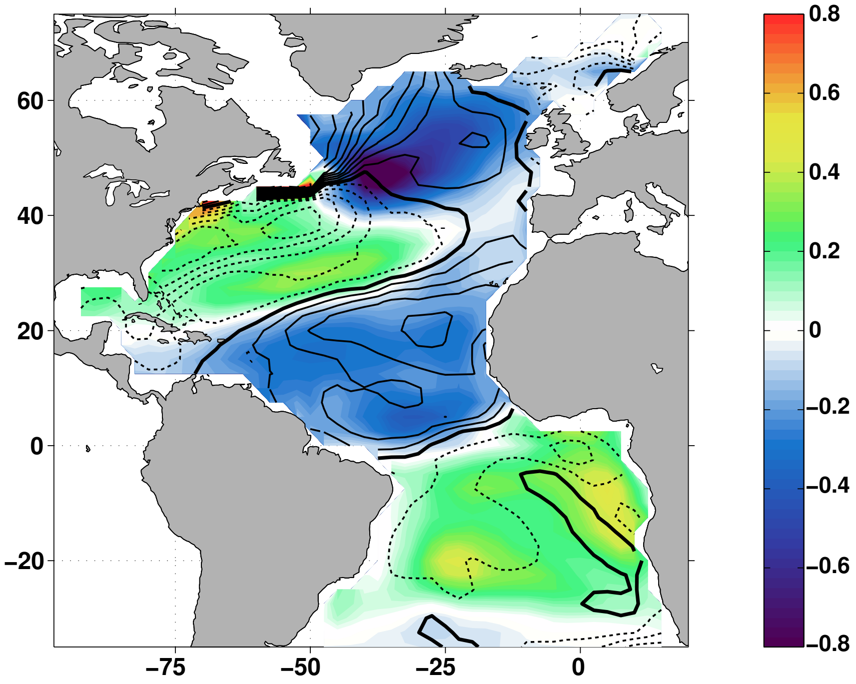
(b) OGCM First Mode of Variability



(c) Observed Time Series (Black), Variable Depth OML (Red), OGCM (Blue)

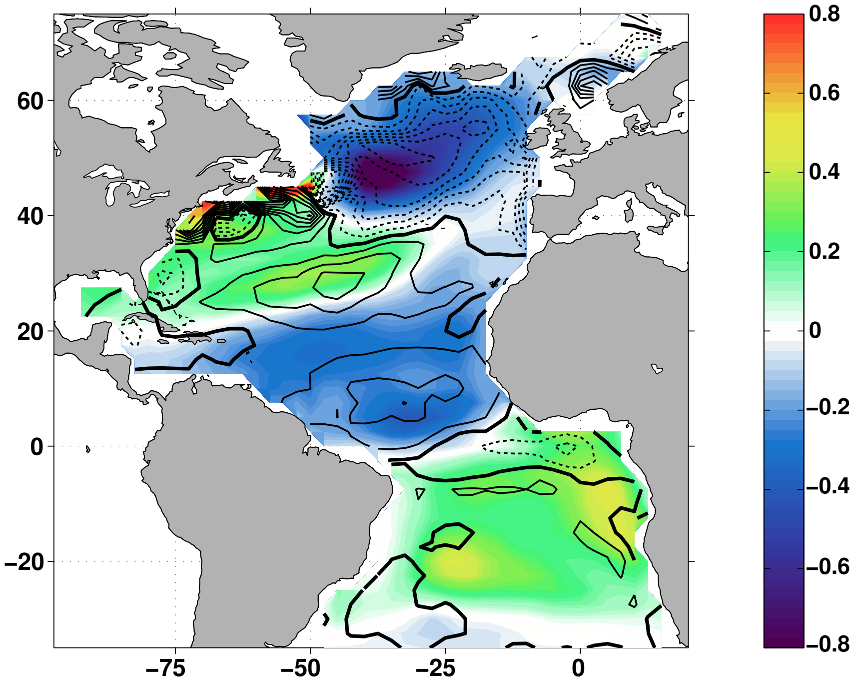


(a)



OGCM Surface Fluxes

(b)



OGCM Ocean Transport

1969 **1989**

OGCM Sea Surface Temperature

(a) (b)