The Southern Ocean: A Possible Source Region for Tropical Atlantic Variability?

by

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We accept this thesis as conforming to the required standard

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Abstract

A coupled model of intermediate complexity is used to examine the effect of Southern Ocean salinity, temperature and wind anomalies on tropical Atlantic variability. We find that positive freshwater, temperature and wind anomalies in the Southern Ocean all cause increased advection into the Atlantic basin, resulting in anomalously cold and fresh water in the eastern Atlantic Ocean at 25°S. Over time, this anomaly spreads northwards and away from the coast, into the tropics. The magnitude of the response increases linearly with the strength of the forcing. The results of these experiments show that southeastern Atlantic Ocean temperature and salinity are particularly sensitive to changes in the Southern Ocean. Limited observational data support the model results. Our findings suggest that this link could be the oceanic branch of a mode of variability linking the Southern and tropical Atlantic Oceans—a possible mechanism for the as-yet-unexplained decadal mode of tropical Atlantic variability.

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Table of Contents

List of Figures

Chapter 1

Introduction and Study Motivation

The Atlantic Ocean has been more widely studied than the Pacific, but its variability is less well understood. There are several reasons for this apparent contradiction. Sea surface temperature (SST) variability in the Atlantic is weaker than in the Pacific (Dommenget and Latif 2000), so patterns are harder to detect. More importantly, while Pacific Ocean variability is dominated by a single mode—the El Niño-Southern Oscillation (ENSO)—the Atlantic appears to contain several competing modes of similar amplitude, which may be interacting with one another.

We are motivated to learn more about the tropical Atlantic system not only to increase our knowledge of this area, but because Atlantic variability is linked to important—and sometimes devastating—climatic processes. The north-south gradient in tropical SST has been closely linked with the well-known droughts in northeastern Brazil (see Figure 1.1, Hastenrath (1976), Moura and Shukla (1981), Hastenrath (1990)), and rainfall variability in the Sahel region of Africa (Lamb 1978, Lamb and Peppler 1992), through its effect on the southern extent of the inter-tropical convergence zone (ITCZ). Atlantic SST is already being used for long-term prediction of rainfall in Brazil and in the Sahel region of Africa (Lamb et al. 1986, Servain 1991, Hastenrath and Greischar 1993). So far, however, these predictions can only be based on the persistence of existing anomalies. Better forecasting of drought or flood years, through an understanding of the mechanisms that produce SST anomalies, has the potential for immense human benefit because it would allow people to plan for such events and to mitigate their effects.

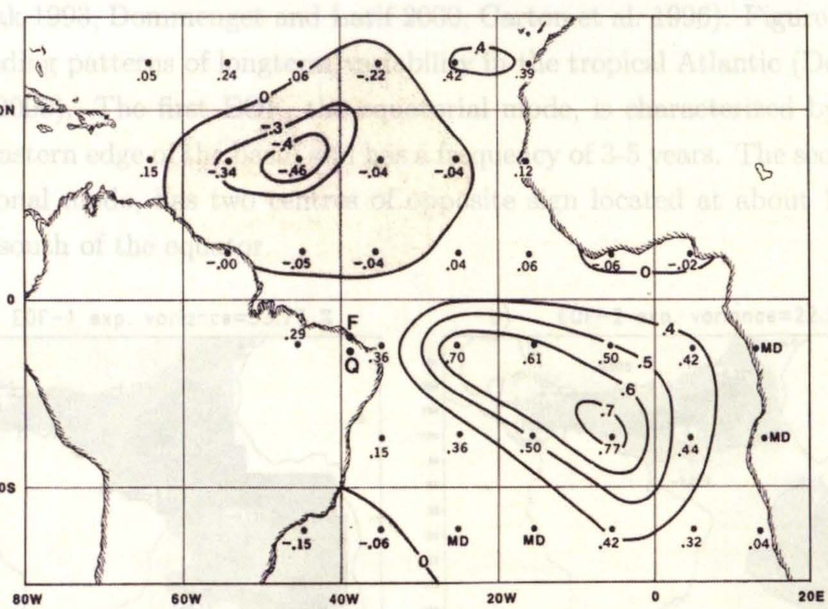

Figure 1.1: **Correlation between monthly SST anomalies in the Atlantic and rainfall at two stations in Northeast Brazil (F and Q). From Moura and Shukla (1981).**

1.1 Observed Patterns of Tropical Atlantic Variability

As a first step toward understanding the system, past variability can be used to establish modes or patterns of variability. Detecting patterns in SST observations is not as trivial an undertaking as it might seem, however. Older records were taken along shipping routes and don't necessarily provide good coverage of the area. Also, the historical measurements have been taken using different methods, of varying reliability. The trick is to match different data records together and to interpolate to cover gaps in the data. There are two popular long-term datasets in use—the Comprehensive Ocean-Atmosphere Data Set (COADS), a 100-year set compiled by daSilva et al. (1994), and the Global Ocean Surface Temperature Atlas (GOSTA) database, a 110-year set of shipboard measurements compiled by Bottomley et al. (1990).

Observed variations in Atlantic SST are generally 0.3 to 0.5 degrees Celsius (Carton et al. 1996). Studies consistently show *interannual* variability on or near the equator and *decadal* variability off the equator (Mehta 1998, Mehta and Delworth

1995, Zebiak 1993, Dommenget and Latif 2000, Carton et al. 1996). Figure 1.2 shows the two leading patterns of longterm variability in the tropical Atlantic (Dommenget and Latif 2000). The first EOF, the equatorial mode, is characterized by variance along the eastern edge of the basin and has a frequency of 3-5 years. The second EOF, the meridional mode, has two centres of opposite sign located at about 15 degrees north and south of the equator.

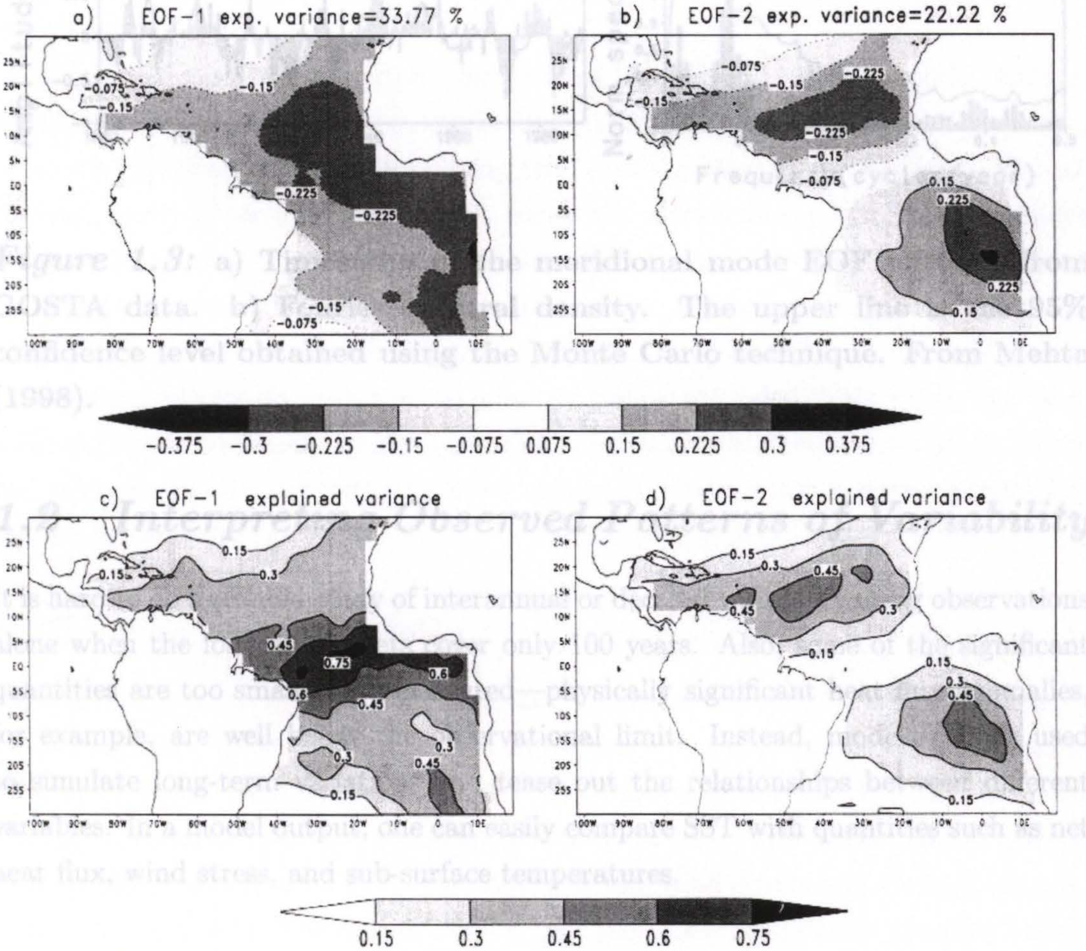

Figure 1.2: **Leading EOFs of annually-averaged SST, from GISST data. Units are degrees. a) 1st EOF, equatorial mode, represents about 34% of the total variance. b) 2nd EOF, meridional mode, represents 22% of the total variance. Panels (c) and (d) are the associated spatial distributions of explained variance. From Dommenget and Latif (2000).**

Figure 1.3 shows a timeseries for the same meridional mode, obtained in a separate study (Mehta 1998). The meridional pattern exhibits an oscillation period of 13 years that is above the 95% confidence limit. Notice that the centres of the meridional mode in Figure 1.2 b) correspond to the regions where SST is correlated with Brazilian rainfall, shown in Figure 1.1.

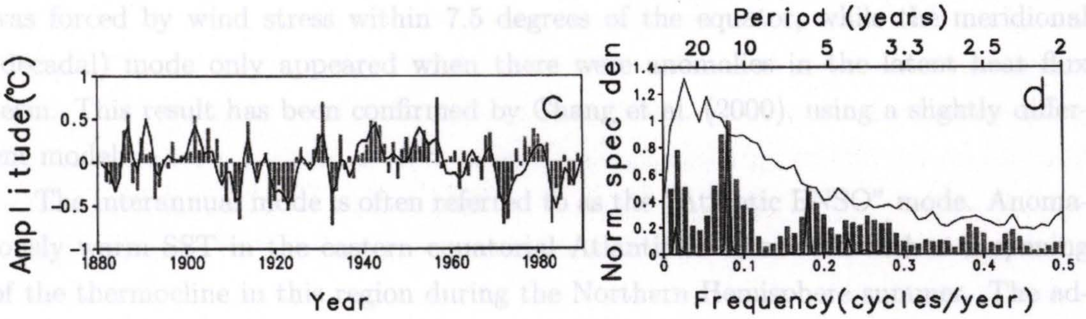

Figure 1.3: **a) Timeseries of the meridional mode EOF pattern, from GOSTA data. b) Fourier spectral density. The upper line is the 95% confidence level obtained using the Monte Carlo technique. From Mehta (1998).**

1.2 Interpreting Observed Patterns of Variability

It is hard to do a reliable study of interannual or decadal variability using observations alone when the longest datasets cover only 100 years. Also, some of the significant quantities are too small to be measured—physically significant heat flux anomalies, for example, are well below the observational limit. Instead, models can be used to simulate long-term variations and tease out the relationships between different variables. In a model output, one can easily compare SST with quantities such as net heat flux, wind stress, and sub-surface temperatures.

1.2.1 The Interannual Mode

The interannual mode is a good example of a phenomenon that appears to have been successfully explained through a combination of observational analysis and computer modelling studies. As an example of one of these studies, Carton et al. (1996) developed a model of the Atlantic where they parameterized the important dynamic interactions and heat fluxes at the sea surface. They succeeded in reproducing the observed anomalies in SST when they forced the model with observed winds. Carton et al. (1996) then separated the model heat flux term at the surface into four components: net shortwave radiation, net longwave radiation, sensible heat loss and latent heat loss. To determine the effect of different components, they performed experiments where they changed only one parameter to determine whether the anomalous SST pattern was still produced. They found that the equatorial (interannual) mode was forced by wind stress within 7.5 degrees of the equator, while the meridional (decadal) mode only appeared when there were anomalies in the latent heat flux term. This result has been confirmed by Chang et al. (2000), using a slightly different model.

The interannual mode is often referred to as the "Atlantic ENSO" mode. Anomalously warm SST in the eastern equatorial Atlantic is associated with a deepening of the thermocline in this region during the Northern Hemisphere summer. The additional warm water is supplied from either the northwestern or the southwestern Atlantic, in response to a relaxation of the trade winds (Carton et al. 1996). Chang et al. (2000) identified an ENSO-like atmospheric response. Ruiz-Barradas et al. (2000) showed that this mode is forced by wind stress, like ENSO—thermodynamic air-sea interactions do not seem to play an important role. This mode has been explored most thoroughly by Zebiak (1993). He found that the Atlantic mode is similar to its Pacific counterpart, but more tightly focused on the equator and situated further to the west.

1.2.2 The Decadal Mode

The dominant mode on decadal time scales, shown in Figure 1.2 b), is more difficult to characterize. It has been identified as a cross-equatorial dipole with a period of around 13 years. Dommenget and Latif (2000) found that these SST anomalies are strongly correlated with wind stress anomalies in the trade wind zones. Ruiz-Barradas et al. (2000), looking at SST, oceanic heat content, wind stress, and atmospheric diabatic heating in a 36-year dataset, found that this "interhemispheric gradient" mode seemed to be controlled mainly by the thermodynamic air-sea interactions, although equatorial dynamics might also play a role. Some modelling studies show that observed wind anomalies can produce the anomalous SST pattern through variations in the latent heat flux term (Carton et al. 1996, Huang and Shukla 1997, Chang et al. 2000), but others demonstrate that SST anomalies can excite the observed atmospheric variations (Sutton et al. 2000, Okumura et al. 2001). It appears that there is a positive feedback between wind stress and SST anomalies, probably through a wind-evaporation-SST feedback.

Chang et al. (1997) put forward the only real theory for explaining the decadal oscillation. They proposed a coupled air-sea interaction in which latent heat flux anomalies act as a positive feedback on SST anomalies in each hemisphere and cross-equatorial advection by mean currents acts as a negative feedback. A number of subsequent studies have found no evidence for this mechanism (Mehta 1998, Sutton et al. 2000), and Chang's own subsequent work has failed to provide strong support for his theory (Chang et al. 2000, Seager et al. 2001). So the questions surrounding decadal variability are still open: which is leading, the wind or SST anomalies, and what is setting the decadal timescale?

One obstacle to fully explaining the dipole pattern is... that it is not actually a dipole! Although many authors first described the cross-equatorial pattern as a "dipole mode" (Chang et al. 1997, Xie and Tanimoto 1998, Yang 1999, Ruiz-Barradas et al. 2000, Okumura et al. 2001), recent studies suggest that there are independent patterns of variability on either side of the equator that only sometimes give the appearance of a dipole. Enfield and Mayer (1997) analysed a 43-year section of COADS data and determined that the regions north and south of the ITCZ are indeed uncorrelated. Dommenget and Latif (2000) used VARIMAX rotated EOF analysis on GISST data and models to confirm that the patterns were uncorrelated. Mehta and Delworth (1995) had originally found evidence of a dipole pattern in their detailed analysis of GOSTA data; however, Mehta (1998) later redid the analysis and revised his earlier conclusions. He found a timescale of 12-13 years (significant to the 95% level) for the variability in the South Atlantic, but no distinct timescale in the North Atlantic. Approximately 80% of the decadal variance in the cross-equatorial SST gradient was explained purely by variation in the South Atlantic (Mehta 1998). Mehta concluded that the apparent dipole was an artifact of the EOF analysis.

Mehta (1998) also found that centres of the anomalies in the North Atlantic appeared to propagate in a clockwise rotation around the basin. Anomalies in the South Atlantic basin propagated in a counterclockwise sense. Hansen and Bezdek (1996) confirmed this rotation. Delworth and Mehta (1998) found that the clockwise propagation of anomalies in the northern hemisphere was more prominent below the surface, but had no explanation for this phenomenon.

1.3 Tropical—Extratropical Connections

The theory of tropical–extratropical connections maintains that when trying to explain variability in the tropics, one should look for a source far away, in the mid to high latitudes. Although somewhat counter-intuitive, this approach makes sense because of the shape of isopycnals, or density levels. Interannual variability is often associated with water properties below the mixed layer, where anomalies can persist from year to year. Properties below the mixed layer are generally affected by ocean–air interactions in the formation regions, the isopycnal outcropping regions at high latitudes.

Gu and Philander (1997) proposed that parcels of water with anomalous properties may subduct at midlatitudes in the Pacific Ocean, travel along isopycnals toward the equator, and then affect equatorial thermocline properties when they arrive in the tropics, years later. Gu and Philander suggested that this could lead to a self-sustaining oscillation with a period set by the travel time of the subducted parcel. This mechanism has been shown to be important in decadal modulation of the ENSO cycle in the Pacific Ocean (Weaver 1999, Cai and Whetton 2001). Alternatively, it has been proposed that low-frequency tropical–extratropical connections may also occur, not as an advection of temperature anomalies by mean currents (vT'), but through variations in the strength of the meridional overturning ($v'T$) (Kleeman et al. 1999, McPhaden and Zhang 2002). Both mechanisms will be considered in this study.

1.4 Extratropical Subduction in the South Atlantic?

Though the theory of tropical—extratropical connections was developed to explain Pacific variability, we have some reason to think that it might also apply to the Atlantic Ocean. Previous research on tropical Atlantic variability has suggested the possibility of an extratropical source. Häkkinen and Mo (2002) wrote that their model experiments "suggest that the southern anomalies have to arise from other processes [not surface heat flux anomalies] such as wind-driven circulation changes." Mehta (1998) noted that his analysis of observational data "suggested possible connections between tropical and extratropical SST variations in each hemisphere." He found a low squared-coherence (0.1-0.4) between the tropical and midlatitude North Atlantic SST variations at decadal timescales. The squared-coherence increased to 0.6 at high latitudes with a lag of a few years [1]. Mehta (1998) wrote that "the results of the present study suggest an alternate paradigm [to Chang's] of the tropical Atlantic SST

[1] The squared coherence is the frequency-domain equivalent of the correlation coefficient. Instead of being a correlation of two timeseries, it is essentially a correlation coefficient of the power spectra of two time series, which shows the degree of correlation as a function of frequency. The squared coherence thus measures the extent to which one series (in this case, one hemispheric pattern) can be considered as a linear function of the other

variations in which tropics-extratropics interactions via the ocean and the atmosphere may be very important." Xie and Tanimoto (1998) have also suggested that the decadal timescale may be set in the extratropics.

There is observational evidence that subduction in the *northern* high latitudes plays a role in the tropical Atlantic. Curry et al. (1998) have shown that the thickness of the subducting Labrador Seawater (LSW), a proxy for the strength of the meridional overturning, is linked to mid-depth water temperature in Bermuda, at 32°N. Figure 1.4 shows the negative correlation between subtropical temperatures and subpolar thicknesses with a lag of 6 years (Curry et al. 1998). Yang (1999) has also shown that the tropical Atlantic SST gradient is correlated to the thickness of the LSW, with a correlation of 0.68 (significant at the 5% level) when the LSW leads by 5 years.

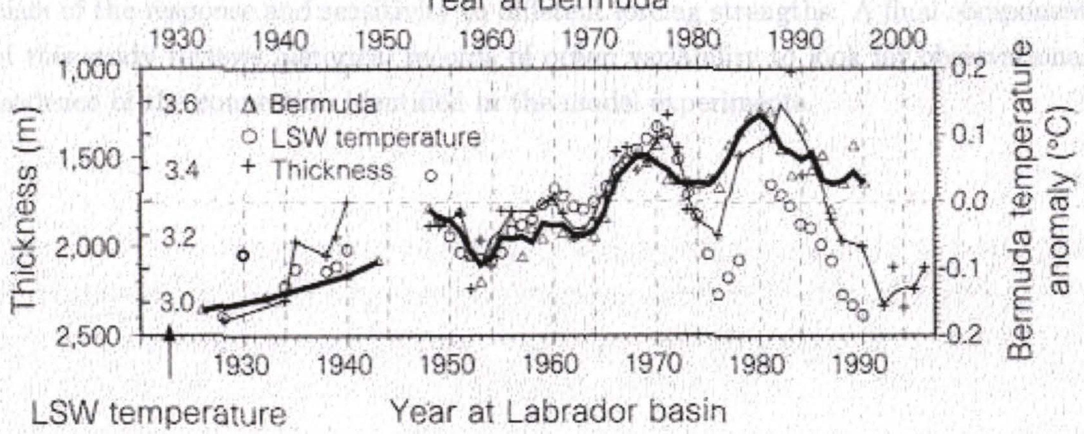

Figure 1.4: **Bermuda temperature anomaly (triangles, thick curve) lagged by 6 years with Labrador Seawater (LSW) source thickness (thin curve and + symbols, note inverted axis) and LSW core temperature (circles). A negative correlation (r<-0.6) is found between LSW thickness and Bermuda temperature anomalies. From Curry et al. (1998).**

There is every reason to believe that this mechanism should be at least as important in the south Atlantic Ocean. In a model experiment where Weaver (1999) showed that northern high-latitude subduction in the Pacific Ocean can influence El Niño on decadal–interdecadal timescales, he commented that an additional experiment, in which perturbations were applied only in the southern hemisphere, gave equally valid results with an even stronger response. If a southern-hemisphere source is possible in the Pacific, it seems even more likely in the Atlantic: Harper (2000) and Jochum and Malanotte-Rizzoli (2001) have shown that the tropical Atlantic thermocline is ventilated almost exclusively by water from the southern high latitudes.

Finally, the southern centre of tropical Atlantic variability is of practical interest. In the study of GOSTA SST observations by Mehta (1998) described in Section 1.2, where he treated each hemisphere separately, showed a timescale of 12-13 years for the variability in the south Atlantic but revealed no distinct timescale for the northern centre (further evidence that coupling between hemispheres may not be important). Mehta (1998) also found that approximately 80% of the decadal variance in the cross-equatorial SST gradient was explained by variations in the southern centre.

In this study, we use a coupled ocean–atmosphere model of intermediate complexity to look at the possible role of the Southern Ocean in tropical Atlantic variability. In model experiments, we find a consistent response to Southern Ocean freshwater, heat, and wind forcing. Further experiments and analysis characterize the mechanism of the response and sensitivity to different forcing strengths. A final component of this study reviews historical records of ocean variability to look for observational evidence of the connection identified in the model experiments.

Chapter 2

Methodology

2.1 The UVic Coupled Model

In these experiments we use version 2.4 of the UVic Earth System Climate Model (ESCM) (Weaver et al. 2001). The UVic ESCM couples a full ocean circulation model and sophisticated sea ice model to a simple atmosphere. It is a useful tool for simulating climate on longer timescales, where high-frequency atmospheric effects are less important than ocean circulation changes.

The ocean component of the model is the Geophysical Fluid Dynamics Laboratory Modular Ocean Model (version 2.2), a fully nonlinear 3-D ocean GCM with 19 vertical levels (Pacanowski 1995). We include the Gent and McWilliams (1990) parameterization for mixing associated with mesoscale eddies, which has been shown to be particularly important for parameterizing processes in the southern high latitudes (Gent et al. 1995, Hirst et al. 2000). The horizontal and vertical viscosity coefficients are set to $2.0 \times 10^9 cm^2 s^{-1}$ and $10 cm^2 s^{-1}$, respectively. We use the Bryan and Lewis (1979) scheme for depth-dependent vertical diffusivity, with values ranging from $k_v = 0.6 \times 10^{-4} m^2 s^{-1}$ at the surface to $k_v = 1.6 \times 10^{-4} m^2 s^{-1}$ at depth.

A reduced complexity, 2-D energy-moisture balance atmospheric model, based on Fanning and Weaver (1996), is used for computational efficiency. Atmospheric sensible heat transport is parametrized as a diffusive process, and moisture transport as a combination of diffusion and advection by winds. Precipitation occurs when the relative humidity in the atmosphere reaches a threshold value of 85%. Precipitation over land is instantaneously returned to the ocean via one of 33 river drainage basins, unless it falls as snow, in which case it is retained locally until it melts.

The energy-conserving thermodynamic–dynamic sea ice model has a subgridscale ice thickness distribution allowing for two ice categories, plus open water, within each model grid cell (Bitz et al. 2001), and an elastic–viscous–plastic representation of sea ice dynamics (Hunke and Dukowicz 1997). The Duffy and Caldeira (1997) parametrization of brine rejection associated with sea ice formation is included.

The atmosphere and ocean are coupled through the exchange of heat and freshwater. Global horizontal resolution for ocean, atmosphere and ice is 3.6° longitude by 1.8° latitude. No flux adjustments are required for this model to keep the simulation of the present climate stable. The only external forcing on the system is solar radiation at the top of the atmosphere and surface wind stress. The model resolves the annual cycle; solar radiation varies seasonally and with latitude and orbital parameters. Applied winds are from NCEP reanalysis monthly mean climatological data (Kalnay et al. 1996).

The UVic model has been extensively and successfully validated against both contemporary climate observations (Weaver et al. 2001) and paleo proxy records (Schmittner et al. 2002a,b). The model has also been shown to simulate realistic ocean circulation and has been used in sensitivity studies for water mass formation (Saenko and Weaver 2001, Weaver et al. 2003, Saenko et al. 2003). In the following experiments, we exploit the UVic model's ability to realistically simulate water mass formation and large-scale ocean circulation, while allowing enough computational efficiency to perform repeated experiments and sensitivity tests.

2.2 Experimental Design

In order to investigate the possible effect of Southern Ocean temperature and salinity variability on the tropical Atlantic, freshwater and heat forcings were applied upstream of the region of interest in two locations (see Figure 2.1). The first, Region A, is a region of enhanced Antarctic Intermediate Water formation south of the core of the Antarctic Circumpolar Current (ACC) (Saenko et al. 2003). The second, Region B, is a more commonly known site of enhanced Antarctic Intermediate Water formation around the tip of South America (McCartney 1977, England et al. 1993). In general, the circumpolar band of open water at latitudes of Drake Passage prevents mean zonal pressure gradients (above the sill depth) and so prevents a net meridional geostrophic transport. However, there is a local pressure gradient west of the Antarctic Peninsula, and this allows a northward transport in this region. West of the Antarctic Peninsula is therefore a location from where the subpolar anomalies can more easily be transported northward across the ACC, to join up with the northward boundary current off South America [Saenko, personal communication]. Because this represents a dynamically viable connection between the high-variability Southern Ocean and the tropics, it was chosen as forcing Region A and used in most experiments.

The model was spun up from rest for 4500 years under climatological winds, and the final steady state served as the initial condition for the idealized sensitivity experiments. In the first set of experiments, uniform surface heat and freshwater anomalies were applied in either Region A or B for between 1 and 5 years (see Figure 2.1). We then continued to integrate the model without anomalous forcing to evaluate the delayed response.

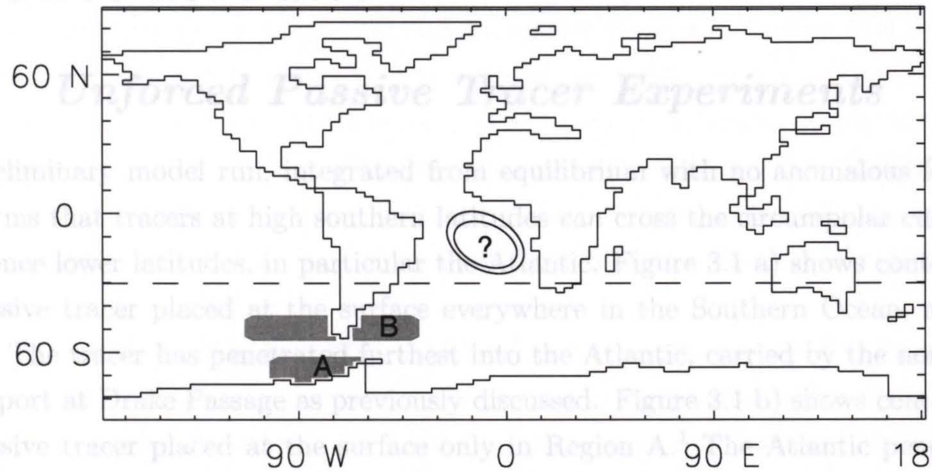

Figure 2.1: **Configuration of the UVic ESCM. Freshwater and heat forcings were applied in Regions A and B. Winds were forced everywhere south of the dashed line. The centre of unexplained decadal variability is marked with a question mark.**

A third set of experiments involved changing the wind stress over the entire Southern Ocean, south of 30°S (see Figure 2.1). The large interannual variations in the characteristics of subantarctic mode water have been shown to depend on the strength of Ekman pumping (Rintoul and England 2002). In our experiment, winds were gradually increased or decreased by an amount proportional to their original strength. The resulting forcing is a smooth 5-year pulse, of sinusoidal shape, with maximum wind stress 1.5 times the original value at each grid point. The relatively large area for the wind forcing, as far north as 30°S, was chosen to avoid unrealistic discontinuities in the wind stress field.

To evaluate the mechanism of the response, a passive "age" tracer was introduced at the surface in the forcing region, in a manner similar to Saenko et al. (2003) and described in England et al. (1993). The passive tracer is advected, diffused and convected in the same way as temperature and salinity in the model, following the movement of water masses without influencing ocean circulation in any way. The tracer values increase at a rate of 1 unit per year everywhere except in the region of tracer release.

Chapter 3

Model Results

3.1 Unforced Passive Tracer Experiments

A preliminary model run, integrated from equilibrium with no anomalous forcings, confirms that tracers at high southern latitudes *can* cross the circumpolar current to influence lower latitudes, in particular the Atlantic. Figure 3.1 a) shows contours for a passive tracer placed at the surface everywhere in the Southern Ocean, south of 60°S. The tracer has penetrated furthest into the Atlantic, carried by the northward transport at Drake Passage as previously discussed. Figure 3.1 b) shows contours for a passive tracer placed at the surface only in Region A.[1] The Atlantic penetration is similar, suggesting that the connection to the Atlantic Ocean comes mainly from Region A. These results confirm earlier work by Saenko et al. (2003), showing that water from Region A can escape northward into the Atlantic basin.

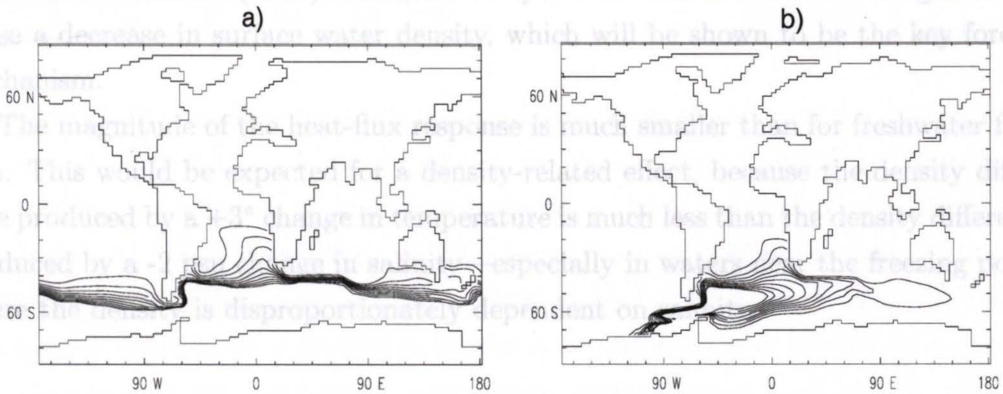

Figure 3.1: **Passive tracer contours at 177 m depth, a) 25 years after placing a passive tracer at the surface in the entire Southern Ocean poleward of 62°S and b) 15 years after placing a passive tracer at the surface only in Region A.**

[1] Tracer values are affected by two quantities: the concentration of tracers from the forcing region, and the tracer's "age" (years elapsed since it left the forcing region). Because tracers are affected by these two separate processes, the values do not have a clear physical meaning. The contours (non-zero values) do, however, show where in the ocean tracers originating in the forcing region have penetrated.

3.2 Freshwater and Heat Forcing Experiments in Region A

Both freshwater and heat forcing in Region A of the Southern Ocean produce a cold and fresh subsurface anomaly in the southeast Atlantic Ocean. Figure 3.2 shows the delayed oceanic response to a 5-year, 0.1 Sv surface freshwater flux applied in Region A, which corresponds to about a -2 psu change in surface salinity. The cold, fresh response in the southeast Atlantic reaches a maximum strength 15 years after the beginning of the perturbation. The response is shown on the sigma = 26.5 isopycnal, where it is strongest. The tracer contours in Figure 3.2 c), which will be discussed in more detail in a later section, show that there is direct influence from the forcing region.

Figure 3.3 shows the result of a 150 W/m^2 anomalous surface heat flux applied for 5 years in Region A, which corresponds to about a 3° increase in surface temperature. As with the freshwater experiments, the response is cool and fresh and limited to the southeast Atlantic. The fact that a *positive* heat flux anomaly leads to a *negative* temperature response in the Atlantic shows that the mechanism of tropical—extra-tropical exchange exhibited here is not a simple advection of anomalies, as proposed by Gu and Philander (1997). Here, the temperature and freshwater forcing similarly cause a decrease in surface water density, which will be shown to be the key forcing mechanism.

The magnitude of the heat-flux response is much smaller than for freshwater forcings. This would be expected for a density-related effect, because the density difference produced by a +3° change in temperature is much less than the density difference produced by a -2 psu change in salinity—especially in waters near the freezing point, where the density is disproportionately dependent on salinity.

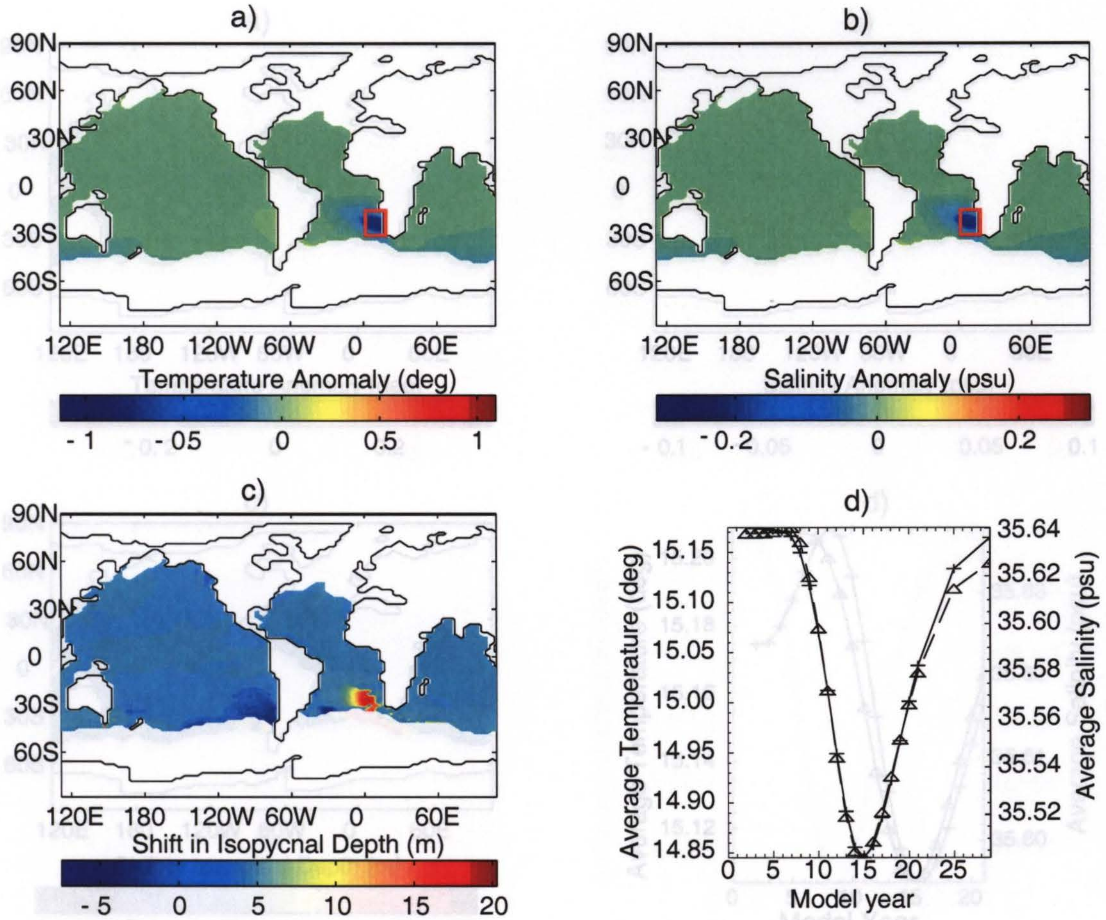

Figure 3.2: **Anomalous a) potential temperature and b) salinity response at year 15, to 5 years of 0.1 Sv freshwater forcing in Region A. Anomalies are shown on the sigma-theta = 26.5 isopycnal. c) Change in the depth of the sigma-theta = 26.5 level from equilibrium conditions. d) Average potential temperatures (crosses) and salinities (triangles) inside the boxed region, as a function of time. Note that the values in panel d) are obtained from averaging over three oceanic grid box levels, ranging from 75 m to 350 m depth. All anomalies represent a one-year average of the experimental run minus a 100-year average of an unforced run.**

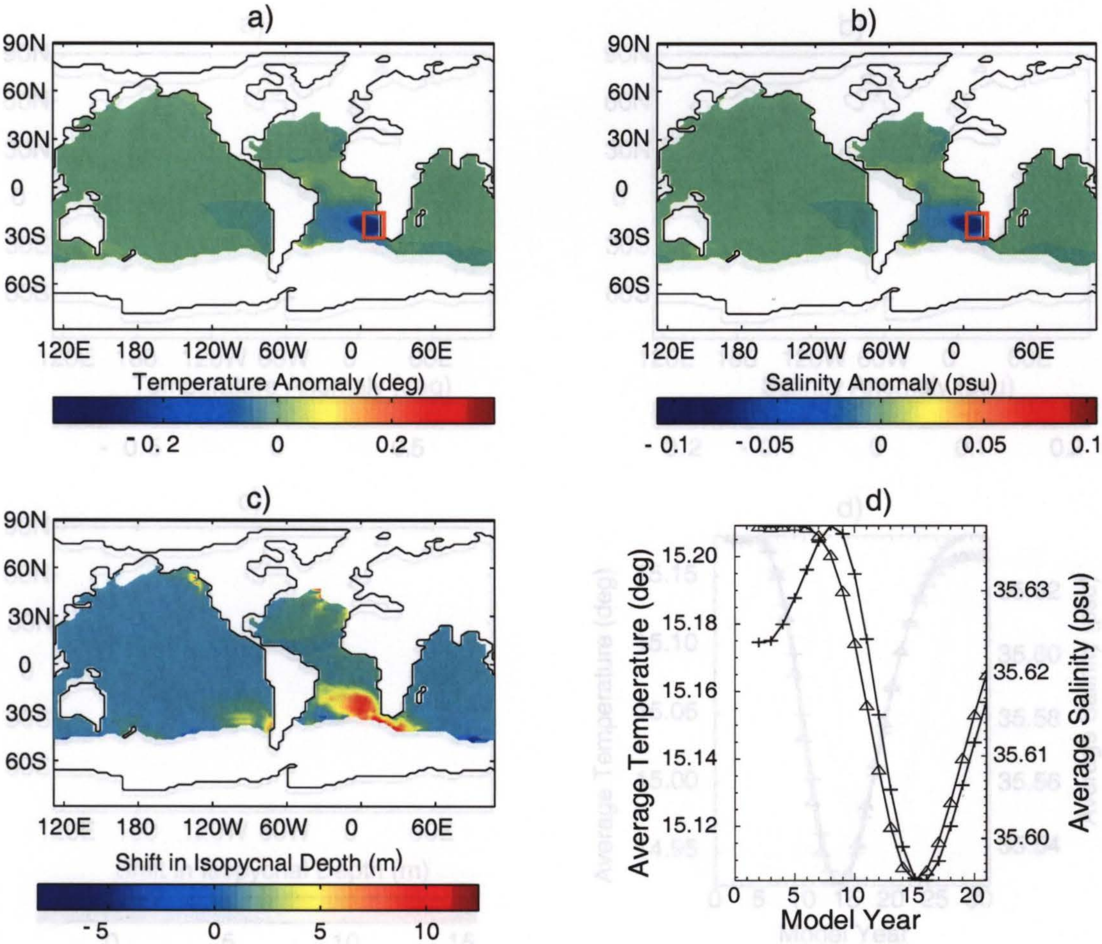

Figure 3.3: **Anomalous a) potential temperature and b) salinity response, in year 15, to 5 years of 150 W/m^2 surface heat flux applied in Region A. Anomalies are shown on the sigma-theta = 26.5 isopycnal. c) Change in the depth of the sigma-theta = 26.5 isopycnal. d) Average potential temperatures (crosses) and salinities (triangles) inside the box, as a function of time (as in Fig. 3.2).**

3.3 *Freshwater and Heat Forcing Experiments in Region B*

Forcing experiments in Region B, a known location of Antarctic Intermediate Water formation, gave very similar results. The results for a 5-year, 0.1 Sv applied surface freshwater flux in Region B are shown in Figure 3.4. Figure 3.5 shows results for a 5-year applied surface heat flux of 150 W/m^2.

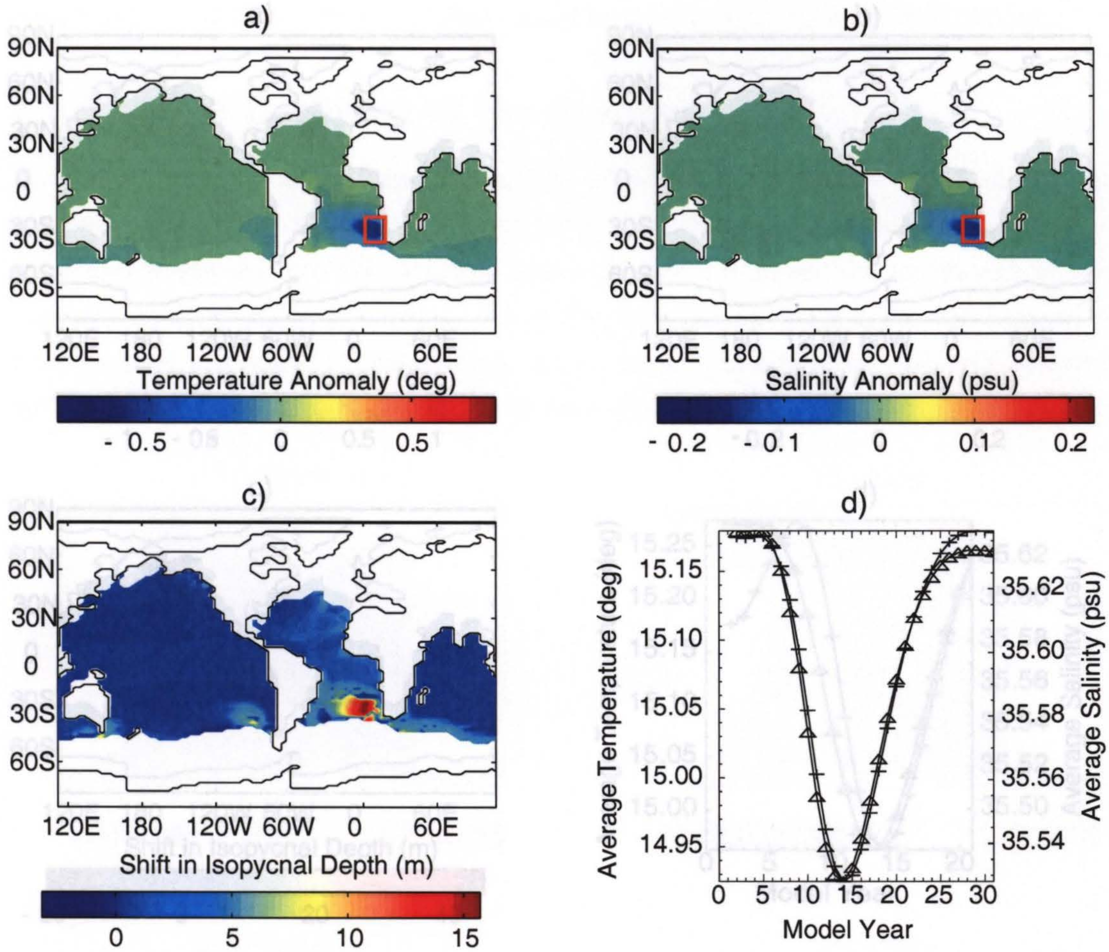

Figure 3.4: **Anomalous a) potential temperature and b) salinity response, in year 14, to 5 years of 0.1 Sv freshwater forcing in Region B. Anomalies are shown on the sigma-theta = 26.5 isopycnal. c) Change in the depth of the sigma-theta = 26.5 isopycnal. d) Average potential temperatures (crosses) and salinities (triangles) inside the box, as a function of time (as in Fig. 3.2).**

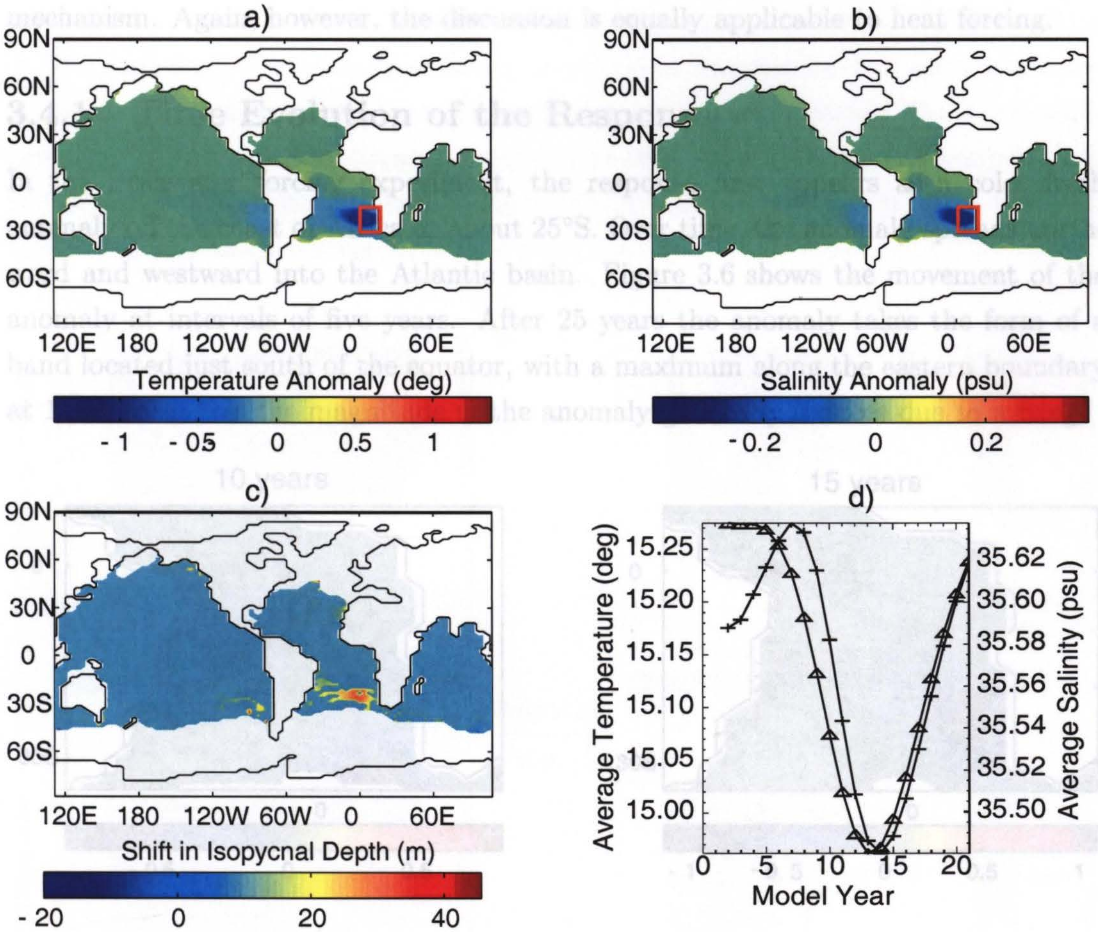

Figure 3.5: **Anomalous a) potential temperature and b) salinity response, in year 14, to 5 years of 150 W/m^2 thermal forcing in Region B. Anomalies are shown on the sigma-theta = 26.5 isopycnal. c) Change in the depth of the sigma-theta = 26.5 isopycnal. d) Average potential temperatures (crosses) and salinities (triangles) inside the box, as a function of time (as in Fig. 3.2).**

3.4 Analysis of the Response

Though the responses to forcing in Region A and Region B are very similar, we concentrate our discussion on forcing in Region A because it is a region of higher variability in the real ocean. We believe that the processes are similar and the discussion applies to both. We also focus on freshwater rather than heat forcing experiments, because interannual variability in Southern Ocean sea ice is a plausible freshwater forcing mechanism. Again, however, the discussion is equally applicable to heat forcing.

3.4.1 Time Evolution of the Response

In the freshwater forcing experiment, the response first appears as a cold, fresh anomaly off the coast of Africa at about 25°S. Over time, the anomaly spreads northward and westward into the Atlantic basin. Figure 3.6 shows the movement of the anomaly at intervals of five years. After 25 years the anomaly takes the form of a band located just south of the equator, with a maximum along the eastern boundary at 15°S. After this the magnitude of the anomaly gradually reduces due to mixing.

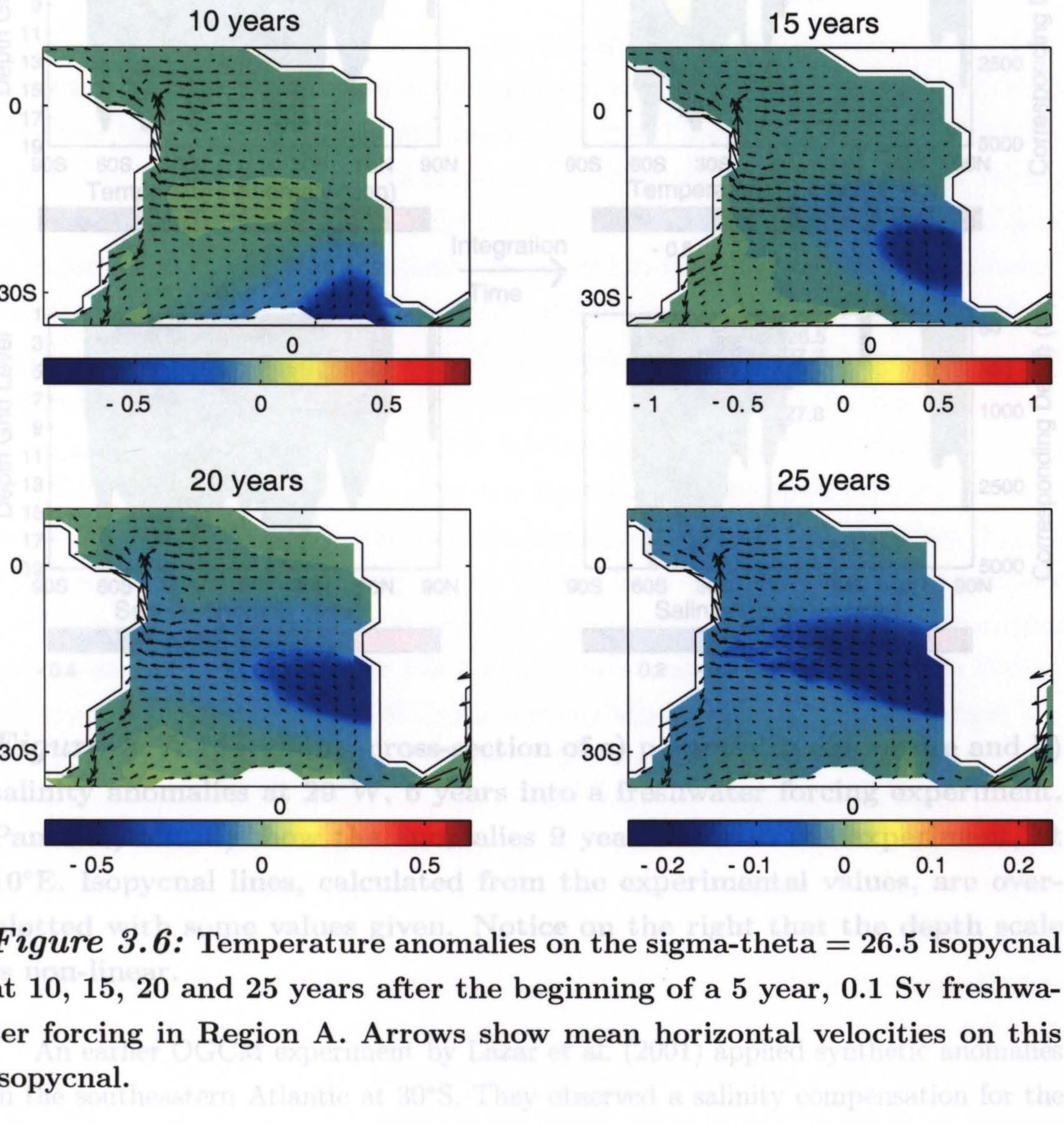

Figure 3.6: **Temperature anomalies on the sigma-theta = 26.5 isopycnal at 10, 15, 20 and 25 years after the beginning of a 5 year, 0.1 Sv freshwater forcing in Region A. Arrows show mean horizontal velocities on this isopycnal.**

Figure 3.7 shows meridional cross-sections from the freshwater experiment. Temperature and salinity anomalies can be seen to propagate northward along the sigma = 26.5 isopycnal. The salinity anomaly is visible at the surface, but surface temperature anomalies are largely attenuated due to air–sea interaction. The northward-propagating, negative temperature anomaly is associated with warmer temperatures in the subducting Antarctic Intermediate Water, indicated by the positive temperature anomalies along the sigma = 27.6 isopycnal.

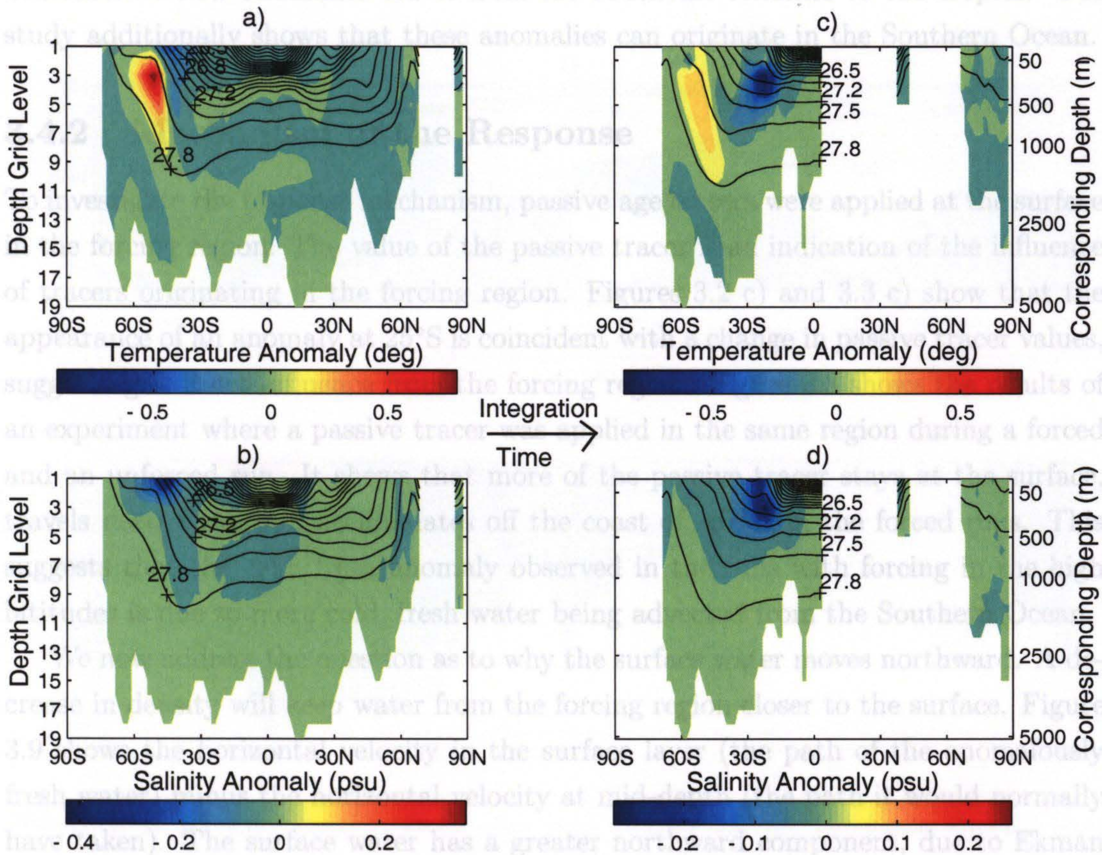

Figure 3.7: **Meridional cross-section of a) potential temperature and b) salinity anomalies at 29°W, 6 years into a freshwater forcing experiment. Panels c) and d) show the anomalies 9 years later in the experiment, at 10°E. Isopycnal lines, calculated from the experimental values, are overplotted with some values given. Notice on the right that the depth scale is non-linear.**

An earlier OGCM experiment by Lazar et al. (2001) applied synthetic anomalies in the southeastern Atlantic at 30°S. They observed a salinity compensation for the temperature-related density change, and showed that the anomalies were carried by mean currents along isopycnals in the mixed layer to the western tropical Atlantic (their experiments applied anomalies on the shallower sigma = 25.3 isopycnal). Our forcings in Region A show a strong response off west Africa, similar to the starting region for Lazar et al.'s experiments, and the anomalies then follow a very similar path to the tropics, though below the mixed layer. The travel time is also slower in our experiments, because of the coarser grid spacing in our model. Both studies demonstrate how anomalies travel from the southeast Atlantic to the tropics. Our study additionally shows that these anomalies can originate in the Southern Ocean.

3.4.2 Mechanism of the Response

To investigate the response mechanism, passive age tracers were applied at the surface in the forcing region. The value of the passive tracer is an indication of the influence of tracers originating in the forcing region. Figures 3.2 c) and 3.3 c) show that the appearance of an anomaly at 25°S is coincident with a change in passive tracer values, suggesting a direct influence from the forcing region. Figure 3.8 shows the results of an experiment where a passive tracer was applied in the same region during a forced and an unforced run. It shows that more of the passive tracer stays at the surface, travels northward, and accumulates off the coast of Africa in the forced runs. This suggests that the cold, fresh anomaly observed in the runs with forcing in the high latitudes is due to more cold, fresh water being advected from the Southern Ocean.

We now address the question as to why the surface water moves northward. A decrease in density will keep water from the forcing region closer to the surface. Figure 3.9 shows the horizontal velocity in the surface layer (the path of the anomalously fresh water) minus the horizontal velocity at mid-depth (the path it would normally have taken). The surface water has a greater northward component, due to Ekman transport acting to the left of the strong westerly winds, and so anomalously light water masses will travel further northward—in this case, past the tip of Cape Horn, out of the circumpolar current and into the tropical Atlantic. More surface water moving north means less subducting in the Southern Ocean—hence the positive temperature anomalies at densities of Antarctic Intermediate Water formation, as seen in Figure 3.7.

The density-forcing mechanism also explains why a *positive* heat forcing in Region A leads to a *negative* temperature response in the southeast Atlantic (as shown in Figure 3.3). Similar to what happens in the freshwater forcing experiment, heating the surface water in Region A makes it less dense, which keeps it closer to the surface, which gives it a greater northward velocity, which results in more Southern Ocean water being carried to the southeastern Atlantic. In Region A, this water was originally warmer than its surrounding environment (due to the heat forcing), but by the time it has travelled to the warmer Atlantic waters, it is colder than its surrounding environment. This transition is confirmed by what we see in the model: an early, positive temperature anomaly in the Southern Ocean becomes a negative anomaly as it travels toward the tropics.

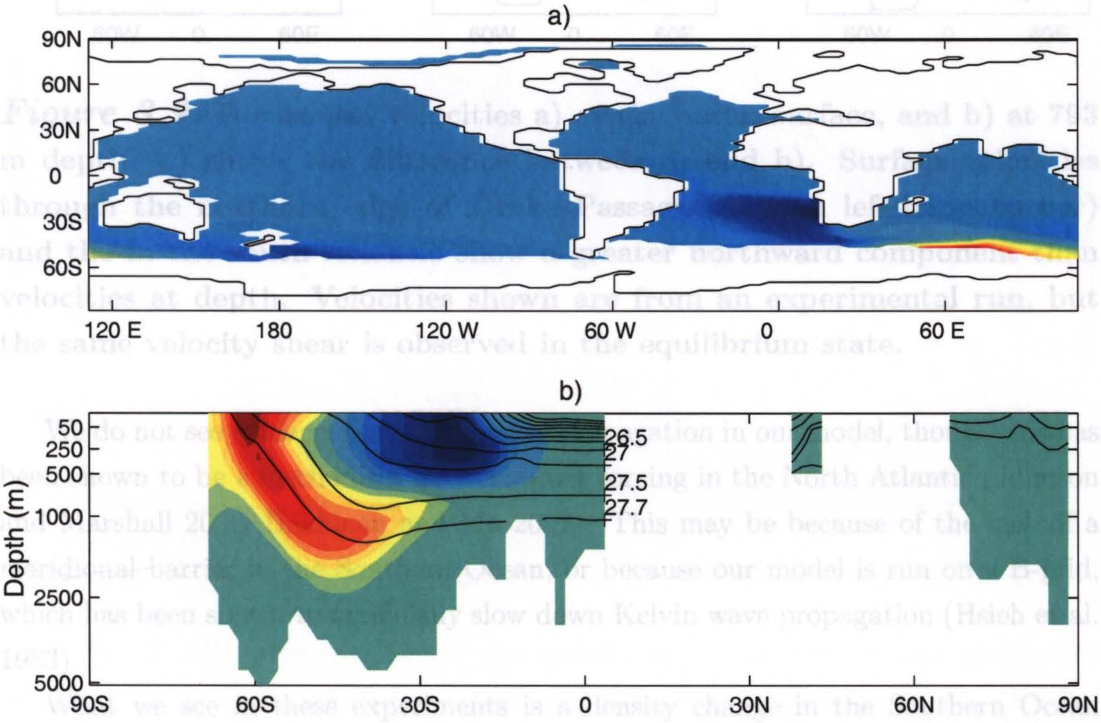

Figure 3.8: **Difference in passive age tracer values between forced and unforced runs. Values are shown on a) the sigma-theta = 26.5 isopycnal, and b) at 10°E, 15 years into a freshwater forcing experiment. Lower tracer values (blue colours) generally indicate a higher presence of tracers from the forcing region; here, blue-coloured regions indicate the presence of more tracers from forcing Region A in the experimental runs. Isopycnal lines shown in panel b) are calculated from the experimental values.**

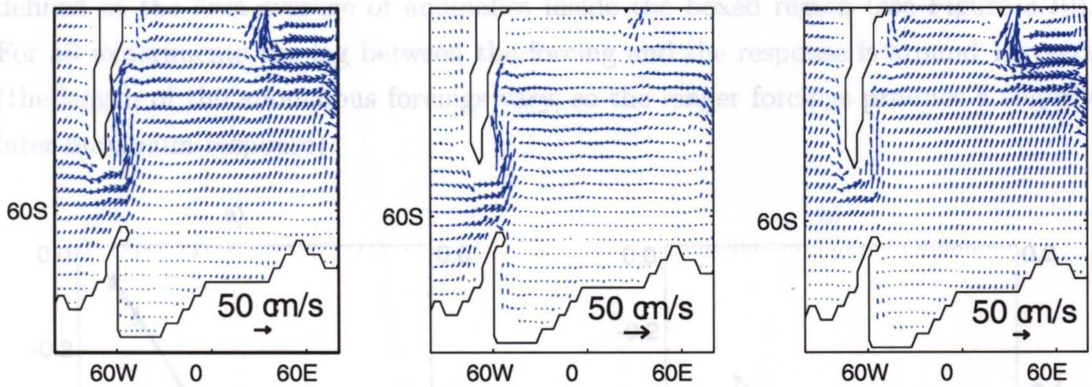

Figure 3.9: **Horizontal velocities a) at the ocean surface, and b) at 793 m depth. c) shows the difference between a) and b). Surface velocities through the northern edge of Drake Passage (bottom left-hand corner) and the in the south Atlantic show a greater northward component than velocities at depth. Velocities shown are from an experimental run, but the same velocity shear is observed in the equilibrium state.**

We do not see evidence for Kelvin wave propagation in our model, though this has been shown to be a mechanism for freshwater forcing in the North Atlantic (Johnson and Marshall 2002, Häkkinen and Mo 2002). This may be because of the lack of a meridional barrier in the Southern Ocean, or because our model is run on a B-grid, which has been shown to artificially slow down Kelvin wave propagation (Hsieh et al. 1983).

What we see in these experiments is a density change in the Southern Ocean surface water that leads to a change in circulation patterns. Lower-density surface water in the southern high latitudes results in more cold, fresh surface water from the Southern Ocean being transported northward and out of the circumpolar current, into the eastern Atlantic basin. The opposite is not true—a negative freshwater forcing in Region A does *not* lead to a warm, salty anomaly in the west African region. The effects of increased salinity are limited to the Southern Ocean basin, largely at depth and east of Drake Passage (results not shown), because the increased density induces increased convection below the sill depth.

3.4.3 Sensitivity Tests

A series of sensitivity experiments show a linear relationship between the strength and duration of the freshwater forcing in Region A and the magnitude of the response, defined as the area-average of anomalies inside the boxed region (see Figure 3.10). For all experiments the lag between the forcing and the response is around 12 years (the length of the anomalous forcings vary, so the longer forcings produce a slightly later maximum response).

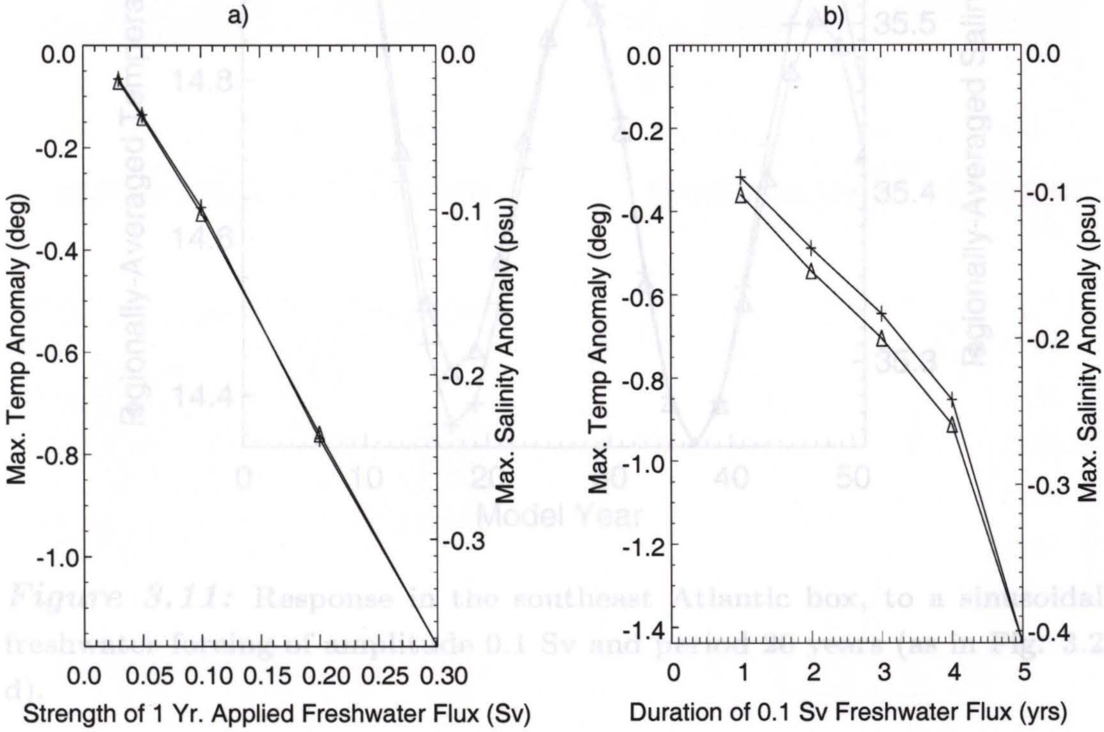

Figure 3.10: **Amplitude of the maximum response in the southeast Atlantic (boxed) region for increasing freshwater forcing strength and duration (as in Fig. 3.2 d).**

An experiment with a sinusoidally-varying freshwater signal produces a sinusoidal response with the same period (see Figure 3.11), showing that decadal variability in the Southern Ocean could be the source of decadal Atlantic SST. Again, the response lags the forcing by 12 years. If we posit a teleconnection feedback, this is a possible candidate for the oceanic branch of a mode of variability.

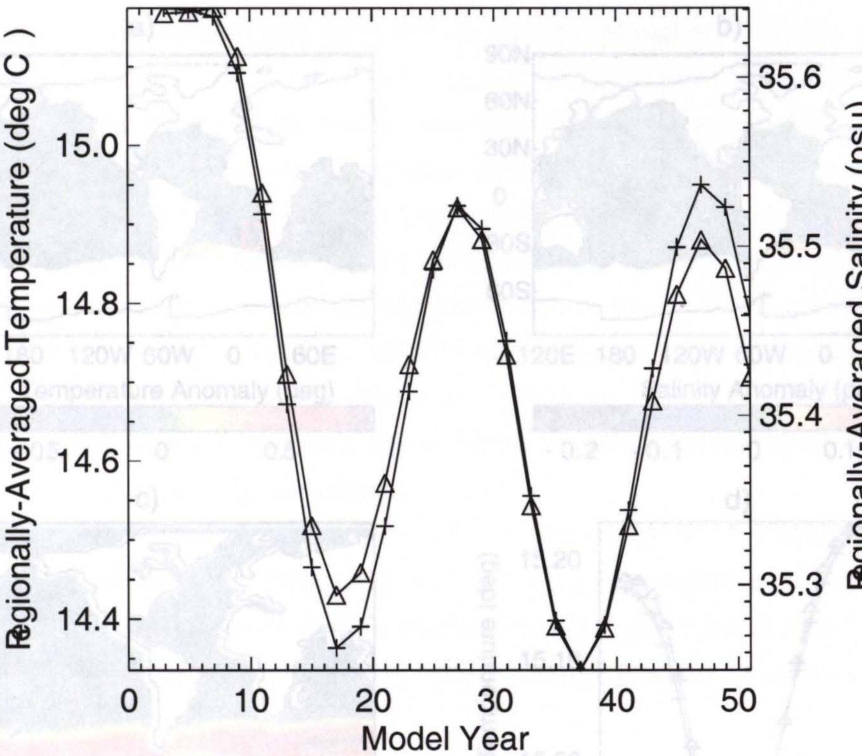

Figure 3.11: **Response in the southeast Atlantic box, to a sinusoidal freshwater forcing of amplitude 0.1 Sv and period 20 years (as in Fig. 3.2 d).**

3.5 Wind Forcing Experiments

In a final experiment, the winds were changed over the entire Southern Ocean in the manner described in Chapter 2. The response to a 5-year, sinusoidal increase in wind strength in the Southern Ocean, shown in Figure 3.12, is again a cold, fresh pool in the southeast Atlantic.

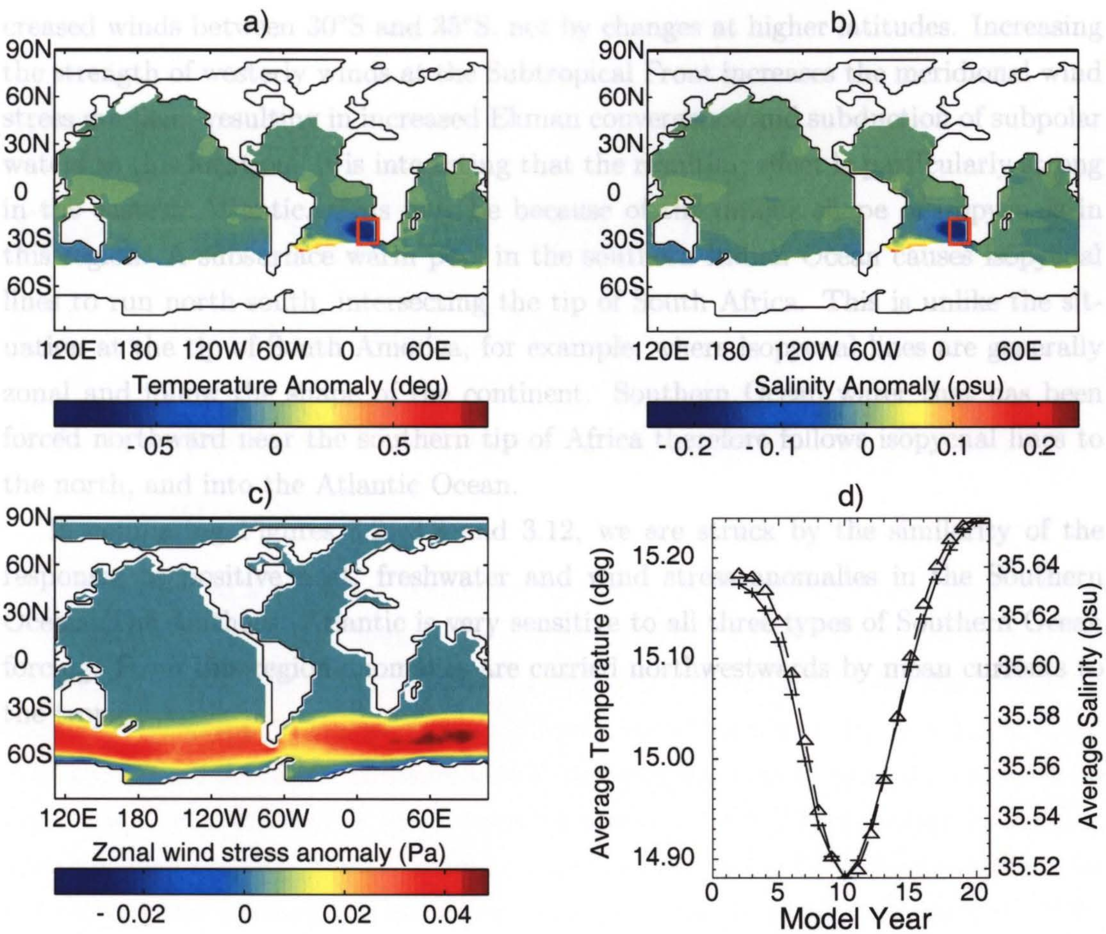

Figure 3.12: **Response to a 5 year pulse of increased winds in the Southern Ocean. Panels a) and b) show the potential temperature and salinity anomalies, 10 years into the experiment, on the sigma-theta = 26.5 isopycnal. Panel c) shows the applied wind stress forcing. Panel d) shows the average temperature and salinity in the boxed region as a function of time (as in Fig. 3.2).**

The original intent of the wind experiments was to force changes in sea ice dynamics, thereby indirectly causing freshwater anomalies in the Southern Ocean. The effect is more direct, however — an experiment where wind stress forcing was applied only on sea ice and not to open water produced negligible anomalies. Instead, increasing westerly wind strength over the Southern Ocean affects surface circulation directly by increasing the northward Ekman transport.

A separate experiment, where winds were increased only over small latitudinal bands, shows that the southeast Atlantic response is caused almost entirely by increased winds between 30°S and 35°S, not by changes at higher latitudes. Increasing the strength of westerly winds at the Subtropical Front increases the meridional wind stress gradient, resulting in increased Ekman convergence and subduction of subpolar waters at this location. It is interesting that the resulting effect is particularly strong in the eastern Atlantic. This may be because of the unique shape of isopycnals in this region. A subsurface warm pool in the southern Indian Ocean causes isopycnal lines to run north-south, intersecting the tip of South Africa. This is unlike the situation at the tip of South America, for example, where isopycnal lines are generally zonal and follow the shape of the continent. Southern Ocean water that has been forced northward near the southern tip of Africa therefore follows isopycnal lines to the north, and into the Atlantic Ocean.

In comparing Figures 3.2, 3.3 and 3.12, we are struck by the similarity of the responses to positive heat, freshwater and wind stress anomalies in the Southern Ocean. The southeast Atlantic is very sensitive to all three types of Southern Ocean forcing. From this region anomalies are carried northwestwards by mean currents to the tropics.

Chapter 4

Comparison of Model Results with Observational Data

4.1 Principal Component Analysis of South Atlantic SST

Having discovered a consistent southeast Atlantic response to Southern Ocean forcing in the model experiments, here we go one step further, and present evidence for this connection in the real ocean. Figure 4.1 shows the model SST associated with the subsurface response in the southeast Atlantic, shown earlier in Figures 3.2, 3.3 and 3.12 a). There are some consistent SST patterns associated with the response to Southern Ocean forcing: a weak negative centre off the African coast at 20°S; a centre of opposite sign in the Agulhas current region, at the tip of South Africa; and a positive band crossing the basin from northwest (at 40°S) to southeast (at 55°S). We decided to look for evidence of this SST pattern in the observational record.

Most previous analyses of tropical Atlantic variability, as described in the introduction, consider only SST data between 30N to 30S. This choice of domain implicitly limits the search to finding connections within the tropics; it is perhaps not surprising that early statistical analyses of tropical Atlantic variability found a fictitious dipole linking variability in the northern and southern tropics. We performed EOF analysis with the hypothesis that there may be a connection between the southern tropical Atlantic and the Southern Ocean.

We performed a principal component analysis (PCA) of SST over the entire South Atlantic region, extending from the tropics all the way down to Antarctica. PCA is a statistical technique useful for identifying dominant patterns of variability in data. The PCA calculates the eigenvalues and eigenvectors (empirical orthogonal functions, or EOFs) of the covariance matrix for the original data. Each EOF, or leading pattern, is then projected back onto the original data to obtain a timeseries of "principal components" for the pattern. The eigenvalues indicate how much of the total variance in the data is explained by each EOF. PC analysis reduces a large dataset with many degrees of freedom to a set of patterns with a smaller degree of freedom that still describe the data. In these experiments, two different long-term data reconstructions were analyzed: the HadISST dataset and the ERSST dataset (descriptions below).

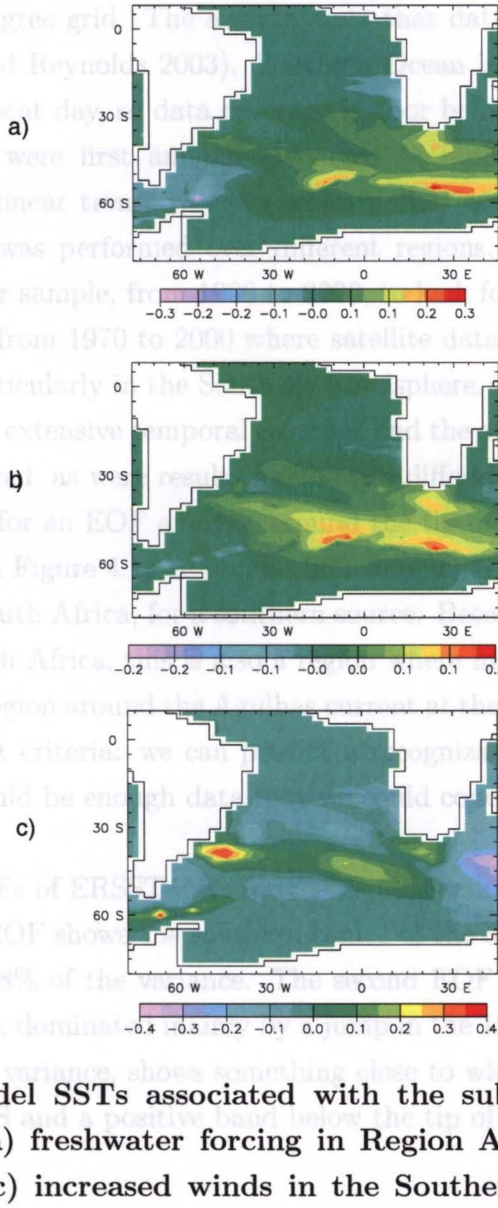

Figure 4.1: **Model SSTs associated with the subsurface response in experiments with a) freshwater forcing in Region A, b) thermal forcing in Region A, and c) increased winds in the Southern Ocean. The SST patterns have three features in common: a weak negative centre off the African coast at 20°S; a centre of opposite sign in the Agulhas current region, at the tip of South Africa; and a positive band crossing the basin from northwest (at 40°S) to southeast (at 55°S).**

4.1.1 NOAA Extended Reconstructed SST (ERSST) data

The NOAA Extended Reconstructed SST dataset was released earlier this year. This is a reconstruction based on the Comprehensive Ocean–Atmosphere Data Set (COADS) SST data from 1854 to the present, averaged as monthly means and interpolated onto a 2 degree grid. The authors note that data points are very sparse before 1880 (Smith and Reynolds 2003). Southern Ocean in situ measurements are sparse even in the present day, so data coverage is poor before the satellite era.

The ERSST data were first annually-averaged to limit the study to interannual variability. The linear trend over the entire period was then removed at each grid point. Analysis was performed over different regions, and over two different timescales: an 100-year sample, from 1900 to 2000, to look for long-term trends, and also a 30-year sample from 1970 to 2000 where satellite data is available and spatial coverage is better, particularly in the Southern Hemisphere. The results of these two time periods (one with extensive temporal coverage and the other with reliable spatial coverage) were compared, as were results for the two different datasets.

Results are shown for an EOF analysis around the tip of South Africa. From the model results shown in Figure 4.1, we predict SST anomalies of opposite sign at 20°S and along the tip of South Africa, for a southern source. Because shipping routes pass around the tip of South Africa, this is also a region where historical SST records are relatively good. The region around the Agulhas current at the tip of South Africa thus satisfies two important criteria: we can predict a recognizable pattern for southern forcing, and there should be enough data that we could conceivably detect a pattern if it was there.

The first three EOFs of ERSST data for the Agulhas current region are shown in Figure 4.2. The first EOF shows the southern centre of the decadal mode, centred at 15°S and explaining 38% of the variance. The second EOF shows a large center at the tip of South Africa, dominated mainly by a jump in the 1960s. The third pattern, explaining 14% of the variance, shows something close to what we are looking for: a negative center at 15°S and a positive band below the tip of South Africa.

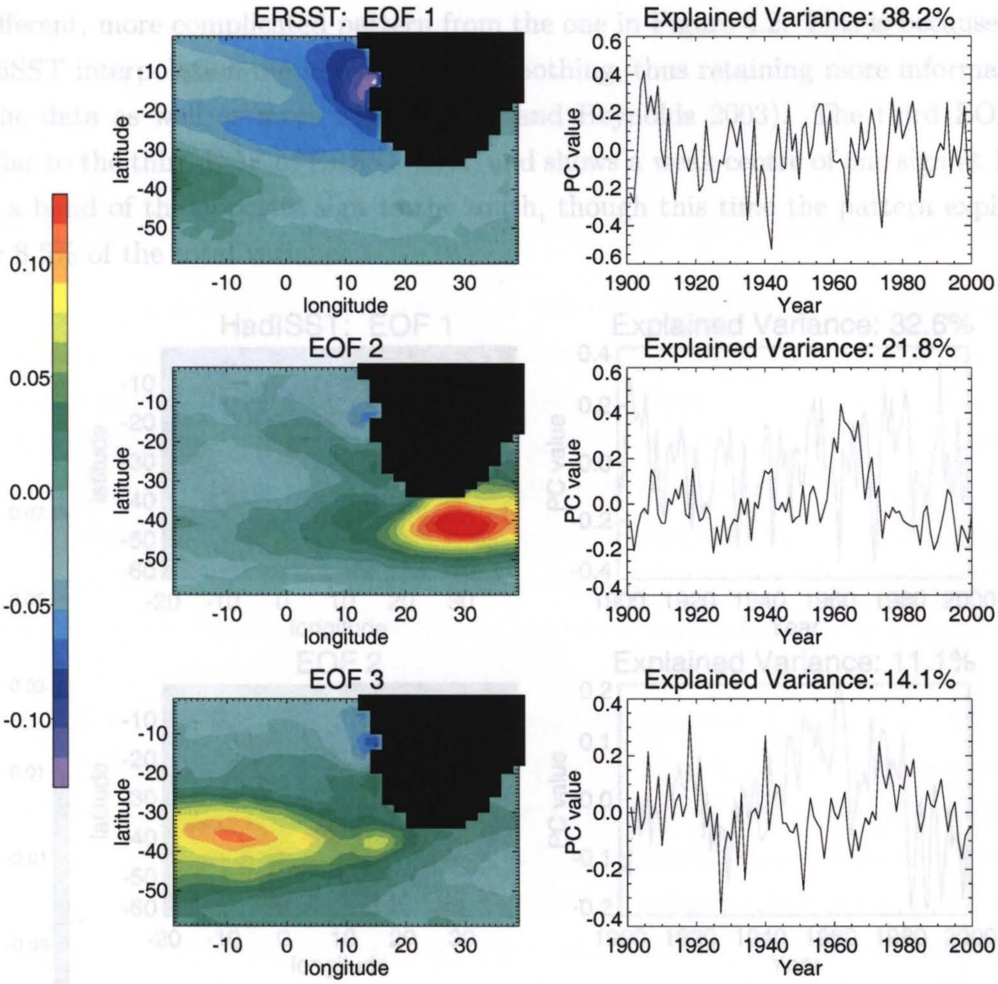

Figure 4.2: **Dominant patterns of variability in ERSST data for the Agulhas current region. Principal components and explained variance are shown to the right of each pattern.**

4.1.2 Hadley Centre (HadISST) Data

The same analysis was done with another major global SST data reconstruction, the Hadley Centre's HadISST v.1.1 (this is an operational version, launched in February 2003, of the older GISST dataset). The HadISST data are monthly mean SSTs on a 1 deg grid, extending from 1870 to present day. Older records are corrected interpolations based on observations, while newer records are a blend of satellite data and observations.

The first three EOFs for the Agulhas current region are shown in Figure 4.3. The first pattern is, again, the southern centre of the decadal mode. The second EOF is a different, more complicated pattern from the one in Figure 4.2. This is because the HadiSST interpolation method uses less smoothing, thus retaining more information in the data as well as more noise (Smith and Reynolds 2003). The third EOF is similar to the third EOF of ERSST data, and shows a weak centre of one sign at 15°S and a band of the opposite sign to the south, though this time the pattern explains only 8.5% of the total variance.

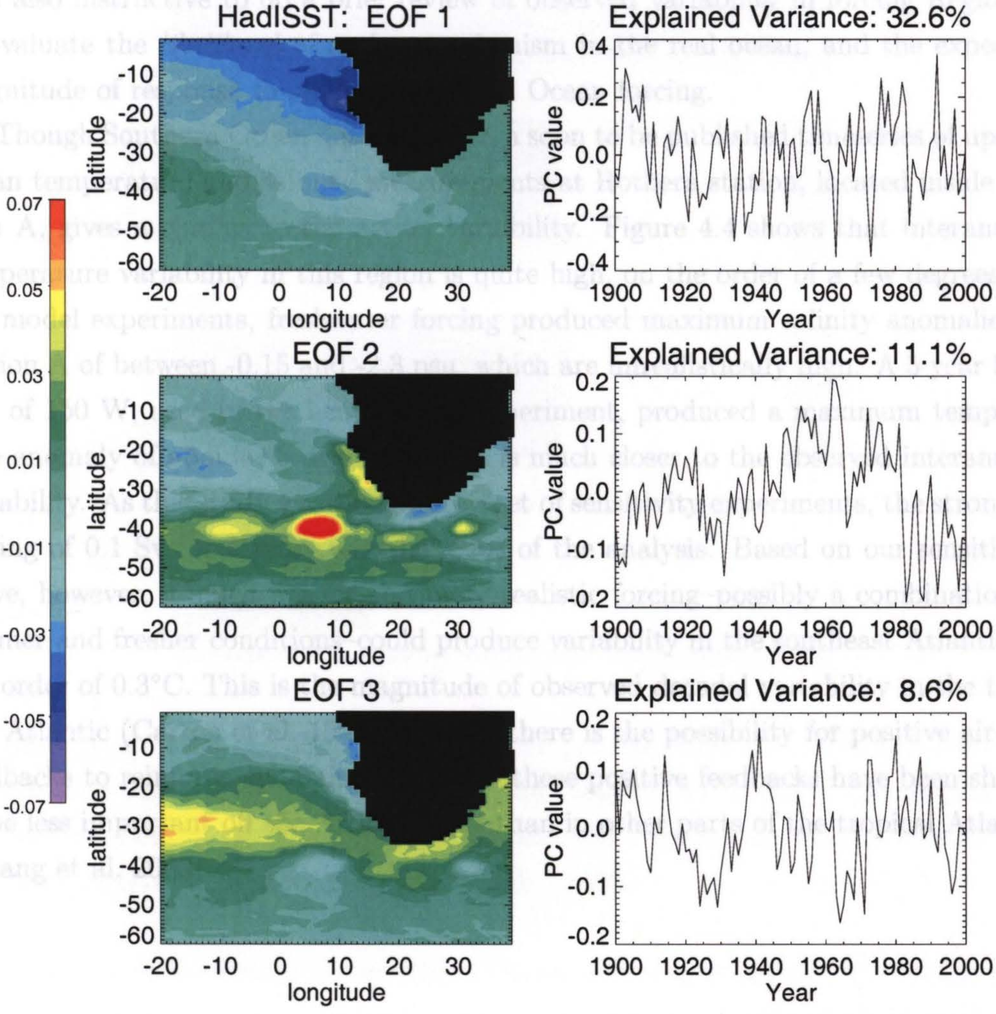

Figure 4.3: **Dominant patterns of variability in HadISST data for the Agulhas current region. Principal components and explained variance are shown to the right of each pattern.**

From these analyses, we find weak support for the Southern Ocean forcing pattern seen in the model. Were the southern centre of decadal variability to be forced entirely by the Southern Ocean, we would expect the first pattern to be associated with a centre of opposite sign to the south. Instead, we find the decadal mode explained mainly by the first EOF, and a weaker pattern, which appears as the 3rd EOF, corresponds to the pattern seen in model results.

4.2 Observed Variability in the Southern Ocean

It is also instructive to do a brief review of observed variability in forcing Region A to evaluate the *likelihood* of such a mechanism in the real ocean, and the expected magnitude of response to realistic Southern Ocean forcing.

Though Southern Ocean data is sparse, a soon to be published timeseries of upper-ocean temperature and salinity measurements at Rothera station, located inside Region A, gives a clue as to the actual variability. Figure 4.4 shows that interannual temperature variability in this region is quite high, on the order of a few degrees. In the model experiments, freshwater forcing produced maximum salinity anomalies in Region A of between -0.15 and -2.3 psu, which are unrealistically high. A 5 year heat flux of 150 W, used in the heat forcing experiment, produced a maximum temperature anomaly of about +3 degrees, which is much closer to the observed interannual variability. As this study was essentially a set of sensitivity experiments, the strongest forcing of 0.1 Sv for 5 years was the focus of the analysis. Based on our sensitivity curve, however, a much weaker and more realistic forcing–possibly a combination of warmer and fresher conditions–could produce variability in the southeast Atlantic on the order of 0.3°C. This is the magnitude of observed decadal variability in the tropical Atlantic (Carton et al. 1996). Finally, there is the possibility for positive air–sea feedbacks to reinforce anomalies, although these positive feedbacks have been shown to be less important off the African coast than in other parts of the tropical Atlantic (Chang et al. 2000).

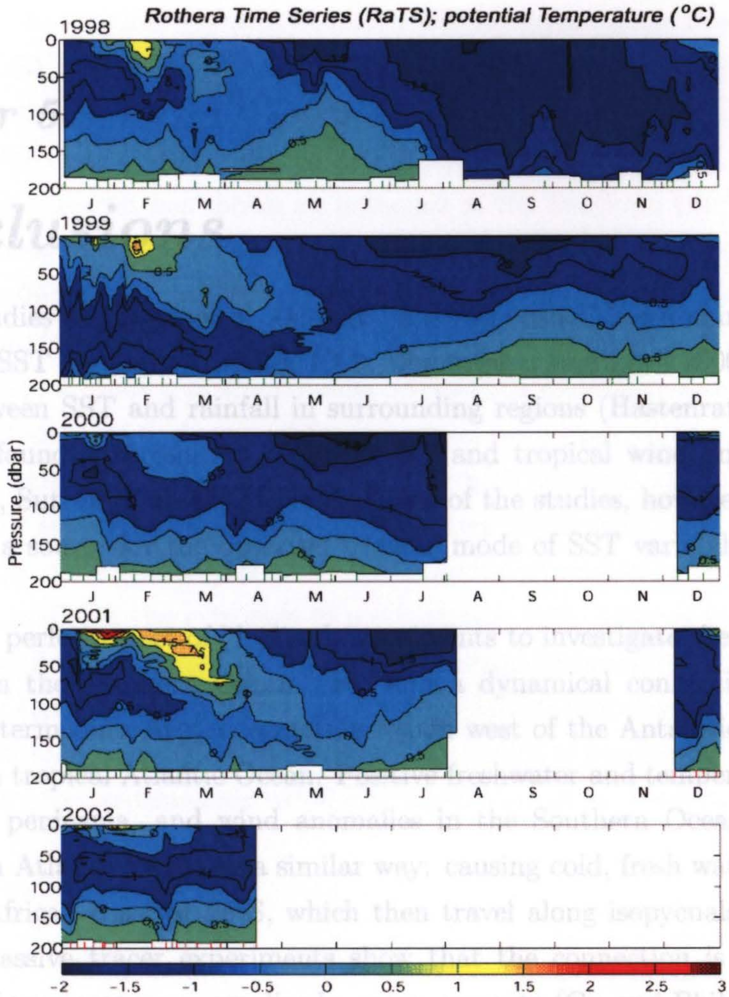

Figure 4.4: **Timeseries of temperature profiles taken at Rothera Station, west of the Antarctica Peninsula (inside Region A). Data is missing in the winters of 2000 and 2001 due to unsafe ice conditions and a laboratory fire, respectively. From Meredith et al. (2003).**

Finally, we ask whether the Southern Ocean shows signs of periodic variability on decadal timescales. The Antarctic Circumpolar Wave, a 4–5 year mode of variability in sea ice, SLP and SST, has been identified in the data by White and Peterson (1996). There is also some suggestion of decadal-scale variability in the ACW (Venegas et al. 1996, Carril and Navarra 2001, Venegas et al. 2001). According to our model results, if this decadal variability does exist, it would translate into decadal variability in the southeastern Atlantic ocean temperatures.

Chapter 5

Conclusions

Previous studies of the tropical Atlantic have established the dominant patterns of interannual SST variability (Mehta 1998, Dommenget and Latif 2000), shown a connection between SST and rainfall in surrounding regions (Hastenrath 1976, Servain 1991), and found a correlation between SST and tropical wind anomalies (Chang et al. (1997), Sutton et al. (2000), etc.). None of the studies, however, has been able to establish a source for the observed decadal mode of SST variability in the South Atlantic.

Here, we perform a set of idealized experiments to investigate the effect of surface variability in the Southern Ocean. We find a dynamical connection between the Antarctic Intermediate Water formation region west of the Antarctic peninsula, and the southern tropical Atlantic Ocean. Positive freshwater and temperature anomalies west of the peninsula, and wind anomalies in the Southern Ocean, all affect the southeastern Atlantic Ocean in a similar way: causing cold, fresh water anomalies off the South African coast at 25°S, which then travel along isopycnals to the tropical Atlantic. Passive tracer experiments show that the connection is neither a strict advection of temperature anomalies by mean currents (Gu and Philander 1997), nor a change in overturning strength (Kleeman et al. 1999); rather, the anomalously light water follows a different isopycnal with a more northerly component, and then affects circulation patterns resulting in more cold, fresh water from the Southern Ocean reaching the low-latitude Atlantic. The pathway of the anomalies from the subtropics to the tropics agrees well with that previously shown by Lazar et al. (2001). Our work may also complement Lazar et al.'s study by providing a more plausible source for generating anomalies, given the high seasonal and interannual variability—in both temperature and salinity—observed in the Southern Ocean (White and Peterson 1996, Jacobs and Comiso 1997, Meredith et al. 2003).

Our sensitivity experiments showed a linear relationship between the strength of the southern forcing and the magnitude of the response, and a sinusoidal forcing was shown to produce a sinusoidal response of the same period. Our analysis therefore suggests that low-frequency variability in the Southern Ocean, as suggested by Carril and Navarra (2001) and Venegas et al. (1996, 2001), would translate into low-frequency variability in the tropical Atlantic. Additionally, an atmospheric teleconnection linking tropical SST back to the Southern Ocean is possible—several studies have found evidence that tropical Atlantic SST influences the North Atlantic Oscillation (Robertson et al. 2000, Ruiz-Barradas et al. 2000, Rajagopalan et al. 1998, Okumura et al. 2001), though none yet has shown an influence in the Southern hemisphere.

The observational data provide some support for the model results. Observed variability in forcing Region A is quite high, and the sensitivity experiments predict that a realistic forcing would produce a magnitude of response comparable with observed decadal variability in the tropical Atlantic. Principal component analysis of historical SST data, to find support for the model connection, give mixed results: the dominant pattern does not compare favourably with patterns seen in the model, but the third EOF suggests that southern forcing may play a role.

A more complete investigation of this oceanic pathway will require more data—this study establishes the existence of a connection and, perhaps more importantly, pinpoints specific areas of the Atlantic as being particularly sensitive to influence from the Southern Ocean. Combined with the earlier results from Lazar et al. (2001), this study suggests that we should look south for the source of tropical Atlantic anomalies, particularly for longer-timescale variability where oceans are likely to play a key role.

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