CLIVAR WORKSHOP

on

SHALLOW TROPICAL-SUBTROPICAL OVERTURNING CELLS (STCs) and THEIR INTERACTION WITH THE ATMOSPHERE

October 9-13, 2000, Venice, Italy

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I. Introduction and Motivation

The Meridional Overturning Circulations (MOCs), in particular the Atlantic Ocean MOC, have been the object of numerous investigations during the last decade. Much less attention has been given to observational and modelling studies of the shallow subtropical/tropical overturning cells (STCs) that can act as a mechanism for transferring mass, heat, salt and tracers between the subtropical and equatorial gyres. Through their effect on the Sea Surface Temperature (SST), the STCs have been proposed as the oceanic component of coupled modes of air-sea variability that influence atmospheric climate on multiple time scales, from the seasonal to the interannual, decadal and multi-decadal.

Hence a workshop was convened in Venice, October 9-13, 2000, under the CLIVAR banner, to bring together observationalists and modellers to assess our present understanding of the structure of these cells and of their influence on the atmosphere. This assessment will be used to develop strategies for future observational and modelling studies, here proposed as recommendations to the CLIVAR Implementation Panels for the three oceans.

I.I Relationship to CLIVAR

Understanding and quantifying the role of STCs in the climate system are relevant to the major objectives of the CLIVAR Program. Specifically, STCs are important in relation to research objectives of CLIVAR-DecCen program D2 (WCRP, 1995): to our knowledge, no meeting/workshop/symposium was ever held before on this subject.

There are several programs proposed under the CLIVAR umbrella that provide a larger scale framework applicable to the STCs: PBECS, TAV/COSTA among others. However, the dynamics of the STCs are common potentially to all oceans. A goal of this workshop has been to design an effort dedicated to observing and modelling the STCs globally.

I.II Objectives

Overall Objective: to assess the present understanding of the structure and dynamic of the STCs and their interaction with the atmosphere and to develop strategies for future observational and modelling studies.

Specific Objectives: Compare model results and observations of the mean and time-dependent STCs to address the following issues:

- What are the sources for and what determines the rate of subduction of subtropical waters that contribute to the shallow cells?
- What are the pathways and time scales from the subtropics to tropical upwelling areas (e.g. western boundary currents and/or interior ventilation)?
- What processes determine the intensity of equatorial upwelling (e.g. local winds versus remote forcing)?
- What are the pathways for the return upwelled waters to the subtropical subduction region?
- What is the role of the global thermohaline circulation in influencing the STC structure and intensity?
- What processes control the effect on the atmosphere of SST variability induced by the STCs?

Based on the understanding of the above issues, propose a strategy for:

- An air-sea network to observe the STCs and their effect on the atmosphere.
- Numerical modelling activities to increase the understanding of the STCs.

II. Structure of the Workshop

The workshop consisted of science sessions with invited overview presentations on the three oceans focussing on a) observations; b) ocean models; c) coupled systems/models. These overviews were given during the first three days, (Monday, Tuesday, Wednesday), with each day devoted to one specific ocean (Pacific, Indian, Atlantic). In parallel to the invited presentations and plenary discussion sessions, at the end of the first two days specific sessions were held on contributed posters, that remained up for discussion for the entire duration of the workshop.

At the end of the Wednesday session, the mandate was specified to the three Working Groups (WGs). They met on Thursday and Friday morning. At the end of the Friday, the WG reports were presented in plenary session.

The detailed workshop agenda is given in Appendix A. The list of participants is given in Appendix B.

III. Scientific Priorities for the Three Oceans

Consensus was reached in the plenary session of Wednesday evening to synthesize the specific objectives of the workshop into the following five questions that constitute the priority issues to be addressed in the WGs for each ocean.

- **Question 1 (Q1):** Do STCs play a role in seasonal to centennial climate variability, and, if so, how?

- **Question 2 (Q2) :** What are the sources and pathways of STCs, including other features such as the Tsuchiya jets, the Meridional Overturning Circulation (MOC) for the Atlantic Ocean, and the Indonesian Throughflow for the Pacific and Indian Oceans?

- **Question 3 (Q3):** How do surface fluxes affect subduction properties and the three-dimensional ocean circulation within the STCs?

- **Question 4 (Q4):** What are the relative mean and time-variable contribution of northern and southern hemisphere STCs to the equatorial circulation?

- **Question 5 (Q5):** How do the STCs affect the mean and time-variable ocean-atmosphere tropical heat budget?

(N.B. Q5 was not separately addressed in the Pacific WG.)

To address the above scientific issues, improved definition and understanding are necessary of the long-term mean and seasonal-to-centennial time scales variations of:

- Water mass properties in the thermocline of the tropical and subtropical oceans.

- The rates and water mass properties of waters subducted in the subtropics, and the regions where subduction occurs.

- Western boundary currents mass and heat transports.

- Equatorial and coastal upwelling rates and source waters.
- Indonesian Throughflow, and its relation to western boundary and interior ocean current transports (Pacific and Indian Oceans).
- Meridional Overturning Circulation (MOC) and how the upper warm return pathways affect the STCs (Atlantic Ocean).
- Eastern boundary termination of major zonal currents such as the Equatorial Undercurrent (EUC), the Tsuchiya Jets, and other thermocline flows.
- The pathways by which upwelled waters return to subtropical subduction zones.
- Surface fluxes of momentum, heat and fresh water.
- Surface and subsurface salinity.

The paramount importance of satellite measurements and atmospheric observations cannot be overemphasized for the understanding of the above seasonal-to-centennial variabilities, and consequently properly address the five major scientific objectives of the workshop. Hence:

- The workshop endorses the development of satellite missions for remote sensing of the global surface salinity field. The European Space Agency has approved one mission (SMOS) and another is under consideration by NASA. The NASA mission is planned for a repeat ground track every 14 days, 70-100 km. Spatial resolution, and an accuracy of O(0.1 psu). These missions have the potential to contribute significantly to studies of the STCs in all ocean basins (Q2 and Q4).
- Continuity of key satellite measurements is essential for climate studies related to the STCs. These measurements include SST, scatterometer winds, sea level, rainfall, insolation and ocean color from which penetrative radiation can be inferred (Q1-4).

- The need is emphasized for improved meteorological packages on vessels of opportunity and increased meteorological observations from moored buoys.

IV. Working Group Reports

IV.I Pacific Ocean Working Group

Evidence for the STCs is found in general in all the three tropical oceans. However, because of extensive ENSO research efforts, and in particular the search for the fundamental causes of ENSO decadal variability, these circulation cells have received significantly more attention in the Pacific than in the Indian or Atlantic Oceans. Pacific modelling efforts are relatively mature, and the Pacific observational monitoring systems have been recently analyzed with the various branches of the STCs and their effect on SST in mind. The modelling and observational experimental strategy discussed in the following builds upon these advances to answer the fundamental questions addressed in the workshop.

Q1) Do STCs play a role in seasonal to centennial climate variability, and, if so, how?

It is well established that tropical Pacific sea surface temperature (SST) anomalies play a central role in determining interannual global climate variability; the global atmosphere is sensitive to tropical Pacific SST on these time scales. There also exist plausible hypotheses to suggest that the climatic influence of tropical Pacific SST variability extend to even longer time scales. Ocean dynamics are clearly important for setting the upwelling rates that help to determine low frequency tropical SST variability; local air-sea fluxes alone cannot close the tropical heat budget on these time scales. In addition, the upwelling strength itself is dependent upon the properties of the tropical Pacific thermocline that is ventilated from the subtropical regions of the Pacific Ocean. Those ocean processes that can potentially impact the tropical upwelling (both interior equatorial *and* coastal) or the mid-latitude subduction processes are therefore of significant interest to those that hope to understand long-term global climate variability That both tropical upwelling and mid-latitude subduction are two of the major branches of the Subtropical Cell (STC) provides a powerful incentive for a program of study directed at understanding the ocean dynamics of this circulation and its role in low frequency climate variability.

Most of our present understanding of the Pacific STCs comes from uncoupled Pacific Ocean models which are robust in generating the canonical mean STC. But much work remains in understanding the variability of the STC and its sensitivity to various model physics, especially the complex diabatic processes of upwelling and subduction that are likely to be model dependent. And coupled ocean-atmosphere models should be designed to isolate the role of the basin-scale STC on the global atmosphere from the more local equatorial circulation. The experimental strategy for answering the fundamental question of the role of the STC in low frequency climate variability is discussed below.

Q2) What are the sources and pathways of STCs, including other features such as the Tsuchiya Jets and the Indonesian Throughflow?

The Pacific STC pathways that are of potential climatic significance must be demonstrated to clearly impact tropical SST, eventually either through equatorial and/or coastal upwelling. To do this requires water mass passage through the complex zonal current systems that reside just off the equator, including the surface countercurrents (both north and south of the equator associated with the atmospheric inter-tropical convergence zones) and the subsurface countercurrents (i.e. Tsuchiya Jets). Furthermore, since the primary canonical thermocline pathways of the North and South Pacific STCs involves passage through the low latitude western boundary current system of the Pacific, the role of those currents, as well as the Indonesian Throughflow, must be properly understood. The ocean models that provide the present guidance for understanding the Pacific STCs do not properly represent the mean subsurface countercurrents, nor have they been critically scrutinized for their accurate representation of the complex western boundary current system. It is equally important to assess the accurate representation of the properties of the source waters at the subduction sites which are likely to be quite sensitive to surface fluxes and model dependent mixed layer processes. Indeed, it is the permanent subduction process that is of most interest where perhaps a low frequency version of the ‘Stommel Demon’, a name meant to represent the seasonal selection of late winter thermocline conditions for setting subducted water mass properties, might be operating. Or perhaps the occurrence of a few very strong high-frequency storms might be the primary flux ‘events’ that set the subducted water mass properties for the STCs. Both of these mechanisms are likely to compete in setting the water mass properties at the STC source regions.

Understanding the sources and pathways of the STC requires a coordinated ocean observation and modelling effort that addresses the fundamental question of what sets the strength of the Pacific STCs. We find that the mean pathways of the Pacific STCs are remarkably consistent among ocean models that are forced with realistic wind stress curl distributions. These pathways are also in accord with fundamental thermocline theory and are supported by the available observed tracer (both active and passive) records. However, the important temporal variability of the Pacific STCs has yet to be firmly established. At the source, what sets the water mass properties of the subducted waters and what is the role of strong very high frequency storm ‘events’ vs. a low-frequency version of the ‘Stommel Demon’ in the permanent subduction process? Within the thermocline, most models show a clear bifurcation of pathways along the western boundary but the fundamental processes responsible for its variability need to be understood. What sets the bifurcation latitude along the western boundary that partitions water between the subtropical gyre and the water bound for the tropics? And how is the STC partitioned between the Indonesian Throghflow and waters that make it into the tropical Pacific circulation? Once the subducted water mass makes it to the tropics it competes with local, closed tropical cells to define the SST. What is the partition of upwelled water between the STC and the local tropical cell? And what are the relative roles of the off-equatorial surface countercurrents and the equatorial undercurrent in moving water eastward and upward to produce tropical SST anomalies? Remarkably vigorous zonal currents exist in the deeper thermocline regions of the tropical Pacific. What is the role of these off-equatorial subsurface countercurrents in re-routing the water bound for the equator? Finally, the termination processes along the eastern boundary determine both equatorial and coastal upwelling and hence SST. How do the Equatorial Undercurrent, surface countercurrents and Tsuchiya Jets terminate in the eastern tropical Pacific and what is the impact of this on the coastal upwelling sites along the Peru-Chile coast and in the Costa Rica Dome?

Q3) How do surface fluxes affect subduction properties and the three-dimensional ocean circulation within the STCs?

Wind-driven Sverdrup theory and elaborations that allow for investigation of vertical structures in the thermocline in response to the combined effects of wind and buoyancy forcing provide a valuable framework for interpreting observations and model simulations. However, quantitative understanding of ocean circulation patterns and water mass properties of the STCs requires accurate knowledge of the surface forcing fields for wind stress, heat flux, and fresh water flux. From available measurements and recent theoretical developments we have gained significant insights into the basic dynamics and thermodynamics of the shallow meridional overturning cells in the Pacific. These insights have allowed us to define the broad outlines of the mean circulation patterns and water mass properties of the STCs, and how the STCs connect to the general circulation of the Pacific basin.

However, progress has been limited by lack of adequate data for accurately specifying the time mean, and the time-space varying surface fluxes, especially in the Southern Hemisphere. Model simulations are very sensitive to the specification of these forcing fields, as well as to parameterizations of mixed layer processes that affect the transfer of momentum, heat, fresh water across the air-sea interface into the ocean interior. As a result, it is not possible to unambiguously determine how momentum (wind stress) and buoyancy (heat and fresh water) fluxes affect the strength, pathways, and water mass properties of the STCs. Also, the interpretation of temperature as a dynamically active vs. passive (“spicy”) water mass tracer is complicated by present gaps in our knowledge of the surface forcing fields. The role of spatial and temporal variability in mixed layer depth and its influence on subduction processes is likewise not well understood. These uncertainties are especially problematic in attempting to determine the relative importance of surface wind stress and buoyancy forcing on decadal time scales since the relevant oceanic and marine meteorological data are spatially inhomogeneous, gappy in time, short in length relative to the time scale of interest, and often of uncertain quality.

Q4) What are the relative mean and time-variable contribution of northern and southern hemisphere STCs to the equatorial circulation?

Significant hemispheric asymmetries exist in the general circulation and water mass properties of the tropical Pacific Ocean because of hemispheric differences in atmospheric forcing and in continental land masses. The mean ITCZ is located north of the equator as a result of coupled ocean-atmosphere-land interactions. In this region, winds are weak and variable, SST and rainfall are high, surface layer salinities are low, the North Equatorial Countercurrent flows eastward against the winds, and the thermocline topography is characterized by significant shoaling in regions of wind curl-induced upwelling. In contrast, the southeastern tropical Pacific is a region characterized by relatively steady trades, cool, westward flowing currents, an excess of evaporation over precipitation, and high salinity surface waters that impart their signatures to the subducted water masses.

Models and observations of the mean thermocline circulation are in rough accord on the pathways and water mass transports between the subtropical and tropical Pacific in both hemispheres. Water mass properties and circulation patterns suggest that about 60% of the water feeding the Equatorial Undercurrent is of Southern Hemisphere origin, with the remaining 40% from the Northern Hemisphere. The pathways by which thermocline waters migrate towards the equator are much more circuitous in the northern hemisphere because the potential vorticity ridge associated with the ITCZ and NECC partially blocks direct interior communication between the tropics and subtropics. There is also rough agreement between models and observations on the magnitude of the time mean Indonesian Throughflow of about 10 Sverdrups. This transport, which exits the Pacific through straits north of the equator, must ultimately be fed by a cross-equatorial mass flux of Southern Hemisphere water.

There are a number of uncertainties though in our description and understanding of the relative contributions of Northern and Southern Hemisphere STCs to the equatorial ocean circulation in the Pacific. Western boundary current transports are less well determined than those in the interior because of the observational challenges associated with measuring swift and narrow currents that are often characterized by energetic eddies and complicated recirculation patterns. Also, though seasonal and interannual time scale variations in the shallow circulation and water mass properties are reasonably well described for the interior ocean, they are less so for western boundary currents and deeper flows such the Tsuchiya jets. There are hints of decadal time scale variability, but at present data are too sparsely distributed in space and time to clearly define these fluctuations. The relative percentage of Northern and Southern Hemisphere waters to equatorial upwelling is not well known, and the pathways by which upwelled waters make it back to subtropical subduction zones are not well understood.

Experiment Strategy

A. Models

To establish the role of the Pacific STCs in seasonal-to-centennial global climate variability requires a carefully constructed modelling experimental strategy that should include:

- analyses of archived model experiments.
- both uncoupled ocean and atmospheric model experiments (with ‘forward’ and data assimilating model experiments).
- coupled model experiments (with both reduced physics and general circulation models).

We discuss the contributions that each of these modelling categories can make in answering the four questions posed above.

A.1) Analyses of Archived Model Experiments. There exists global ocean community modelling efforts with high-enough spatial resolution to have properly resolved the major components of the Pacific STCs but have yet to be analyzed for these pathways; they should now be analyzed with the above four questions in mind. For example, the Los Alamos Parallel Ocean Program global ocean model data set is available and it would be of interest to specifically determine how well the off-equatorial surface and subsurface current systems have been modelled and, in general, how well the entire ‘plumbing’ and source water processes that defines the Pacific STCs have been represented (Questions 2 and 4). And diagnostics must be established that can cleanly separate the specific contribution of the STCs to low-frequency tropical SST variability (Question 1). Particular attention should also be paid to the time variability of these archived model data sets.

A.2) Uncoupled Ocean Model Experiments. In general, these ocean-only model experiments should be designed to elucidate the fundamental physics of the STCs and their sensitivity to various model parameterizations. Resolution should be set high enough to resolve important details of the pathways as well as the physical mechanisms at the subduction sites, perhaps with the use of advanced adaptive, nested grid algorithms. And, as is typical with a program that seeks a fundamental understanding, both simplified 'reduced physics' ocean models and ocean general circulation models have important roles to play.

For this category of uncoupled ocean model experiments, we envision both 'forward' (e.g. purely prognostic calculations without any influence from the observations) and data assimilating experiments that are separately discussed below.

A.2.1) Forward Ocean Model Experiments. An *Ocean Model Intercomparison Program (OMIP)* and a number of *Process Studies* are recommended. The OMIP should entrain out best community GCMs configured to test the sensitivity of the various model STCs to surface fluxes, mixing rates and distributions, and spatial and temporal resolution (Question 3). Three recommended Process Studies would focus upon:

- Equatorial upwelling and the relative role of the various pathways of the STC in producing tropical SST anomalies (Question 1).
- Subduction processes, with very high resolution nested grids embedded within the Pacific basin scale circulation (Question 2).
- Bifurcation processes, along both the western and eastern boundaries (Questions 2 and 4).

A.2.2) Data Assimilating Ocean Model Experiments. The implementation of an observational program designed to fully capture the complexities of the Pacific STCs will likely be incomplete (e.g. funding realities) and it is the function of this category of model experiments to make the best use of those observations that will be collected. To begin, a long-term retrospective assimilation experiment should be designed to best represent the mid-70s Pacific climate regime shift. This would accomplish a number of objectives, including a fundamental understanding of perhaps the most robust decadal variability signal available and the development of a trustworthy tool for helping to design the STC observing system. These sets of experiments would be in position to help answer all 4 questions.

A.3) Uncoupled Atmospheric Model Experiments. Our group did not discuss this category but it should be included in any reasonable experimental strategy. The experiment strategy here might parallel some of the uncoupled ocean model experiments, e.g. an AMIP and a series of Process Studies that need to be defined.

A.4) Coupled Model Experiments. Learning from our experience with ENSO modelling, we envision contributions to this category of experiments from a hierarchy of models ranging in complexity from reduced physics atmosphere *and* ocean models, to intermediate coupled models, and finally to fully coupled general circulation models (CGCMs). The simplified reduced physics models should build upon the more advanced versions of fundamental ocean thermocline theory (e.g. with an active mixed layer). This paradigm is known to produce the STCs but with *specified* SSTs. The next step would be to relax this constraint and allow the (now coupled) model to produce its own SSTs. The CGCMs, on the other hand, should be specifically designed to simulate the structure and intensity of the ITCZs in both hemispheres, which are thought to be so important in setting the STC pathways and transit times in both hemispheres. This is a difficult problem that present day coupled models have not succeeded in reproducing. Besides these Pacific Ocean/Global Atmosphere coupled models, the question of whether the subduction process itself is a *local* coupled air-sea process should be investigated, perhaps with advanced coupled nested grid systems. These sets of experiments would directly address Question 1.

B. Observations

The observational strategy to address the role of the Pacific STCs in seasonal-to-centennial global climate variability involves the collection of new measurements to improve the definition of the relevant phenomena and to improve our understanding of key physical processes. Two types of observational effort are envisioned, one involving long term, basin scale, sustained observations and one involving short duration, geographically focussed process-oriented studies to intensively address a specific set of scientific issues. The new observations will contribute directly to modelling as well as empirical studies of the STCs. The data will be available for assimilation into model-based ocean analyses of temperature, salinity and velocity, and atmospheric model-based analyses of surface fluxes. The new observations will also be valuable for model validation, and for the development of improved model parameterizations of sub-grid scale processes.

The observational strategy is based on the premise that certain critical observing systems, namely the ENSO Observing System and high-resolution VOS/XBT lines will be continued for the duration of CLIVAR. The ENSO Observing System for this purpose is taken to include the repeat hydrographic and ADCP sections from the cruises that service the moorings of the TAO/TRITON array. Our strategy also requires the continuity of satellite missions for global analyses of SST, surface wind velocity, sea level, surface insolation, penetrative radiation (ocean color) and rainfall. Finally, in addition to encouraging the collection of new observations as described below, we also endorse the continuation of data archeological efforts that add previously inaccessible historical data to the global ocean database.

Recommendations are presented separately for sustained in situ observations and process studies. The questions that these measurements address are indicated in parentheses. Implicitly, all the proposed new observations will contribute to Question 1 to the extent they facilitate model development and model studies.

B.1) Sustained In-Situ Observations

- Implementation of Argo, with initial deployments in the southeastern tropical Pacific, followed by deployments in the northeast Pacific. Floats should resolve mixed layer temperature and salinity variability (Questions 2 and 4).
- More salinity sensors deployed on platforms of the ENSO Observing System, e.g. moorings and drifters (Questions 2 and 4).
- More moored mixed layer velocity measurements as part of the ENSO observing system (Questions 2).
- Deployment of surface flux reference stations in regions of subduction and upwelling (Question 3).
- Development of techniques to monitor western boundary current transports of mass and heat (Questions 2 and 4).

B.2) Process Studies

- A field program to study mixed layer and thermocline processes that cause surface waters to be subducted into the thermocline. Regionally, this study should take place in the stratus deck region of the Northeast or Southeast Pacific, and address coupled ocean-atmosphere interactions involved in the production of subducted water masses (Questions 2 and 3).

- A field program to determine the rates, source waters, and dynamical controls on equatorial upwelling. A similar experiment for coastal upwelling would be desirable (Question 3).

- A field program to study the bifurcation of the North Equatorial Current and its time varying contribution to the Kuroshio and Mindanao Currents. Connectivity of the Indonesian Throughflow to these currents systems should be addressed as well (Questions 2 and 4).

- A field program to study the termination of major zonal currents in the thermocline along the eastern boundary, and the fate of water masses in those currents. This study should include an investigation of the equatorial 13C water and its relationship to the Tsuchiya Jets (Questions 2 and 4).

IV.II Indian Ocean Working Group

Of the three oceans considered in this workshop, least is known about STC circulation and variability in the Indian Ocean. Although the Indian Ocean has been observed and modelled during the past 40 years (see review by Schott and McCreary, 2000), our understanding of its circulations and its role in climate change remain rudimentary. The problem results from a basic lack of long-term observations there (as compared, for example, to the Pacific), a severe impediment to documenting the Indian Ocean's role in climate variability.

Q1) Do STCs play a role in seasonal to centennial climate variability, and, if so, how?

It is not known if Indian Ocean STCs affect climate variability mainly because so little is known about the fundamental structure of STCs there.

Although water is subducted almost everywhere along the southern boundary of the South Indian subtropical gyre, it is subduction in the southeastern Indian Ocean that is believed to be the source of subsurface water that flows equatorward to participate in the STCs. (Subduction in southwestern is likely confined to a westward recirculation.) There is seasonal-to-interannual SST variability in the southeastern ocean, which could be generated by oceanic dynamics in a variety of ways, and this variability could affect the STCs after it is subducted.

Another source of water for the Indian Ocean STCs is the Indoneisian Throughflow (ITF). Therefore, there is the possibility that multi-decadal STC signals could be related to the ITF, and hence forced by variations in the Pacific wind stress, rather than by regional forcing. What is the role of the ITF in the variability of the southeastern Indian Ocean, is it related to variability in the STC?

In the Arabian Sea, where most of the water in the Cross-Equatorial Cell (CEC) upwells, SST anomlaies are correlated with monsoon strength; however, the anomalies appear only to respond to the monsoon winds, and there is little evidence that they feedback to influence monsoon strength. There are also suggestions of a correlation between SST variability there and longer time scale variations, including the tropical biennial oscillation (TBO) and decadal variability. What causes the TBO? Are the ocean and STC involved? If so how?

Q2) What are the sources and pathways of STCs, including other features such as Tsuchiya Jets and the ITF?

Despite much effort in the past decades, we do not even know the fundamental structure of the circulation of subtropical-tropical cells in the Indian Ocean, including the mean circulation rate, its pathways, and where the subducted STC water is returned to the surface. Defining the fundamental structure of the STCs is a high priority. The STCs consist of subduction in the southeastern Indian Ocean (about 15-20 Sv), and upwelling in either the northern hemisphere (for the CEC) or possibly in a band from 5-10S in the central and western ocean. (Unlike the Atlantic and Pacific Oeans, in the Indian Ocean there is no strong climatological upwelling along the equator.)

In the CEC, the upwelling occurs mostly in the Arabian Sea (7-10 Sv). The upwelling along the Somali/Omani coast is South Indian and ITF water. This upwelling is also fed by a small amount of subduction within the Arabian Sea. An equatorial roll is believed to be embedded in CEC. Does the equatorial roll have a diapycnal flux? Models suggest that at least during boreal spring there may be a diapycnal flux associated with the equatorial roll. At that time of the year, the equatorial roll extends to a depth of 60 m, below the minimum mixed-layer depth. As a consequence, the upwelling arm of the equatorial roll should bring cold water into the surface mixed- layer, via cross-isopycnal flow. This should cause a net surface heat flux into the ocean in a narrow zonal band just south of the equator.

There is evidence of some STC closure by upwelling at 5-10 S (7-10 Sv) in modelling studies and CZCS data. On the other hand, there is not a strong upwelling signature in SST from satellite IR, which could be masked by clouds and humidity. The potentially strong influence of this upwelling on the atmosphere and the Findlater jet (as compared for example with the Arabian Sea coastal upwelling) results from SST in the region exceeding the critical temperature for the onset of strong atmospheric convection (about 27-28°C). Better observations of the SST variability and its causes are therefore highly recommended.

Little is known about the contribution of the ITF to STC variability, as regards subduction and upwelling, both along 5-10S and in the Arabian Sea. How does ITF affect subduction rates? What is the effect, if any, of the ITF on the atmosphere and specifically the TBO and decadal variability?

There are also questions about the larger scale connection of STCs to deeper upwelling in the Indian Ocean and to circulations in the other oceans. Concerning the conversion of cold water to warm, the magnitude of the bottom water flow into the Indian Ocean, and the effects of topographically related mixing have not been quantified. What is the variability, on all time scales, of the ITF in terms of transport and properties? In addition to the ITF, do inter-ocean exchange with the Atlantic around Africa, and with the Pacific around Australia, also affect the STCs?

Q3) How do surface fluxes affect subduction properties and the three-dimensional ocean circulation within the STCs?

Interannual variability in the surface forcing fields over the Indian Ocean has been documented. The subduction rate is controlled by the strength of the Ekman pumping rate and by mixed layer properties, such as the density and depth of the late winter mixed layer. Thus, climate variability in the South Indian Ocean is likely to have major impact on the subduction rate and water mass subducted. However, we do not have enough climate data to explore this sensitivity, and this is a critical area of focus for the study of the Indian Ocean.

Q4) What are the relative mean and time-variable contribution of northern and southern hemisphere STCs to the equatorial circulation?

Not applicable for the Indian Ocean.

Q5) How do STCs affect the mean and time-variable ocean-atmosphere tropical heat budget?

On the annual mean, there is a net heat input into the northern Indian Ocean through the surface, which must be removed by the CEC. Moreover, the contribution of the ITF to the heat flux is about equal to the total integration of the surface flux. The situation is further complicated by the strong seasonal cycle in the monsoon, which gives rise to an annual reversal in Ekman transport and cross-equatorial heat transport. Important issues in the Indian Ocean for the heat budget relate to the interannual variability in the heat balance (cross equatorial heat fluxes, air-sea heat fluxes, and heat storage, etc.), and how the STC affects the balance between these terms of the budget. How is the net southward cross-equatorial heat flux accomplished in the Indian Ocean? What is interannual variability of heat flux and heat transport? Can the Indian Ocean change the heat transport or storage on interannual time scales, and do these changes affect SST?

Implementation

The priority of international projects for the Indian Ocean should be focused on first understanding the fundamental structure of the circulation and its variability on biennial to decadal time scales. In the following, the specific issues for the Indian Ocean related to STC variability and their interaction with the atmosphere are listed. In addition, the workshop questions to which they relate are identified in parentheses (Question 4 is not relevant for the Indian Ocean). Some of the items discussed here are similar to issues raised in the "Implementation Plan for the CLIVAR Asian-Australian Monsoon Research".

A. Ocean Modelling Efforts

The recommended modelling efforts for the Indian Ocean include basic simulations of the quasi-steady state circulation, its variability over a range of time scales, the structure of the Indian-Ocean Dipole, and coupled modelling with a focus on determining how SST anomalies affect the atmosphere. Numerical simulations should be carried out within global models, to be able to investigate the role of the global thermohaline circulation and ITF within the Indian sector. A model inter-comparison study is desirable, because of the sensitivity of STC strength and structure to model parameterizations.

A.1) The quasi-steady state of the general circulation in the Indian Ocean:

1) The basic structure of the Subtropical Cell and the Cross-Equatorial Cell (Question 2): It should be emphasized that the circulation in the North Indian Ocean is dominated by a strong seasonal cycle; thus understanding STC structure is fundamentally a four-dimensional problem. Several questions need to be explored through numerical modelling and comparison with observations. These include:

What sets the strength of the STCs? What is the structure of the deep meridional cell, if it exists at all? Does the deep cell influence the STCs? What is the basic structure of the equatorial current system, including the Wyrtki Jets and equatorial undercurrents? What is the influence and fate of ITF waters?

2) Subduction and obduction (Question 2): What are the distributions and rates of subduction and obduction? What are the pathways of the subducted water masses throughout the Indian Ocean?

3) Heat transport and balance of the Indian Ocean (Question 5) Heat transport in the Indian Ocean occurs primarily via the shallow meridional overturning cells. Heat balance and storage are closely related to SST anomalies, and thus to the climate variability. Results from modelling studies should be compared with observations What are the mechanisms that affect variability in physical processes controlling the heat storage, that control change of heat storage due to warming on decadal time scale, and the potential feedback to the atmospheric circulation?

A.2) Climate variability (Question 1): An initial focus should be on shorter term climate variability, such as the TBO. How is the TBO related to anomalous external forcing, including wind stress, heat flux, and freshwater flux. Other issues that require modelling attention include the Indonesian Throughflow, interaction with the Antarctic Circumpolar Current and the Atlantic and Pacific Oceans. Exploratory work is also needed on coupled modes of variability on other time scales.

A.3) The structure of the Indian Ocean Dipole (Question 1): Simulations are needed of the Indian Ocean Dipole. How is it related to the TBO and to anomalous transport in the ITF?

B. Observations

B.1) Priorities for the Indian Ocean are to measure and observe mean, and seasonal to long term variability of:

1) ITF transport and properties.

2) Magnitude and sources of water upwelling along 5-10S.

3) Upwelling export from offshore Somalia and the northwest Arabian Sea.

4) East/west exchanges of mass, heat and freshwater across the equatorial region between 3N and 10S.

5) Pathways and southern export, specifically the southward transport through and modification within the Mozambique Channel, and southern inter-ocean exchange as part of the wider STC loop.

B.2) Recommendations for implementation of observations:

1) Implementation of ARGO float deployments as a basic requirement for large scale water mass coverage after the end of WOCE. It is recommended that some floats be parked at shallow depths (about 200 m, say), in order to drift with the subsurface STC pathways (Questions 1, 2, 3, 5).

2) Augmention of ARGO with salinity, ADCP, and high-resolution observations in key locations via volunteer observing ships and along the IX1, IX3, IX7, IX12 lines (Questions 1, 2, 3, 5). These data will contribute toward documenting interannual variability of the upwelling in the northwest Arabian Sea and along 5-10S, and the Leeuwin Current.

3) TAO Buoys between 60-90E and 3N-10S (3 in east, 3 in west, and 8 in central region). These buoys will give air-sea fluxes, winds, and subsurface instrumentation to observe equatorial upper-layer currents and cross-equatorial exchanges. Their instruments indluce Seacat T/S recorders, current meters, and at key locations also ADCPs. As part of servicing these buoys hydrography/tracer sections can be obtained seasonally. Among other things, these observations will allow calculation of a regional heat budget (Question 3 in particular, also 1, 2, and 5).

4) A process subduction experiment (will contribute to Questions 1, 2, 3, and 5) to quantify subduction rates, to understand what controls subduction rates, and to determine the relationship between STC variations and subduction rates.

5) A process experiment to study the 5-10S upwelling, because of its potential climate significance. (Questions 1, 2, 3, and 5).

C. Air-Sea Coupled Modelling Effort

One of the most important questions is, how much SST anomalies can affect the atmospheric circulation over a range of time scales.

C.1) Close examination of existing model runs: Model runs should be carefully examined for the climate variability, especially the variability of the upwelling, plus other components of the oceanic circulation system that have the potential to impact the climate.

C.2) Long-term simulation of high-resolution models are needed to provide new ideas for further study, including field observations.

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IV.III Atlantic Ocean Working Group

Knowledge of the existence of Atlantic STCs dates back to the Meteor cruises of the 1920s. By the 1970s it was apparent that the shallow circulation in the Atlantic was asymmetric, with the Southern Hemisphere providing much of the high salinity waters of the Equatorial Undercurrent. However, despite these early advances development of the connection of the STCs to climate variability is less well advanced in the Atlantic than in the Pacific and is complicated by the central role of boundary processes and the presence of multiple modes of climate variability. Recent deployment of a moored array, expansion of Lagrangian observations, and rapidly increasing theoretical and modelling experience should provide us much additional information to address these questions in the coming years.

Q1) Do STCs play a role in seasonal to centennial climate variability, and, if so, how?

Superimposed on a strong seasonal cycle there appear to be at least two additional modes of variability. The first of these, an Atlantic counterpart to ENSO, is primarily a phenomenon of the Northern Hemisphere summer and is expressed as a quasi-biennial warming of the eastern basin, a relaxation of the equatorial trade winds, modification of the African monsoon, and a south and eastward shift of convection. Modelling studies suggest that this mode is either slowly growing or damped. Interestingly, its growthrate is connected to the mean equatorial stratification, and thus to the STCs.

The second mode of variability was originally identified in conjunction with decadal fluctuations in Brazilian rainfall patterns and their connection to dipole-like meridional gradients of anomalous SST. Subsequent observational studies have shown that this pattern of SST is also related to shifts in oceanic heat storage, in the latitudinal position of the trade wind systems and the Intertropical Convergence Zone (ITCZ), as well as in the pattern of diabatic heating that drives the tropical troposphere. Modelling studies, in turn, have shown that changes in the wind field give rise to changes in latent and sensible heat flux responsible for much of the variability of extra-equatorial SST.

Even more so than the Atlantic Nino, there are ongoing debates regarding the extent to which this dipole-like interhemispheric mode is local to the tropical Atlantic and the extent to which it is connected to ENSO and to the North Atlantic Oscillation. One intriguing hypothesis suggests that the anomalous interhemispheric gradient of SST arises from fluctuations in high latitude ocean convection. We can expect rapid progress on these and related issues as attention is focused on further observational and modelling studies.

Q2) What are the sources and pathways of STCs, including other features such as the MOC?

Our current knowledge of the STCs in the Atlantic and their interactions with the Meridional Overturning Circulation (MOC) is still primitive, but improving rapidly. Our knowledge is greatest regarding the mean STCs and least regarding their potential connection to SST and to meteorological variability.

The Atlantic is distinguished by a vigorous MOC supplying in excess of 20 Sv to the deep ocean at northern latitudes. Southward deep ocean transport is balanced by a net northward mass transport of some 13-15 SV through the tropical zone at intermediate and shallow depths. At intermediate depths, low salinity Antarctic Intermediate Water penetrates into the Northern Hemisphere between sigma levels of 26.8 and 27.4. However Antarctic Intermediate Water transport is insufficient to balance southward transport at deeper levels. Thus, much of the northward transport occurs within the thermocline and mixed layer and thus is directly connected to the STCs.

The waters of the Equatorial Undercurrent (EUC) are distinguished by high salinity. The equatorial thermocline as well shows high salinity tongues penetrating equatorward from both hemispheres. Based on this and further water mass analysis in the 1970s Metcalf and Stalcup argued that the source of the southern subsurface salinity maximum and the main source for the EUC lies in the western subtropical Atlantic (15S-25S, 20W to the boundary). In this region large-scale atmospheric subsidence linked to deep convection over the Amazon results in low rainfall and high surface solar heating, and thus to high evaporation rates and high salinity.

Observational and modelling evidence suggest that western boundary processes play an important role in the northward transport of this Southern Hemisphere water. A significant portion of this water may upwell either along the equator or in the region of surface divergence associated with the ITCZ a few degrees north of the equator.

In the subsidence region of the northern subtropics high salinity water is also being formed (based on surface salinity distributions the source region is 20N to 30N, 20 to 50 W). Waters subducted here may be transported southward along the path of the North Equatorial Current. Some of this water enters the western boundary where it retroflects and becames entrained in the NECC/NEUC system of zonal currents.

Q3) How do surface fluxes affect subduction properties and the three-dimensional ocean circulation within the STCs?

Modelling and theoretical studies of thermocline ventilation provide a broad outline of the processes involved in subduction. A key aspect is the conditioning of the mixed layer due to radiative, sensible, evaporative, and freshwater fluxes as well as entrainment. Also important are the constraints associated with wind-driven Sverdrup transport and vorticity conservation.

Throughout the basin warm dry trade winds blowing off the African continent lead to high rates of evaporation. It is only in the narrow band of latitudes near the ITCZ where the freshwater supplied through rainfall exceeds that lost to evaporation. River discharge is also important in the Atlantic. During Northern Hemisphere summer and fall freshwater from the Amazon is carried eastward by the NECC, while in spring much of the Amazon water as well as that of other rivers of northern South America are carried northwestward toward the Caribbean.

The results of this complex set of freshwater fluxes are zones in the western tropics and subtropics of both hemispheres where surface salinities may exceed 37 PSU. Although these evaporative zones are important source regions for the STCs little is known yet about how they fluctuate in time and affect the supply of warm water moving equatorward.

Surface net radiation has strong seasonal variations in due to variations in clouds and solar declination. Evaporative heat loss also varies due to variations in surface wind speed. These changes in surface heat flux together with seasonal changes in entrainment are largely responsible for the seasonal variations in SST in the subtropics and in the western tropics. There is developing evidence that long decadal surface heat flux variations may be responsible for changes in tropical SST at decadal time-scales as well. Along the equator and in the east equatorial and coastal upwelling become important factors and are most closely connected to local and remote wind stress forcing.

Q4) What are the relative mean and time-variable contribution of northern and southern hemisphere STCs to the equatorial circulation?

The asymmetric nature of the mean Atlantic circulation impacts the STCs of the Northern and Southern Hemisphere. The result is to emphasize the southern STC and in the mean and to de-emphasize the northern STC. Indeed, water mass studies in the equatorial zone strongly indicate that most of the water (14 Sv) entering the Equatorial Undercurrent is of southern origin, which upwells and may then be transported into the northern gyre.

The primary interface between the northern and southern STCs appears to be the zonal NECC/NEUC current system lying between 3N-10N. To the north of this system waters of the northern STC are common, while further south they are not. This current system weakens at thermocline levels in Northern Hemisphere spring while surfaces of constant density flatten, reducing the barrier in potential vorticity and allowing for a seasonal transport of Southern Hemisphere water into the northern subtropical gyre. Some mechanisms by which this transport occurs have been identified, including the North Brazil Current and its rings. Other interior routes are possible as well. The position and strength of the NECC are directly connected to the ITCZ, thus providing a potential source of interaction between the tropical atmosphere and ocean.

Q5) How do the STCs affect the mean and time-variable ocean-atmosphere tropical heat budget?

Most of what we know about the heat budget in the tropical Atlantic is in the time mean and on seasonal timescales. In the time mean local heat storage is small compared to surface flux and divergence of heat transport. The surface heat flux is controlled by the bright solar conditions in the tropics that lead to net surface heating throughout the eastern basin at a rate of 50-70 W/m^2. This net increase in heat is reflected in the enhancement of northward heat transport in the northern tropics, which in Northern Hemisphere spring may exceed 1PW. Wind-driven mixed layer transport plays an important role in the northward movement of warm water.

The time mean heat budget is modulated by seasonally varying surface heat flux and by seasonally varying upper layer geostrophic and ageostrophic currents. During Northern Hemisphere summer and fall a wind-induced convergence of meridional heat transport causes warm water to accumulate in the zonal band between the equator and the NECC/NEUC current system. How these processes play out at interannual and decadal time-scales, the importance of slow changes in surface flux and their origins, and the implications of these processes for climate variability in the tropical Atlantic are all issues that need to be addressed through observational and modelling studies.

Experimental Strategy

A. Models

Because of the decadal and longer timescales involved in climate variability in the tropical Atlantic a complete scientific assault must rely heavily on modelling and analytic studies. Some of the questions we have identified include:

- Nature of hemispheric asymmetry (Questions 2, 4). From the discussion above it is clear that the asymmetric nature of the MOC is connected with the predominance of the Southern Hemisphere STC in supplying water to the equatorial zone. How do these processes interact? What processes regulate the mean STC pathways?

- Role of the seasonal variability (Question 1). The NECC/NEUC current system represents a significant potential vorticity barrier to meridional particle movements. This current system must also be supplied with mass from the STCs whose source regions are subjected to seasonally varying fluxes. What is the effect of the seasonal variability of the current system on the STCs and vice versa? How does seasonal variability of the STCs impact SST?

- Role of intraseasonal processes (Question 2 and 5). Does temporal variability, for example associated with the TIW, provide a significant interior pathway for water parcels? What is the fate of water carried eastward in the EUC and NECC/NEUC systems? Do these processes affect the heat budget and SST on longer timescales?

Ultimately these questions need to be addressed using coupled atmosphere/ocean models.

A.1) Analyses of Archived Model Experiments. A number of interesting model experiments have been carried out under a variety of initial and boundary conditions that have yet to be analyzed from the point of view of the tropical STCs. Similarly, the results of the current WOCE model-data synthesis efforts for the Atlantic should be specifically analyzed to assess such STC properties as pathways, strength, variability, and correlations with wind strength and SST. Analysis of models with different horizontal and vertical resolutions, vertical differencing schemes, with and without freshwater sources, and with modified MOCs offer a great deal of information about the dynamics of this system. Already a variety of studies are underway examining both Lagrangian and Eulerian statistics, as reported at this conference, and more can be expected soon.

A.2) Uncoupled Ocean Model Experiments. The characteristics of Atlantic model experiments follow those given in the Pacific. Because climate variability in the Atlantic has generally longer time scales than the Pacific simulations and assimilation studies need to span decades to capture important time scales. Also, because of the broad spatial pattern of the SST anomalies and the likely involvement of thermal fluxes, models need to span the tropics of both hemispheres and need to include nonadiabatic processes. For example, some of the more interesting CME-type experiments have a southern boundary that is place unfortunately close to the equator and thus does not include the full southern STC.

A.2.1) Ocean Model Simulations. Modelling studies can be used to gain a quantitative understanding of how subduction rates and properties (temperature, salinity, and other tracer signatures) are set by atmospheric forcing (Question 3). Further sensitivity studies with realistic models can show how controlled perturbations to climatological surface fluxes (for instance, changing the wind or heat flux over a specified region on a specified time scale) drive perturbations in the STC rates, pathways, etc. STCs can transmit extra-tropical anomalies to the equator by advecting temperature anomalies or by propagating, presumably via wave dynamics, changes in the volume transport of the STC. The importance of each of these mechanisms should be analyzed in the Atlantic context.

Interestingly, the subduction zones are generally downstream of the greatest oceanic cooling, and more work needs to be done to identify which factors actually set the properties of the subducted water. Given the atmospheric state, the quantitative modification to tropical SST brought about by the oceanic circulation (particularly the STC and the NADW cell) needs to be investigated further (Question 5).

Looking more generally at the tropical heat budget with time-varying forcing, we know that the heat storage term is negligible for sufficiently long time scales, and is of leading order for seasonal time scales, but we need models to clarify its importance on interannual to inter-decadal time scales.

A.2.2) Data Assimilating Ocean Model Experiments. Work is proceeding in this area, although it was not specifically discussed in our working group session. Data assimilation-based reanalyses offer a way to make the best use of historical observations. However, for such reanalyses to resolve important time-scales (Question 1) they must span multiple decades. In order for them to include important physics they must explicitly include freshwater cycling. Comparisons of different reanalysis products show wider differences in the Atlantic than in the Pacific, reflecting the weaker climate signals and smaller observation base.

A.3) Uncoupled Atmospheric Model Experiments. Among the issues that need to be addressed include: the mechanisms by which the tropical atmosphere responds to local anomalies and gradients of SST, the role of diabatic heating over the continents and the Amazon in particular, and the connection of atmospheric circulation in the tropical Atlantic sector to variability in the Pacific and the subtropical and midlatitude Atlantic. In particular, experiments can show how the atmospheric response to SST anomalies differs between the Atlantic and the better-studied Pacific. There is also some speculation that upwelling perturbations under the ITCZ in the Northern Hemisphere may influence the atmosphere more strongly, via the Hadley Cell, than similar perturbations to colder equatorial upwelling.

Atmospheric models can also show whether tropical Atlantic SST anomalies can influence extra-tropical phenomena such as the North Atlantic Oscillation. Previous atmospheric simulations disagree in the tropical Atlantic. These model outputs need to be further analyzed to identify key physical processes that cause this disagreement. The simple dynamical models that have proven so useful in ENSO studies are not as successful in the Atlantic because of the presence of strong continental convective zones. Thus, substantial effort needs to be directed toward more complex intermediate and full general circulation models.

A.4) Coupled Model Experiments. Ultimately, issues relating to the nature of positive and negative feedbacks between the ocean and atmosphere need to be addressed as coupled atmosphere-ocean problems. We need coupled models to understand the potential relationships between changes in surface fluxes in subduction zones, changes in the strength of the STCs, impacts on SST, and the resulting changes in surface heat, freshwater, and momentum fluxes. In particular, coupled models may be useful in determining whether the observed decadal power of tropical SST anomalies is related to STC dynamics, or if the ocean acts merely as a large heat reservoir, which reddens the spectrum of atmospherically generated variability (and how the atmosphere reacts to this reddened spectrum).

We also need coupled models to show how remote processes can drive variability within the tropical Atlantic as well as through ‘local’ feedbacks. Key examples include ENSO and the North Atlantic Oscillation (NAO). Models should be used to test the hypothesis that the NAO can serve as a two-way bridge for the tropical and extratropical North Atlantic to interact, particularly on decadal and longer time scales.

Continental processes can influence atmospheric convection, and oceanic SST variability to a greater degree in the Atlantic than the Pacific. The importance of these processes means that land surface models as well as the impact of continental runoff must be included eventually as well.

B. Observations

The observational strategy to address the role of the Atlantic STCs in seasonal-to-centennial global climate variability involves the collection of new measurements to improve the definition of the time-mean STCs and process-oriented studies to address such mechanisms as the role of surface fluxes in regulating SST. The observations will also be available for assimilation studies.

B.1) Sustained In-Situ Observations. An expansion of the current observational suite is needed. Direct measurements of velocity from Acoustic Doppler Current Profilers (ADCP) should be collected as a long term program along sections across the tropical Atlantic making use of all available ship resources (Questions 2 and 4).

- PIRATA experiment: The PIRATA experiment consists of an array of Atlas moorings (Servain *et al.*, 1998) that are serviced one or twice a year. It is recommended to collect ADCP data during those cruises along the 38W and 10W meridians.

- Ships of opportunity: An ideal transect is along the AX8 XBT. This transect is serviced by 2 commercial ships, the NOLIZWE and the NOMZI, which occupy a transect between South Africa and the East Coast of the US once every 40 to 60 days. If at least one of the two vessels could be equipped with a hull mounted ADCP, then a realization every 2 months can be obtained. Both ships are UK flagged.

- Research vessels of opportunity: Request to all research vessels that cross the tropical Atlantic periodically to collect and made available for this study meridional sections of ADCP. Examples of ships that do these crosses are European vessels serving southern ocean territories (POLARSTERN and MARION DU FRESNE).

Flux measurements in subduction zones are vital. The main subduction zones are shown schematically in Figure 1 together with upwelling zones. Flux measurements in those regions are recommended as PIRATA extensions. Finally, a node to collect and quality control these observations is recommended.

B.2) Process Studies. Even though large efforts were put into the understanding of the dynamics of one of the critical western boundary currents in the region, the North Brazil Current, very little is yet known about equatorward flows and their pathways. It is recommended that experiments be conducted in the Western Boundary Currents to determine the sources and pathways of the STCs (Q2 and Q4). Examples are as follows:

- Direct measurements of equatorward western boundary current transports and water mass properties related to the STCs. These observations can be obtained through specially designed process studies based on repeated hydrographic cruises or moored instrumentation. As part of the German CLIVAR program, an experiment is already in place off Brazil near 11S since March 2000 to observe the variability of the equatorward warm water transport. Also, a US experiment consisting of hydrographic cruises will start on March 2001 (AOML) north of Brazil between 4 and 10N.
- Lagrangian measurements in the subduction areas and western boundary current. RAFOS floats deployed in the thermocline at fixed depths (e.g. 200 m) or along isopycnals can provide information on the pathways of the subducted flow. Experiments can also be built upon augmentation of the already existing ARGO project but it must be kept in mind that profiling floats are only quasi-Lagrangian.

Deliberate tracer experiments to follow the pathways of subducted North Atlantic Water to the equator, to verify an interior pathway for the STCs as suggested by model experiments. Tracer analysis of existing data can also provide valuable information on the pathways.

To understand how STCs affect the tropical heat budget it is recommended that experiments in the subduction and upwelling zones and in the areas of greater atmospheric impact be conducted (Q1, Q3 and Q5). Examples are as follows:

Important upwelling zones are the NECC trough and the NW and SW African coasts The COSTA report recommended that ocean-atmosphere interaction buoys be deployed in these areas to quantify the fluxes of heat and momentum. The workshop reiterated this recommendation.

Process oriented experiments are to be conducted (floats, CFCs, and or deliberate tracers) to quantify and understand the processes that control subduction rates and their relationship to STCs.

B.3) Data Analysis (Q1, Q3 and Q5. It is recommended to study the subduction rates from existing CTD data. Also the properties of the subducted regions can be studied from the data to be collected by the international ARGO program.

It is recommended that analysis of all available data, including satellite data and model products, be conducted as a first step for designing fieldwork experiments (Q2 and Q3). There is a need to determine if the distribution of the subduction regions produced by the general circulation models is realistic or not. It is recommended that examinations of available model re-analyses be conducted, and that further efforts be made to produce re-analyses that can be used to examine subduction processes. Mixed layer depths and potential vorticity distributions from re-analyses would form an important constraint on numerical models. For that, it will be important that a reanalysis of all available hydrographic data be performed to obtain an improved observational field of these parameters to compare with the model products (Q1, Q3 and Q5).

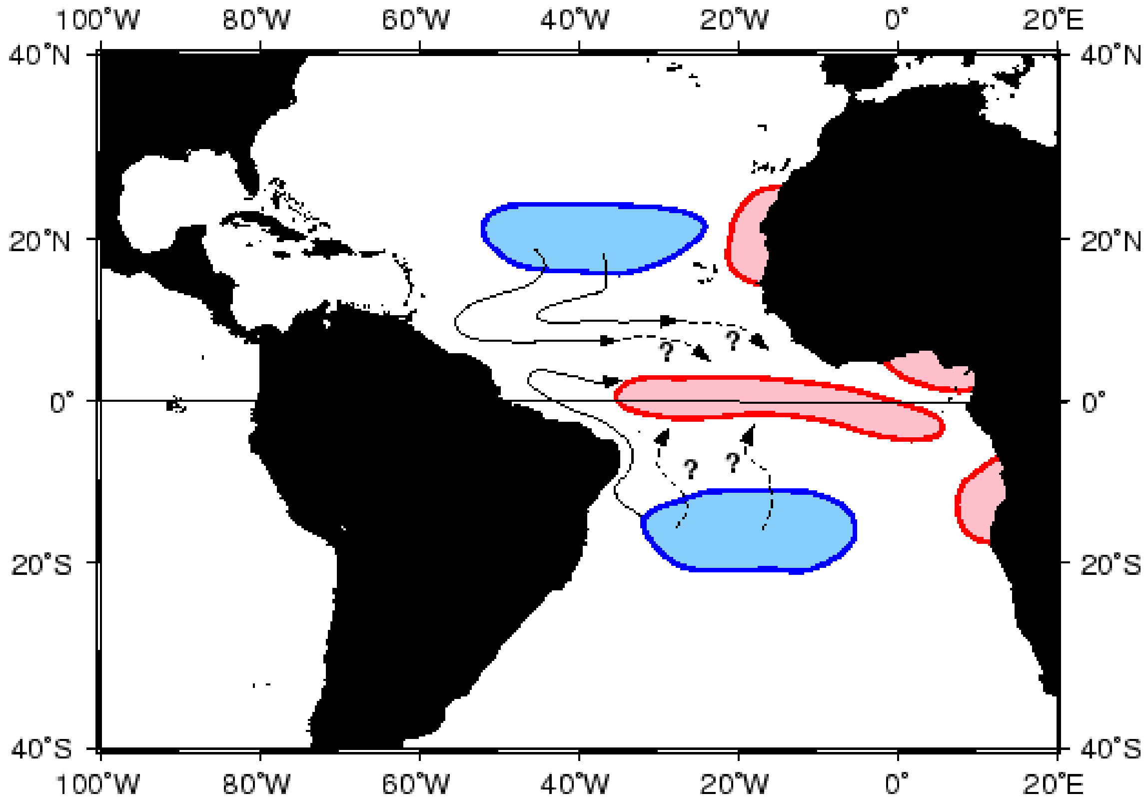

Figure 1: Schematic of the most probable locations of the subduction (blue) and upwelling (red) zones. Actual sizes and locations change depending on different data analysis or modelling results. Even though the relative contributions of northern and southern hemisphere STCs to the equatorial circulation are not well established, it is assumed that the main contribution comes from the south. Question marks indicate uncertainties on the water pathways.

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V. Overview Presentations

V.I Observational Review of the Pacific Ocean Shallow Tropical-Subtropical Overturning Cell *(Gregory C. Johnson)*

The subtropical cells (STCs) cycle water subducted in the subtropics to the equator, where it upwells and flows back poleward at the surface, exchanging heat and freshwater with the atmosphere. In the Pacific Ocean, observations are sufficient to describe a mean subtropical cell, including: subduction areas and volumes; equatorward transport pathways and partition between hemispheres, interior and western boundary; equatorial (and other) upwelling; and the surface return flow to the subtropics. However, there are still some outstanding issues even for the mean circulation. For instance, about 10 ◊ 10^6 m^3 s^{-1} of water must start from the South Pacific, cycle through the STC to cross the equator and gain North Pacific characteristics through air-sea exchange, and then flow out of the Pacific into the Indonesian Throughflow. While there are limited observations of variability of various components of the STC, the data have been taken at various times so that putting them together to assess how the system varies would be a difficult task. Observing how the entire system interacts with the atmosphere and exploring possible modes of coupling over longer time-scales will be a significant challenge.

Introduction

The Subtropical Cell (STC) is here loosely defined as the part of the ocean circulation which brings water subducted at higher latitudes to the equatorial band, effects the near-equatorial sea-surface temperature (SST) by upwelling and diapycnal mixing, and then moves water back poleward at the surface, where it can exchange heat and freshwater with the atmosphere. Model results suggest STCs play a potential role in modulating climate through equatorial SST either by advecting water subducted with anomalous temperature and/or salinity in the subtropics to the equator where it upwells [Gu and Philander,1997], or by varying the amount of subtropical water adverted to the equator [Kleeman *et al.*, 1999] which may in turn modulate processes such as ENSO that control equatorial SST. It should be kept in mind that the STCs are not the only candidates for decadal modulation of ENSO dynamics and predictabililty [Kirtman and Shopf, 1998]. In addition the STCs are not closed circulations, but play a role in the global ocean circulation. For instance, in the Pacific Ocean, about 10 ◊ 10^6 m^3 s^{-1} of water from the southern hemisphere must cross the equator to feed the Indonesian throughflow, and this water is cycled through the Pacific STC on its way north [Wijffels, 1993].

This review will focus on observations relevant to the STC in the Pacific Basin. Of the three oceans, the STC is perhaps best defined in the Pacific. The mean pathways of the Pacific STC are fairly clear from published observations. As in the other oceans, variability of the Pacific STC is difficult to study from observations alone. However, in the Pacific some tantalizing observational evidence on STC variability has been found. Here the STCs will be followed in four limbs: (1) their origin at extratropical subduction regions, (2) their equatorward progress in the pycnocline, both at the western boundary and in the interior, (3) their upwelling sites on and off the equator, and (4) their poleward return at the surface. Observations relevant to each of these limbs will be discussed first in terms of an elusive and perhaps misleading “mean” state of the STC, and then (when available) in terms of the variability about that mean.

Subduction

Subduction refers to the process by which surface water, under the influences of seasonally varying buoyancy and momentum fluxes, moves from the seasonal pycnocline into the permanent pycnocline, where it is shielded from direct air-sea interaction. Subduction is picked as the starting point for the STC discussion because it is where water within the STC first acquires either anomalous temperature-salinity properties or anomalous formation rates. Analysis of climatological winds and ocean structure suggests subduction rates are fairly high in a band across the northern subtropical Pacific, but highest in the west [Huang and Qiu, 1994]. These high subduction rates lead to an enhanced presence of certain density classes within the pycnocline, hence reduced stratification, which are known as “mode waters”. In the North Pacific, there are three subtropical modes waters of increasing density from west to east [Ladd and Thompson, 2000]. The most eastward and equatorward of these mode waters are the ones which flow westward and equatorward to participate in the STC, rather than eventually turning north at the western boundary to join the subtropical gyre [Gu and Philander, 1997]. In the South Pacific, the highest subduction rates are equatorward and eastward [Huang and Qiu, 1998].

Estimating subduction rates is an involved process that requires interior ocean data and air-sea flux data, and these estimates have been limited to “means”. However, the atmospheric forcing can clearly change the rates of subduction or the properties of the subducted waters, and there have been some attempts to examine changes in the properties of the subducted waters from observations. In the North Pacific, one manifestation of the Pacific Decadal Oscillation [Mantua *et al.*, 1997] is a contrast between anomalously cold SSTs in the center of the Basin with anomalously warm SSTs around the eastern boundary, or vice versa. Analysis of subsurface temperature data suggests these anomalous SSTs appear to propagate into the ocean interior to slowly cool or warm it as water is subducted in the winter season, moving equatorward as the move into the pycnocline [Deser *et al.*, 1996]. However, without salinity data it is difficult to determine whether these warm and cold anomalies are the result of changes in the temperature-salinity (T-S) properties of subducted water or changes in the pycnocline depth owing to wind forcing changes. Another study analyzed the northern hemisphere propagation of isopycnal thickness anomalies (assuming a climatological T-S relationship towards the equator [Schneider *et al.*, 1999]. This study, using observing winds to force an ocean model, concluded that the anomalies may have been advected by the mean circulation as far south as

18°N, but equatorward of that latitude anomalies were the result of changes in the pycnocline forced by local wind variability.

Considerably less work has been done on variability of air-sea fluxes and subduction in the southern hemisphere. This lack of work is due in part to the relative dearth of oceanic and atmospheric observations there, and in part to the opinion that the region exhibits little low-frequency variability. However, a meteorological reanalysis product does show low-frequency variability in the Southern Hemisphere of character and magnitude similar to that seen in the north [Garreaud and Battisti, 1999]. One of the few studies of oceanic variability focuses on the T-S properties on the core isopyncal of the salty tongue of subtropical water that reaches the equator to feed the equatorial undercurrent [Kessler, 1999]. This study shows considerable interannual variability along 5-10°S. The interannual variability was largely attributed to advection and changes in the South Equatorial Current associated with El Niño. However, there was also and interdecadal trend over the record length from 1983-1997, with a tendency towards a warmer, saltier tongue. Kessler [1999] could not find a satisfactory explanation of this trend.

Equatorward Pycnocline Flow

Equatorward flow within the pycnocline occurs in broad flows in both the ocean interior and along the boundaries in western boundary currents. In the Pacific, the western boundary transports toward the equator seem to be of roundly equal magnitude each hemisphere [Butt and Lindstrom, 1994], however, the interior flow in the southern hemisphere is much large than that in the northern hemisphere [Johnson and McPhaden, 1999]. This imbalance in the STC interior equatorward flow could provide the mass to ultimately feeds the Indonesian throughflow.

Interior Flow. The location of the InterTropical Convergence Zone (ITCZ) in the northern hemisphere makes the interior flow there considerably weaker than that in the southern hemisphere. However, attention has been drawn to the northern hemisphere interior pathway by tritium data [Fine *et al.*, 1987]. The wind stress field associated with the ITCZ creates high potential vorticity within the pycnocline which pushes equatorward flow in the North Pacific toward the western boundary [McCreary and Lu, 1994]. Both the tracer data, (transient [Fine *et al.*, 1987] and traditional [Tsuchiya, 1968]) and the wind stress data [McPhaden and Fine, 1988] suggest there is some interior flow to the equator in the northern hemisphere. The tracer data additionally give a decadal time-scale for this northern interior route. However, the wind stress data clearly show that the southern interior flow to the equator is much larger, more direct, and possibly faster than the northern.

The results from the wind stress data are confirmed by a recent analysis of historical CTD data, where salinity, acceleration potential, and potential vorticity distributions all suggest a weak and circuitous northern hemisphere interior pathway with southward interior flow having a strong westward flow in the North Equatorial Current (NEC) north of the ITCZ, shifting to eastward in the North Equatorial CounterCurrent (NECC), and then westward again in the northern branch of the South Equatorial Current (SEC). By contrast, the southern hemisphere these properties all show a direct route northwestward in the southern branch of the SEC toward the equator in the ocean interior. The relative contributions of the equatoward interior geostrophic flows are estimated from the historical CTD data to be 5 (±1) ◊ 10^6 m^3 s^{-1} in the north and 15 (±1) ◊ 10^6 m^3 s^{-1} in the south [Johnson and McPhaden, 1999].

Western Boundary Currents. The other pathway for subtropical water to flow to the equator is along the western boundary. In the Pacific, the topography of the western boundary is quite complicated. In the south there are numerous island chains and passages among them that allow multiple western boundary currents [Butt and Lindstrom, 1994]. Assessing the magnitudes of these various boundary currents, much less their variability, is a difficult task. In the north, there is only one western boundary current, the Mindanao Current (MC), which flows south towards the equator along the Philippines. However, the major egress for the Indonesian ThroughFlow (ITF), a substantial amount of Pacific water which flows to the Indian Ocean through the Indonesian Archipelago, is into the Celebes Basin though a passage located just south of Mindanao at around 5°N. The ITF drains water from the MC just as it is turning east to feed the NECC, and eventually the EUC. The historical geopolitical difficulty in measuring the ITF has made estimating the contribution of the MC to the equatorial Pacific difficult. In addition there are the complicating influences of the Mindanao and Halmahera eddies, offshore recirculations that make determining the net western boundary current transports more difficult.

A fair amount of effort has been put into measuring the western boundary currents in the low-latitude South Pacific. A major passage for the primary western boundary current, the New Guinea Coast Undercurrent (NGCU), is the Vitiaz Strait between Papua New Guinea and New Britain. Two shipboard CTD/ADCP surveys have been combined with limited moored current data to estimate a transport of 8-14 ◊ 10^6 m^3 s^{-1} between a depth range of 20-300 m, through Vitiaz Strait [Lindstrom *et al.*, 1990], with a smaller 1-4 ◊ 10^6 m^3 s^{-1} passing through St. Georges Channel between New Britain and New Ireland in the same depth range. In addition, a CTD/ADCP survey off the east coast of New Ireland revealed another, smaller western boundary current, the New Ireland Coastal Undercurrent (NICU), carrying about 5 ◊ 10^6 m^3 s^{-1} of water within the density range feeding the EUC.

An analysis of eight CTD/ADCP surveys off Mindanao along 8°N [Wijffels *et al.*, 1995] results in a southward transport within the MC of 23 (±4) ◊ 10^6 m^3 s^{-1} above = 26.7 kg m^{-3} (or around 350 m). However, as much as 9 (±3) ◊ 10^6 m^3 s^{-1} of this water exits the Pacific into the Celebes Basin to feed the ITF [Gordon *et al.*, 1999], leaving perhaps as much as 14 (±5) ◊ 10^6 m^3 s^{-1} to reach the equatorial Pacific from the MC. However, it seems likely that not all of this water reaches the EUC, with some of the denser MC waters feeding a portion the Northern Suburface Countercurrent (NSCC), which itself carries about 7 ◊ 10^6 m^3 s^{-1}, and some of the lighter MC waters being too light to supply the EUC.

While the low-latitude WBCs in the south total more when larger density ranges are considered, the amount of water estimated to actually supply the EUC (24.5 kg m^{-3} < <26.5 kg m^{-3}) from southern hemisphere low-latitude WBCs is about 9 - 12 ◊ 10^6 m^3 s^{-1} [Butt and Lindstrom, 1994], which leaves from 13 -16 ◊ 10^6 m^3 s^{-1} to of northern hemisphere water to make up the difference for an EUC of 25 ◊ 10^6 m^3 s^{-1} at around 153°E. These numbers are in accord with the difference of the MC and ITF transport estimates given above. Thus, while none of these estimates are from long-term means, they do tend to hang together within their error bars, suggesting that roughly equal amounts of southern and northern hemisphere waters feed the EUC via the low-latitude WBCs.

Upwelling

The next component of the STC is upwelling, where the relatively cold water subducted in the subtropics reaches the ocean surface, effecting large heat fluxes and strongly influencing the atmospheric circulation. This majority of this upwelling takes place along the equator, where the EUC shoals to the east under the influence of the easterly trade winds, upwelling owing to the equatorial divergence caused by poleward Ekman fluxes. However, there is also some water from the eastern end of the EUC, and perhaps the Southern Subsurface Countercurrent (SSCC), that upwells along the coast of Peru. This coastal upwelling and advection of cold water offshore is certainly associated with the stratus cloud decks that exist off Peru, although a causal relationship either way is not established. Finally, there is some potential for upwelling in the Costa Rica Dome and under the ITCZ, where thermocline and even sub-thermocline water is brought very close to the surface.

Equatorial Upwelling. Instantaneous vertical (or for that matter diapyncal) velocities are very small even on the equator, and mean vertical velocities are a few orders of magnitude smaller. Thus, vertical velocities are just about impossible to measure directly. Observational estimates of vertical velocity mostly come from continuity, assuming zero vertical velocity at the surface, and integrating horizontal divergence vertically. These calculation are inherently noisy, since they involve taking derivatives to estimate the divergence.

An early calculation of equatorial upwelling through 50 m was made over a fairly large area, between 170°E and 100°W, and 5°N and 5°S [Wyrtki, 1981]. The upper-ocean zonal transports and their vertical distribution with respect to 50 m were estimated at the end longitudes of the box from the literature. Dynamic height differences across end latitudes were used to get a equatorward geostrophic velocities along the northern and southern boundaries of the box. A wind stress climatology was used to estimate the poleward Ekman transports across these same boundaries, which was applied above 50 m. There was a range of estimates involving different zonal convergence and divergences. That most favored by the author involved 51 ◊ 10^6 m^3 s^{-1} of water upwelling across 50 m within the box. The large latitudinal range was required to perform the geostrophic and Ekman calculations sufficiently far from the equator (±5°) to avoid complications from higher order dynamics.

A more detailed diagnostic model based on the same dynamics [Bryden and Brady, 1985], but this time limited to between 150°W and 110°W, gave an upwelling transport 22 ◊ 10^6 m^3 s^{-1} across 62.5 m, again between 5°N and 5°S. These authors pointed out that the thermocline shoals up to the east, and the EUC generally follows that trend. An analysis of flow across isotherms, rather than isobars, showed that only 7 ◊ 10^6 m^3 s^{-1} actually crossed the 23°C isothermal surface. that because the thermocline shoals up toward the east. Tracers, such as ^{14}C and CO_2, incorporated in a box model [Quay *et al.*, 1983], set a lower limit for water feeding the upwelling of 225 m, or = 26.5 kg m^{-3}. An analysis of tropical Pacific ocean tritium distributions [Fine *et al.*, 1983], set a lower limit for the upwelling at about 16°C.

More direct measurements of equatorial upwelling have been mostly limited to small-scale mooring arrays. These arrays are used to estimate a vertical profile of local horizontal divergence. This profile is then integrated downward from the surface, assuming no flow through the surface, to infer the vertical velocity. A recent example is a year-long deployment of a 4° longitude by 2° latitude diamond-shaped array at 140°W [Weisberg and Qiao, 2000]. Analysis of data from this array gave a maximum mean upwelling velocity of 2.3 ◊ 10-5 m s^{-1} at 50 m, and showed that synoptic vertical velocities associated with tropical instability waves could be an order of magnitude larger than the mean. Surface drifters also give an estimate of equatorial upwelling [Poulain, 1993; Johnson, 2001], although these instruments do not stay long in this area of large divergence.

Shipboard ADCP data give a larger-scale means of estimating vertical velocity by integrating horizontal divergence. Analysis of 27 sections from the Hawaii-Tahiti shuttle experiment in the central Pacific [Johnson and Luther, 1994] suggests a mean structure of vertical velocity with upwelling centered at 2°N with downwelling at 2°S and 6°N. A more recent analysis of 85 meridional sections taken over the 1990s from 95°W to 170°W, mostly during maintenance of the TAO mooring array, shows poleward near-surface Ekman flow and equatorward subsurface flow. This structure results in upwelling centered on the equator, with downwelling in the NECC at about 7°N [Johnson *et al.*, 2001]. Both of these analyses are plagued by noise from tropical instability waves, which have meridional velocities about an order of magnitude larger than the means. In addition to showing equatorial divergence, drifter data suggest surface convergence, hence downwelling, around ±4° from the equator [Johnson, 2001]. This evidence of off-equatorial downwelling is reminiscent of the shallow “tropical cells” predicted by models [Lu *et al.*, 1998]. These cells recirculate water downwelled off the equator through the EUC.

Seasonal to interannual variability of equatorial upwelling have been examined using TAO mooring data in conjunction with various wind analyses and shipboard data [Meinen *et al.*, 2001]. This basin-scale analysis uses geostrophic and Ekman dynamics to the TAO mooring array and is again made between 5°S and 5°N. Over seasonal time-scales the zonal geostrophic divergence and Ekman divergence are the dominant terms. During the 1997-1998 El Nino, the meridional geostrophic convergence is nearly eliminated as the thermocline flattens and the Ekman divergence is reduced even more, halving the upwelling across 50 m.

Peru-Chile Coastal Upwelling. What remains of the EUC at the Galapagos Islands turns southeastward towards the west coast of South America [Lukas, 1986], and feeds the poleward flowing Peru-Chile Undercurrent. The transport of this current at 10°S has been estimated from several hydrographic sections as $1 \lozenge 10^{6}$ m^{3} s^{-1}, sufficient to supply coastal upwelling to at least 15°S [Huyer *et al.*, 1991]. An older mass budget calculation based on a singly hydrographic survey suggests requires about that much upwelling through the 100 m from 6°S to 15°S within about 200 km of the coast [Wyrtki, 1963]. These numbers are quite small in terms of the equatorial upwelling and the mass budget of the STCs, it is important to note that the cold, nutrient-rich water upwelled off the west coast of South America has a significant effect on local climate and fisheries.

Costa Rica Dome Upwelling. A final possible location for upwelling is the Costa Rica Dome, where the top of an extremely sharp and shallow thermocline is often within 10 m of the sea surface [Wyrtki, 1964]. Surface waters around the dome, near 89°W, 9°N, are slightly cooler and saltier than surrounding waters. In addition, these waters are undersaturated in oxygen and relatively nutrient-rich. All of these hydrographic signatures suggest some upwelling is present in the dome. A very crude estimate of upwelling implied about 10^{5} m^{3} s^{-1}, two or three orders of magnitude less than on the equator [Wyrtki, 1964]. Even the implied vertical velocity over the 200 to 300-km dome dimensions, about 10^{-6} m s^{-1}, is an order of magnitude smaller than equatorial values. Again, upwelling of this magnitude is small in terms of the mass budget of the STCs, but it could be locally important for fisheries in bringing nutrients to the surface. In addition, it must be kept in mind that only very small vertical velocities are needed to force significant horizontal geostrophic velocities. Johnson and McPhaden [1999] speculate that Costa Rica Dome upwelling might be associated with northward flow of the NSCC.

Poleward Surface Flow

Since most of the upwelling limb of the STCs occurs on the equator, with only small contributions through coastal and other upwelling, it follows that most of the surface return flow must originate from the equator and flow poleward. Surface drifters clearly show the poleward component of Ekman flow driven by the easterly trade winds in both hemispheres, both at large scale [Reverdin *et al.*, 1994], and even with tens of km from the equator [Poulain, 1993]. Seasonal variations in near- surface equatorial divergence are in phase with changes in the zonal winds [Poulain, 1993]. These poleward flows are superimposed on the geostrophic currents, which are predominantly zonal and, except for the NECC and the weaker SECC, westward. Shipboard ADCP data give some indication of the vertical distribution of poleward Ekman flow, and it appears that these flows reach below the mixed layer [Johnson *et al.*, 2001].

Little has been done observationally to quantify the time-scales and pathways of these surface flows. Time scales estimated from surface drifter data for surface flows to reach the subtropics from the equator are of the order of two years [Johnson, 2001], somewhat faster than the decadal time-scales for subsurface flows from the subtropics to reach the equator. The pathways tend to result in this surface flow reaching the subtropics in the western half of the basin [Johnson, 2001]. Since subduction of water taking part in the STCs occurs in the east, it would seem highly probable that the poleward surface flow and this subduction are not connected. That is to say, the STCs are not closed circulations, but are liked to the global ocean circulation.

Another link to the global circulation for the Pacific STC is the Indonesian throughflow. More thermocline water reaches the equator from the southern hemisphere than the northern. This water must modify its potential vorticity from negative to positive to reach the throughflow in the northern hemisphere. There are two possible routes. This first is a western boundary current, where lateral friction allows modification of potential vorticity. The second is the sea surface, where surface forcing can play a role in changing the sign of potential vorticity. SADCP velocity data suggest a larger poleward near-surface transport in the northern interior than in the southern [Johnson *et al.*, 2001], but this difference is not significant within the noise. The actual route for this cross-equatorial flow is unclear.

Discussion

The mean Pacific STC is fairly well defined. Extratropical subduction that feeds the STCs is stronger in the southeast than the northeast Pacific [Huang and Qiu, 1994; 1998]. There is evidence of variations in these subducted waters in both the northern [Deser *et al.*, 1996] and southern [Kessler, 1999] hemispheres. The equatorward thermocline flow toward the equator in low-latitude western boundary currents is roughly equal in magnitude [Butt and Lindstrom, 1994]. The equatorward thermocline flow in the ocean interior is much larger and direct in the south than in the north [Johnson and McPhaden, 1999]. Equatorial upwelling brings much of this equatorward thermocline flow to the surface in the EUC, with most of of this water following isopyncals as they surface in the eastern equatorial Pacific, and the remainder mixing across them [Bryden and Brady, 1985]. As expected, there are significant variations in the strength of the equatorial upwelling on seasonal to interannual time-scales [Meinen *et al.*, 2001]. Surface flow takes this upwelled water back to the subtropics in the western half of the basin in both hemispheres [Reverdin *et al.*, 1994; Johnson, 2001]. While it is evident that the thermocline flow of Southern Hemisphere water is greater than that to the north, it is not clear how this water makes its way north from the equator to feed the Indonesian Throughflow. While the water feeding the throughflow must ultimately come from the south, water properties do suggest that it bears the imprint of North Pacific surface forcing [Ffield and Gordon, 1992].

As of yet there is only tantalizing observational evidence of STC variability in some of the limbs, most notably the properties of the equatorward thermocline flow and the strength of the equatorial upwelling. The STCs clearly respond to atmospheric forcing. There is obvious potential for the atmosphere to respond to changes brought about by modulation of the Pacific STC, especially through changes in equatorial sea-surface temperature or the dynamics of El Niño. Observations of a coupled interaction will require a basin-wide effort that persists in the ocean and atmosphere for at least a decade.

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V.II Shallow Overturning Cells in the Pacific Ocean (*Julian P. McCreary, Jr.*)

During the past decade, it has become clear that the equatorial Pacific Ocean is not dynamically isolated, but rather is an essential part of the Pacific general ocean circulation. Indeed, the near-surface equatorial currents are branches of shallow, overturning cells that extend to mid and higher latitudes. These cells include the North and South Subtropical Cells (STCs), in which water flows out of the tropics in the surface layer, subducts in the subtropics, returns to the tropics within the thermocline, flows eastward in the Equatorial Undercurrent (EUC), and upwells in the eastern equatorial ocean. They are climatically important because they carry cool subsurface water into the equatorial region, and hence are an essential part of the equatorial heat balance. (Somewhat deeper cells involve the North and South Subsurface Countercurrents, also referred to as Tsuchiya Jets and their counterparts in the Atlantic Ocean are called the North and South Equatorial Undercurrents.) This talk focuses on the modelling studies that have helped to provide a more complete picture of the STCs, discussing their three-dimensional structure, dynamics, variability, and possible influence in coupled models.

1. Structure

A number of theoretical studies using simple ocean models ($2\frac{1}{2}$-layer systems) and idealized forcings have investigated the pathways by which subtropical, thermocline water flows to the equator to join the EUC. Pedlosky (1987, 1988) and Pedlosky and Samelson (1989) obtained solutions by matching the geostrophic interior flow field of Luyten *et al.* (1983) to an equatorial, inertial boundary layer, thereby explicitly pointing out the close connection that exists between the subtropical and equatorial oceans. McCreary and Lu (1994) extended this work by finding solutions in a closed basin, showing that the subsurface, equatorward flow was one branch of a closed STC.

Forcing their $2\frac{1}{2}$-layer model with realistic winds, Lu and McCreary (1995) determined further that most of the subtropical water flows to the equator via tropical western-boundary currents (the “western-boundary” pathway), rather than in the interior ocean (the “interior” pathway). Indeed, in the North Pacific region of their solution *all* of the subtropical water followed the western-boundary pathway because of a “potential vorticity barrier” generated by Ekman pumping associated with the ITCZ. This property is consistent with observed tracer distributions, which also appear to be blocked near the latitude of the ITCZ at 10°N (see Wyrtki and Kilonsky, 1984, for example).

Recently, Lu *et al.* (1998) described the STC pathways in a solution to their $3\frac{1}{2}$-layer model driven by annual-mean Hellerman and Rosenstein (1983) winds. For simplicity, the following discussion describes only the solution’s more complex, northern STC. (The southern STC is similar but simpler, because there is no analog to the North Pacific subpolar region in the southern hemisphere.) The fate of the water that subducts in the northern subtropics depends on where it subducts. Water that subducts too far to the west recirculates within the Subtropical Gyre, so that only the water that subducts in the eastern subtropics participates in the STC. Consistent with the Lu and McCreary (1995) solution, most of this water flows across the interior of the tropical ocean and moves to the equator via a western-boundary current, where it joins the EUC and eventually upwells into layer 1 in the eastern, equatorial Pacific. A small portion, however, also follows an interior pathway. After upwelling at the equator, some layer-1 water flows directly from the tropics into the eastern subtropics where it subducts and returns directly to the equator, thereby providing a *direct* closure for the northern STC. However, much of the water that subducts in the northeastern subtropics originates in the subpolar ocean, and there is no direct pathway by which layer-1 water can flow from the tropics into the subpolar region. Instead, to reach the subpolar ocean it first subducts into layer 2 in the central or western subtropics, recirculates in the northern Subtropical Gyre, and enters the subpolar ocean in the western-boundary current. There, it upwells into layer 1 via Ekman suction, flows southward into eastern subtropics via Ekman drift, where it can finally subduct into layer 2 to close the STC. (This second STC pathway actually defines another shallow overturning cell, the Subpolar Cell. This cell connects the subtropical and subpolar regions of the North Pacific, and as discussed above is tightly linked to the North Pacific STC.) Finally, a third indirect pathway bypasses the subpolar region altogether; in this pathway, layer-1 water subducts in central or western subtropics, circulates about the Subtropical Gyre near its outer edge, and thereafter moves into the tropics.

Solutions to general circulation models (GCMs) also develop STCs (Bryan, 1991, see his Figure 2; Liu *et al.*, 1995; Blanke and Raynaud, 1996; Rothstein *et al.*, 1998; Huang and Liu, 1999). The GCM solutions tend to allow more flow via the interior pathway than the layer models do. The reason for this different behavior is likely that the latter have stronger vertical mixing, which can break the potential-vorticity constraint. Alternately, it may result from the GCMs having different background stratifications that can alter characteristic pathways (see below) or higher vertical resolution. The existence of this pathway is supported by their being a tritium maximum along the equator in the central Pacific (Fine *et al.*, 1981, 1987), which suggests that some high-tritium, subtropical water converges onto the equator in that region (McPhaden and Fine, 1988).

In the GCM solutions of Rothstein *et al.* (1998) and Huang and Liu (1999), some of the subsurface water that takes the western-boundary pathway retroflects in the western ocean to join a deep portion of the North Equatorial Countercurrent (NECC). It flows westward in the NECC and only flows to the equator in the central ocean. This interesting pathway is not present in the Lu *et al.* (1998) solution for reasons that are clear. It provides another means for explaining the central Pacific tritium maximum noted above.

2. Influence of Indonesian Throughflow

Significant water exchange between the Pacific and Indian Oceans occurs in the Indonesian Archipelago, where the Indonesian Throughflow allows an outflow of upper-ocean water estimated to be 5-15 Sv (Fine, 1985; Gordon, 1986; Godfrey, 1989; Hirst and Godfrey, 1993). It is balanced by an inflow of intermediate waters from the Antarctic Circumpolar Current (ACC; Reid, 1997). We define the Pacific Interocean Circulation (IOC) to be all the pathways that connect this southern inflow to the Indonesian outflow. Defined in this way, the IOC is the Pacific branch of the hypothesized global "conveyor belt" (Gordon, 1986).

Influences of the Pacific IOC in the equatorial ocean have been discussed in several recent modelling studies. For example, Semtner and Chervin (1992) noted that 4-10 Sv of deep IOC water mixes upward into the equatorial thermocline in their solution. Blanke and Raynaud (1996) reported that IOC water flows in their model EUC, and as a result $\frac{2}{3}$ of the extratropical water in the EUC is of southern-hemisphere origin. Shriver and Hurlburt (1996) outline the IOC pathways present in a solution to their 6-layer GCM. In all of the pathways, IOC water enters the Pacific in the southeastern basin, and eventually moves to the equator near the western boundary. In one of the pathways, IOC water within layer 3 is then allowed to flow directly out of the basin in the Throughflow. In another, IOC water within layer 2 flows eastward in the EUC, upwells into layer 1 in the eastern, equatorial basin, flows into the northern subtropical ocean where it subducts back into layer 2, and only then leaves the basin in the Throughflow. Lu *et al.* (1998) simulated the IOC by specifying an inflow through the southern boundary of their basin and an outflow through an idealized representation of the Indonesian passages north of the equator. They found pathways similar to those in the Shriver and Hurlburt (1996) solution, and an additional one involving upwelling in the subpolar region (as discussed above). They also concluded that $\frac{2}{3}$ of the EUC extratropical water was of southern-hemisphere origin because of the IOC.

The IOC thus forces a north-south asymmetry to the Pacific equatorial circulation. It intensifies the amount of southern-hemisphere water that upwells along the equator, and strengthens the southern STC relative to the northern one. As such, the IOC plays an analogous role to the deep thermocline cell in the Atlantic.

3. Dynamics

3.1 Subtropical intrusions: A key property of many (if not all) of the above studies is the existence of a Shadow-Zone characteristic (Luyten *et al.*, 19xx) that extends equatorward from a point at the eastern boundary of the basin x_e and at the ventilation (or "subduction cutoff") latitude y_d. A typical value for y_d is 15-20° north and south of the equator. The layer-2 flow to the east of this characteristic is at rest (*i.e.*, the Shadow Zone). More interesting for our purposes, though, is that subtropical water "intrudes" into the tropical ocean west of the streamline. The intrusion does *not* occur in a linear model, resulting from nonlinearities in the divergence term of the continuity equation. Although strictly incorrect (due to the failure of geostrophy), the characteristic can be extended to the equator. It intersects at a distance L from the eastern boundary given by

$$L = \frac{1}{2}\frac{g_{12}'}{\left|\overline{\tau^x}\right|}(H_2 + 2H_1) \qquad (1)$$

where g_{12}' is the reduced gravity coefficient between layers 1 and 2, H_1 and H_1 are the thicknesses of layers 1 and 2 at the eastern boundary of the basin, and $\overline{\tau^x}$ is the average zonal wind stress along L. Equation (1) provides insight into whether an interior pathway exists: If L is larger than the width of the basin, then the Shadow-Zone characteristic intersects the western boundary before intersecting the equator and there is no interior pathway. Note that L depends on the stratification parameters g'_{12} and H_2, and that an interior pathway is more possible when these parameters are smaller. This implies that an interior pathway will exist at shallower levels in the real ocean, consistent with models and observations. Also note that L is smaller when $\overline{\tau^x}$ is larger, a consequence of the system then having a larger near-equatorial interface tilt, and the system therefore being more nonlinear.

3.2 Strength: The basic reason for the existence of STCs is that there is a net divergence of upper-layer water from the tropical ocean: This poleward volume transport in layer 1 must be compensated for by an equatorward volume transport in layer 2. McCreary and Lu (1994) derived a simple expression for the upper-layer divergence in their analytic model [see their equation (17)], showing that it was determined mostly by the Ekman transport across the tropical boundaries

$$E(y_d) = \int_{x_w}^{x_e} \tau^x(x, y_d) \; / f(y_d) d_x \tag{2}$$

where y_d is the subduction cutoff latitude north or south of the equator, and x_w is the longitude of the western boundary.

According to (2), STC strength is *not* determined by midlatitude wind-stress curl, as might be expected. It is true that an increase in midlatitude wind curl increases subduction by a proportionate amount, but the additional subducted water can recirculate within the Subtropical Gyres, and need not flow into the tropical ocean. It will do so only if the wind-curl anomaly also acts to increase the Ekman divergence out of the tropics.

4. Variability

The work summarized above only discusses steady-state solutions for the STCs and their dynamics. There have been almost no studies done to understand the processes that govern STC variability. Variability of the STCs has been proposed to be important for climate variability in two different ways. In one first scenario anomalous temperatures are advected by the mean STC (labelled $\bar{v}T'$), and in the other the mean temperature field is advected by variations in STC strength ($v'\bar{T}$).

Gu and Philander (1997) first proposed the former idea ($\bar{v}T'$), hypothesizing that midlatitude SST anomalies generated by surface heat fluxes could be subducted into the subsurface branch of the STCs, advected to the equator, and upwelled there to affect the Pacific cold tongue. A number of modelling studies have explored this idea. They suggest that midlatitude SST anomalies are weakened by a variety of processes by the time they reach the equator, and hence that their effect on equatorial SST anomalies is quite small (*e.g.*, Xie and Nonaka, 2000).

Kleeman *et al.* (1999) first proposed the latter idea ($v'\bar{T}$), showing that changes in the wind stress along y_d altered the strength of the North Pacific STC in their coupled, ocean-atmosphere model, leading to changes in the size and strength of the cold tongue. Klinger *et al.* (2000) used the Kleeman *et al.* (1999) ocean model to investigate wind-forced STC variability in greater detail, forcing the model with idealized wind patches of zonal wind that were either switched-on or oscillatory. Among other things, they find that significant equatorial SST anomalies are generated by changes in the zonal wind field along y_d, generally consistent with the measure of STC strength in (2). In further support of the idea, Nonaka (2000; see abstract) analyzed the output of a GCM forced by interannual winds, showing that at decadal time scales tropical SST anomalies are closely correlated with variability in the STC heat transport whereas the relationship did not exist at all at interannual time scales.

5. Coupled Models

Kleeman *et al.* (1999) and Solomon (2000; priv. comm.) identified the $v'\bar{T}$ mechanism of STC variability as an important process as a critical process in their coupled models. In both of these studies, the ocean component is a variable-temperature version of the Lu *et al.* (1998) $3\frac{1}{2}$-layer model and the atmospheric component is a statistical model based on correlations between observed wind stress and SST variability. Two oscillations develop, namely, an interannual (ENSO) oscillation generated by equatorial coupling and a decadal oscillation generated by mid-latitude processes. The decadal oscillation produces zonal wind anomalies along y_d, which, as just discussed, lead to changes in STC strength and changes in the size and strength of the equatorial cold tongue. This alters the equatorial SST field, which feeds back to cause decadal variability in the interannual oscillation. Thus, ENSO decadal variability in the solutions result from the STC “bridge”, which links the subtropics to the tropics. Thus, in these models the dynamics of ENSO decadal variability are very different from the dynamics of ENSO itself. Specifically, during an ENSO warm event, equatorial westerly anomalies decrease the tilt of the thermocline along the equator, thereby deepening the thermocline in the eastern basin and weakening the cold tongue. In contrast, the decadal variability results from off-equatorial wind anomalies along y_d, which deepens or shallows the thermocline *throughout* the equatorial ocean.

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V.III The Coupled System in the Pacific Ocean *(Antonio J. Busalacchi)*

Research on ocean-atmosphere coupling in the Pacific Ocean, as it pertains to tropical-subtropical interactions, is primarily on time scales longer than interannual. Work on this topic goes back over forty years when the possibility of ocean-atmosphere interactions external to the tropics, on seasonal or longer time scales, was originally hypothesized by Namias (1959, 1969) and Bjerknes (1964). Namias argued that sea surface temperature (SST) anomalies in the North Pacific can change the transient eddy activity in the atmosphere, which in turn changes the mean westerly flow reinforcing the initial SST anomalies. Bjerknes suggested that decadal variability in the North Atlantic Ocean involved interactions of the westerly wind and the subtropical gyre. These early studies still have considerable relevance to today's study of the coupled system in the Pacific Ocean. At the present time there is considerable uncertainty regarding the mechanisms of Pacific decadal variability and the geographic centers for ocean-atmosphere coupling. A number of different scenarios or possible coupled modes have been proposed. For example:

- Unstable air-sea interaction between the subtropical gyre circulation in the North Pacific and the Aleutian Low-pressure system
- Decadal variability originating in the tropics with higher latitude variability resulting from ENSO-like atmospheric teleconnections
- Air-sea coupling in the northeastern subtropics that involves transport variations of the North Pacific Subtropical Cell
- Decadal variability set by a subduction time scale
- Pacific decadal variability as a result of global ocean-atmosphere interactions

Within the context of this workshop, these hypotheses can be grouped into three general and broad categories of coupling within midlatitudes, coupling from the tropics, and tropical-midlatitude coupling.

Coupling Within Midlatitudes

The seminal work on coupling within midlatitudes was performed by Latif and Barnett (1994, 1996) in which they proposed that decadal variability in the North Pacific is based on a cycle involving unstable ocean-atmosphere interactions over the North Pacific. The observational basis for this comes from North Pacific SST and sea-level pressure that are observed to vary in phase on decadal time scales (Figure 1). Latif and Barnett used similar indices to analyze coupled GCM simulations characterized by a large spatial coherence of decadal SST in their solutions at midlatitudes. SST regression analyses centered near the Kuroshio suggested that important changes in the meridional temperature gradient were occurring on decadal time scales with subsequent implications for the forcing of the overlying atmosphere. The scenario that arose from the coupled simulations was that a spin-up of the subtropical gyre would cause positive SST anomalies in the North Pacific, a subsequent weakening of the meridional SST gradient, and a weakened Aleutian Low. Continuing through this cycle, associated with the weakened Aleutian Low were reduced westerlies, a weakened gyre circulation, coming back full circle to an enhanced meridional sea-surface temperature gradient. Ancillary atmospheric general circulation model (AGCM) experiments forced by a weakened meridional temperature SST gradient resulted in a weakened Aleutian Low. This anomalous high pressure resulted in a midlatitude forcing of the Pacific North America pattern, and an anomalous heat flux feedback into the ocean in the area off Japan. Reconstruction of the anomalous heat content from the output of their coupled simulations (Figure 2) suggested that the time scale of order twenty years for this coupled mode was set by westward propagating Rossby waves that modify the strength of the subtropical gyre circulation, its associated poleward heat transport, which then eventually leads to the generation of North Pacific SST anomalies.

Subsequent work by Xu *et al.* (1998), using a hybrid coupled model, indicated the decadal oscillation was mainly driven by surface wind stress. The use of a hybrid coupled model allowed these authors to separately partition the role in the coupling of the net surface heat flux versus that of the wind stress. When their atmospheric model responded only to the surface heat flux, the sea surface temperature signal at (25-35^{o}N) (150^o-180^{o}E) resulted in a much slower evolution of order forty years, and a larger amplitude than in the fully coupled simulations. The response to this heat flux coupling was attributed to a large amplitude standing oscillation. In contrast, when the atmospheric model only responded to surface wind stress, the period of the oscillation was closer to twenty years.

Beyond the confines of the Pacific Ocean basin, both coupled model and observations indicate there is a strong wintertime relation between the strength of the Aleutian Low and North America air temperature reflecting a projection onto the Pacific North America pattern. As a result, there exists a potential basis for predictability over North America. As demonstrated by Latif and Barnett (1996), there is a strong lead-lag relationship between the North Pacific heat content and sea-level pressure. This would suggest that a determination of the phase of the North Pacific heat content may lead to an ability to predict future sea-level pressure and down stream effects over North America.

Among the outstanding issues or questions with respect to this unstable air-sea coupling, are the processes that give rise to the sea surface temperature anomalies and the significance of the response of the atmospheric circulation. In Latif and Barnett (1996), the role of advection of mean temperature by anomalous currents was assessed versus anomalous temperature by mean currents. Both advective terms in the temperature equation were the same magnitude, but only temperature advection by anomalous currents showed a fairly consistent relation with the curl of the wind stress. One of the important aspects with respect to coupling to the atmosphere is that AGCM experiments suggest that the midlatitude response to SST is modest compared to internal atmospheric variability and may have complex seasonal and nonlinear dependencies. In addition to this sensitivity that varies from model to model, Weng and Neelin (1998) suggested that the period of the coupled oscillation is sensitive to a length-scale feedback from the atmosphere (i.e., the existence of a coupled Rossby mode), and that this would be a mechanism by which the amplitude of the coupled mode could break above the background noise of the atmosphere. Outstanding issues also extend to interpretation of the observational record. Correlations between a North Pacific SST index (25^{o}N-40^{o}N, 170^{o}E-160^{o}W) have been used to suggest that the tropics are of minor importance in this coupling in that the centroid of the coupling is within midlatitudes. For example, SST anomalies in the Equatorial Pacific have been characterized as being of opposite sign relative to the main anomaly in North Pacific and "relatively weak". This would suggest that the decadal mode has its origins in midlatitudes. However, a close inspection of such correlation analyses indicates that the correlation between North Pacific SST and tropical SST can be as high as –0.8 (Figure 3). Lastly, with specific regard to the topic of this workshop, subtropical circulation cells do not play an active role in this particular mechanism for decadal variability.

Coupling From the Tropics

The possibility of coupling from the tropics, for example via an atmospheric bridge, arose out of studies of the 1976-1977 "climate shift" such as a deepening of the Aleutian Low and a drop in sea surface temperature in the central Pacific (cf., Miller *et al.*, 1994). Prior to this, Bjerknes (1966, 1969) suggested that anomalous heating from the equatorial ocean would accelerate the Hadley circulation and midlatitude westerlies. Power spectrum analyses of a North Pacific sea level pressure index indicate a broad spectral peak at periods longer than twenty years.

Changes in sea surface temperature throughout much of the tropical Pacific have been found to lead this index by about three months (Figure 4, Trenberth and Hurrell, 1994). Atmospheric GCM studies forced by such an SST signal have demonstrated a causal link between SSTs in the tropics, and the North Pacific circulation with a deeper Aleutian Low and changes in high frequency storm tracks set up as a teleconnection due to tropical heating (Blackmon, 1983; Lau and Nath, 1994; Yukimoto *et al.*, 1996).

Similarly, coupled model studies indicate decadal variability in the North Pacific may be remotely forced by decadal variation of tropical SST (Figure 5). The coupled model solutions of Yukimoto *et al.* (1996) suggest that the subsurface ocean variability at midlatitudes is an enhanced gyre spin-up process responding to changes in the overlying atmosphere forced from the tropics. In these solutions, there is a very strong negative correlation between the North Pacific and tropical SST. In association with this, there is also a strong negative correlation between the decadal SST in the tropics and the North Pacific wind stress. Heat content anomalies were found to migrate around the subtropical gyre in response to the changes in the wind stress induced at midlatitudes.

Another form of coupling in the tropics was noted by Knutson and Manabe (1998). Their coupled GCM solutions exhibited decadal variability (of order 12 years) with a pattern involving tropical winds and heat content reminiscent of the delayed action oscillator. In contrast to the ENSO variability in this coupled model, a decadal mode was noted with westward progression of off-equatorial heat content anomalies centered at a slightly higher latitude (order 12^{o} vs. 9^{o}N). At higher latitudes, negative sea level pressure anomalies were induced that strengthened the Aleutian Low and resulted in cool SST anomalies. This midlatitude response was considerably stronger than that due to ENSO. In addition, the global coupled climate model of Meehl *et al.* (1998) contained coherent decadal climate variability that extended over the entire Pacific basin and was associated with processes in the Atlantic and Indian Ocean regions (i.e., the Pacific decadal variability could not be studied in isolation from that of the other oceans).

Among the pertinent issues for coupling in the tropics are questions related to the causes of decadal variability in tropical SST. Is the variability in coupled models a result of coupled interaction between the tropics and midlatitudes? If so, what sets the time scale? Is the decadal variability separate and distinct from ENSO, or is it a subharmonic of ENSO? In the limited analyses performed to date, the subsurface propagation of North Pacific anomalies in these coupled solutions do not appear to propagate to the equatorial thermocline to initiate the subsequent phase. In fact, meridional sections from the work of Yukimoto *et al.* (1996) suggests that, if anything, there is a possible link between the thermal structure south of the equator and the overlying decadal wind stress, but no evidence of such relation for the North Pacific. Similar to some of the questions for the midlatitude coupling, the mechanisms by which a strengthened Aleutian Low develops in response to tropical SST has not been rigorously diagnosed.

Tropical-Midlatitude Coupling

Several of the studies in Section III pointed to a relation between decadal midlatitude variability and ENSO-like decadal variability in the tropics. Two processes have been proposed to explain such tropical-midlatitude coupling. Barnett *et al.* (1999) proposed an atmospheric teleconnection in which decadal wind anomalies generated at midlatitudes extend far enough into the tropics to force decadal ocean circulation variability. An out of phase relation between North Pacific and equatorial SSTs was used to suggest that the connection is via the atmosphere and not the ocean. Since the strongest correlation between the SST and these two regions is contemporaneous, a fast bridge mechanism has been suggested to be at work. Hence, it has been argued that the connection must be in the atmosphere and not in the ocean. Moreover, coupled model results with and without an active ocean have been used to indicate that the stochastic components of midlatitude decadal climate variability extend into and force the tropics. Barnett *et al.* (1999) showed that aspects of the SST and zonal wind stress relationship exhibited in their fully coupled simulations could be reproduced, in part, when only a slab ocean or mixed layer ocean was coupled to an AGCM. Since there was no active ocean in these experiments, the decadal signal could not be induced by ENSO, nor could the oceanic gyres play an active role. However, it is important to note that a large section of the equatorial signature was not being accounted for when the ocean component was inactive.

Another approach to this problem was proposed by Gu and Philander (1997), invoking an oceanic teleconnection for interdecadal variability in which midlatitude temperature anomalies are advected to the equator within the subsurface branch of the North Pacific tropical cell (Figure 6). Under this scenario, the subtropics are linked to the tropics via subduction of temperature anomalies via the low latitude western boundary current and interior pathways. Alternatively, changes in the intensity of the shallow merdional overturning cells could also induce advective temperature changes associated with this subduction. In either advective mechanism, changes in tropical SST induced by the STC would feed back to the overlying atmosphere. The AGCM studies by Lau and Nath (1994, 1996) support the notion that tropical SST anomalies play a much larger role than midlatitude SST anomalies in driving anomalous midlatitude circulation.

Various aspects of this hypothesis have been recently investigated. Wang *et al.* (2000) argue that the characteristic subduction time scale (of order 6 years) and relative phases of the observed decadal variability between the midlatitude North Pacific and equatorial Pacific are consistent with a negative delayed action oscillator (in the meridional plane) for the observed decadal variability, but not interdecadal. Kleeman *et al.* (1999) used a hybrid coupled model to demonstrate that the decadal modulation of the equatorial SST is coincident with decadal variations in subtropical wind stress. Within the context of this model, equatorial SST was inversely related to upwelling strength. Anomalous equatorial upwelling was determined by the meridional transport in the upper ocean out from the tropics, i.e., SST anomalies in the tropics are generated by heat transport anomalies associated with variations in the North Pacific STC. Model experiments that restricted the coupling to various geographic regions point to the northeastern subtropics as being the key region of ocean-atmosphere interaction. Correlation analyses of their solutions indicated that the positive feedback for the decadal oscillation was provided by latent heat flux anomalies in the northeastern subtropics due to wind speed variations. The delayed negative feedback arose from anomalous horizontal advection and convective overturning in this simulation. These solutions suggest that the decadal modulation of tropical SST is determined by decadal variation in subtropical wind stress (Figure 7).

A number of the outstanding questions for tropical-midlatitude coupling are similar to that of the previous sections. Among these, what is the relative importance among transport changes, temperature advection, and forcing originating external from the tropics? What are the related changes in water mass attributes, and are they significant? What is the relative importance of the influence of midlatitude versus tropical SST on a midlatitude atmosphere? Ultimately, one of the most important questions is what is the atmospheric response to SST variability induced by the subtropical cells?

Summary

As it pertains to coupled hypotheses of decadal variability in the Pacific Ocean, the potential role of the STC is uncertain at best. A variety of coupled modes have been proposed to account for the observed decadal variability:

A) A coupled mode intrinsic to the midlatitude North Pacific involving changes in the gyre circulation in response to changes in the merdional gradient of SST. However, the atmospheric response to midlatitude SST anomalies is generally too weak to be simulated unambiguously in many GCM experiments;

B) Decadal tropical SST inducing changes in the midlatitude atmosphere and subsequent changes in the ocean gyre circulation. The source of the decadal SST at low latitudes remains uncertain;

C) Decadal ENSO-like variability at low latitudes with delayed action oscillator attributes over a larger meridional extent;

D) Tropical-midlatitude coupling via subduction of temperature anomalies. A delayed action oscillator mechanism operant meridionally with time scale set by subduction pathways from North Pacific; and

E) Tropical-midlatitude coupling via changes in the strength of the North Pacific STC.

In the context of this workshop, this range of hypotheses and related questions that remain unanswered provide a basis and framework for observational, modeling, and process studies needed to enhance our understanding of the role of shallow tropical-subtropical circulation cells and their interaction with the atmosphere. The results of such studies, in turn, will also lead to improved understanding of the coupled climate system in the Pacific Ocean basin on time scales longer than seasonal to interannual.

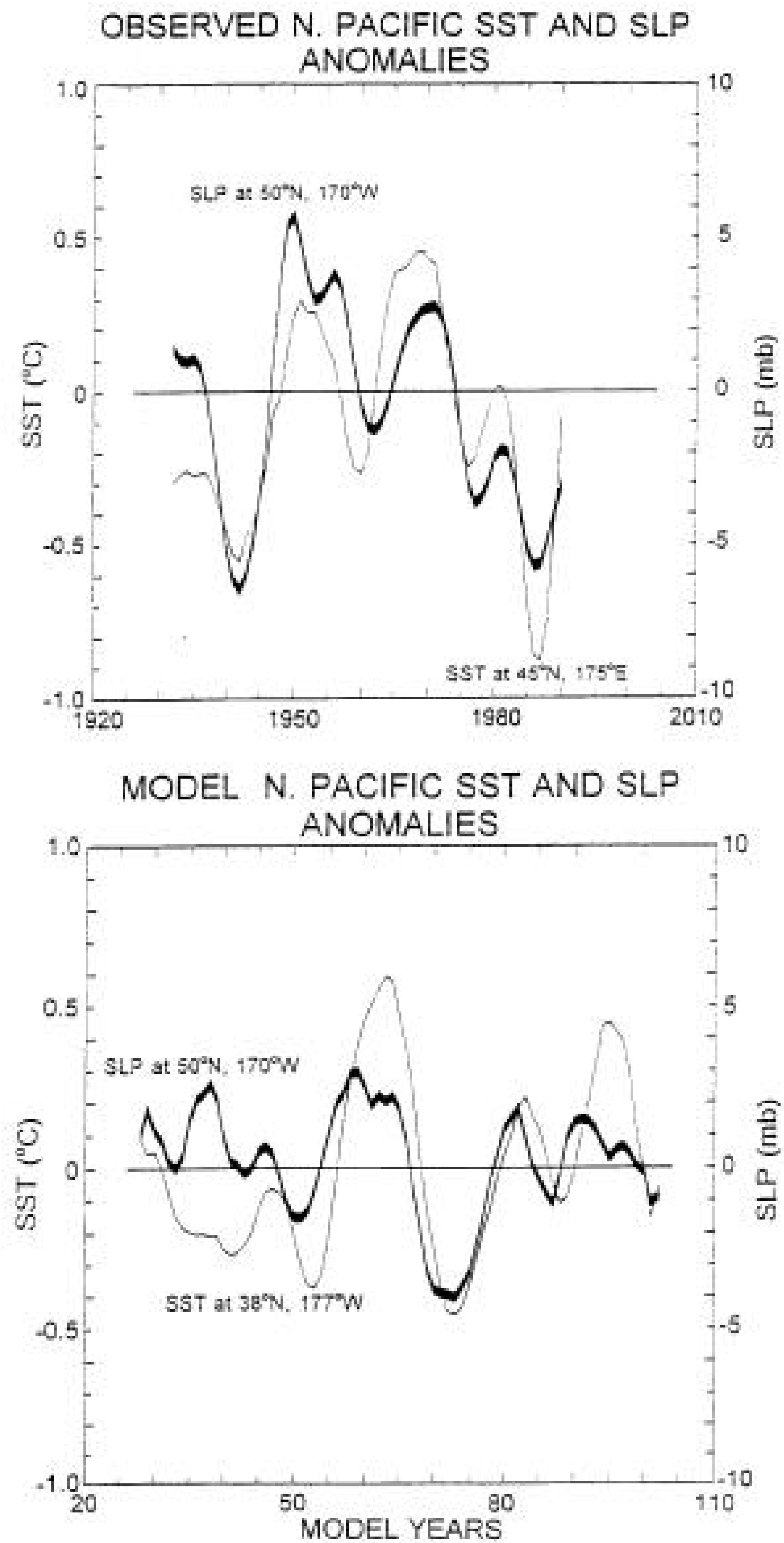
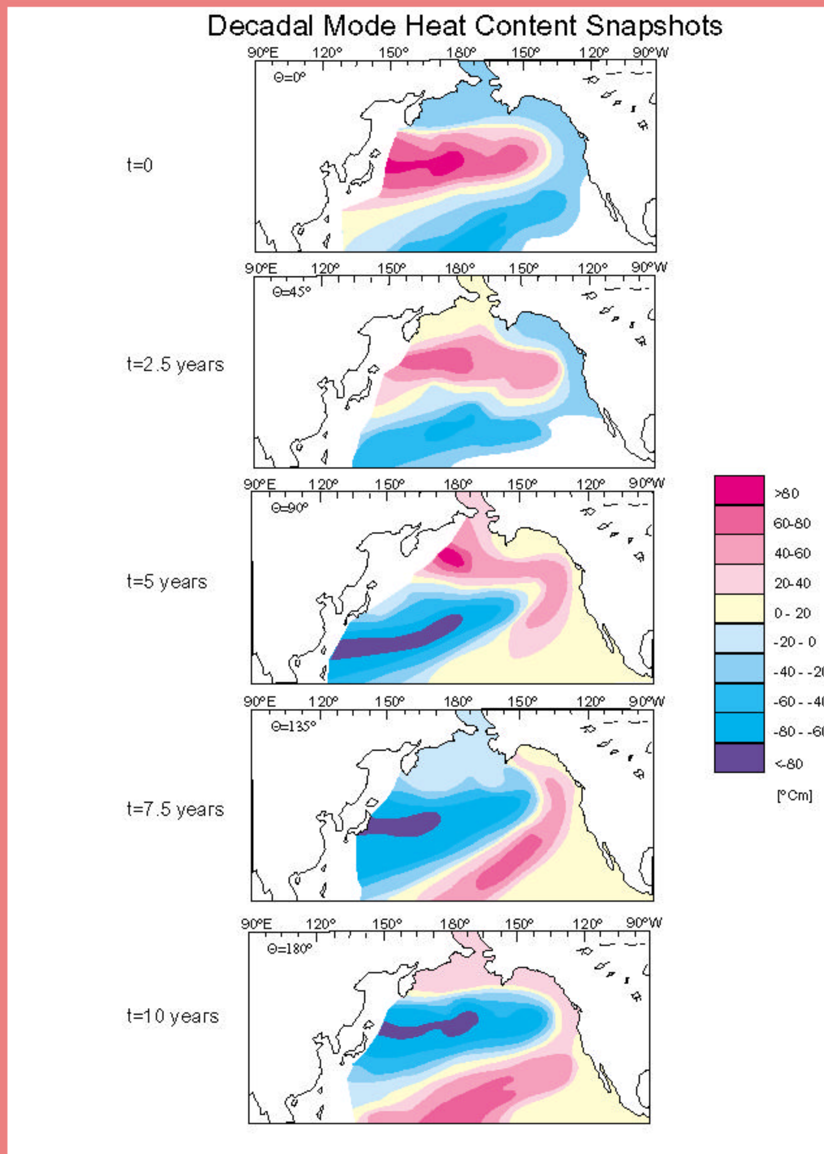


Figure 1: Low-pass filtered anomalies (retaining variability on time-scales longer than 5 years) of North Pacific SST (°C) and SLP (hPa) as observed. From Latif and Barnett (1996).

Decadal Variability in the North Pacific

- The Latif-Barnett Mode-



Reconstruction of anomalous heat content from the output of a coupled model run in the Pacific Ocean. The individual panels show the progression of heat content anomalies at approximately 2.5 years apart (Latif et al., 1994, Science, 266, 634-637).

AV/D4/99-1

Figure 2

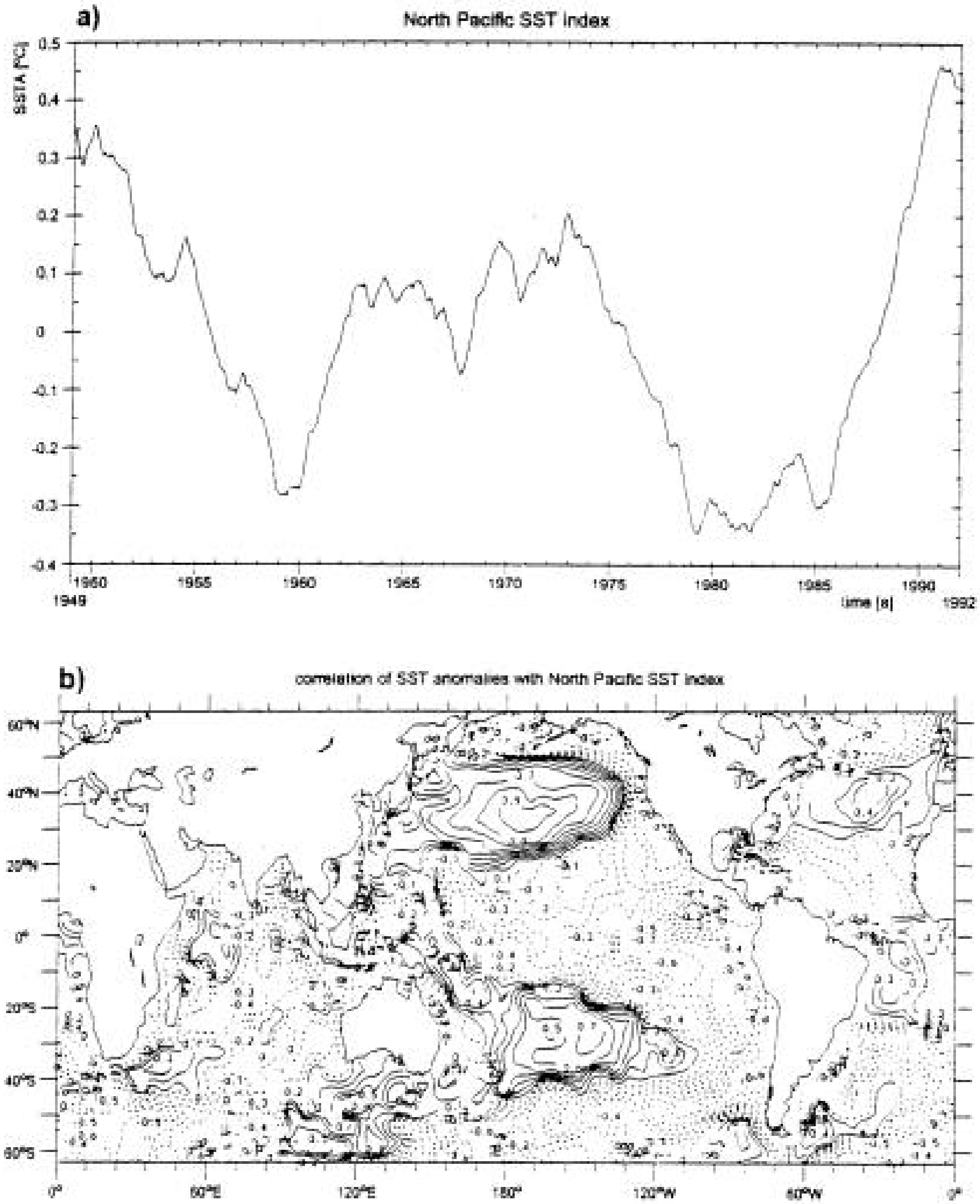


Figure 3: (a) Time series of North Pacific SST anomalies averaged over the region 25°-40°N, 170°E-160°W and spatial distributions of correlation coefficients of (b) observed SST anomalies with the index time series shown in (a). All data were detrended and smoothed with a 5-yr running mean filter prior to the correlation analysis. From Latif and Barnett (1996).

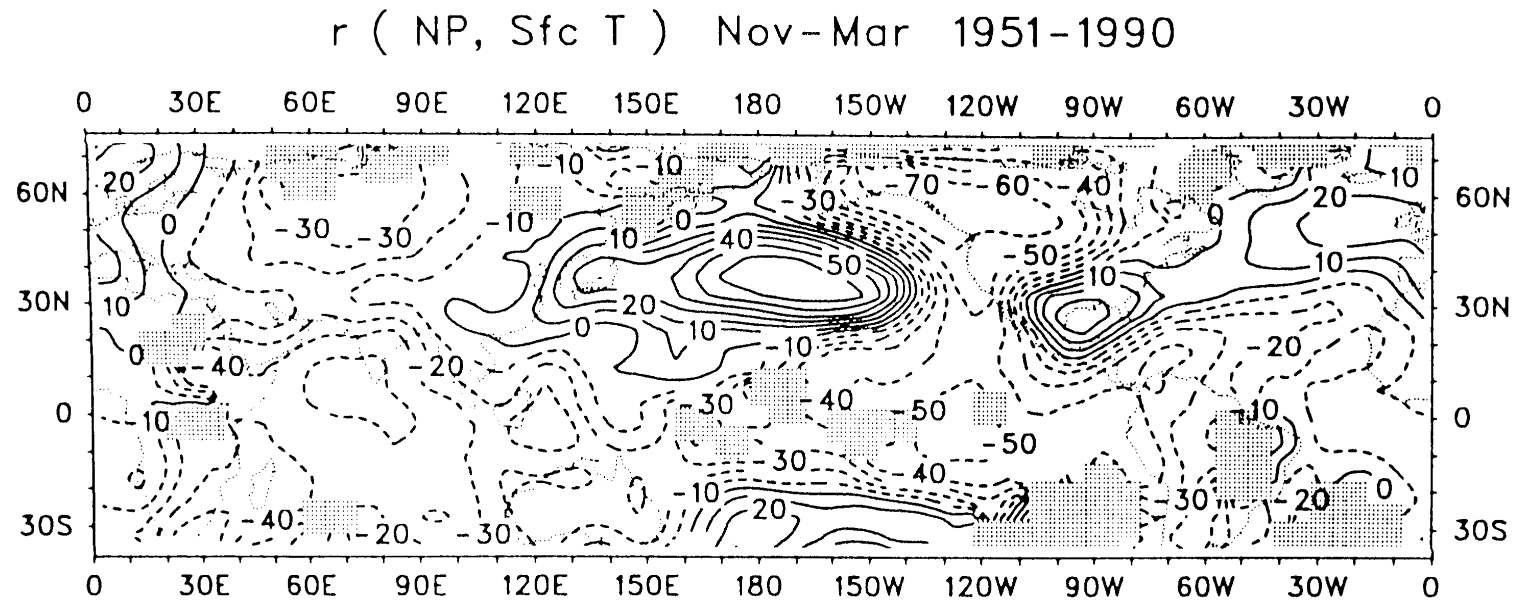


Figure 4: Correlations (%) of the five month mean November-March NP index with surface temperatures at zero lag for 1951-1990. Negative values are dashed and areas of insufficient data are stippled. From Trenberth and Hurrell (1994).

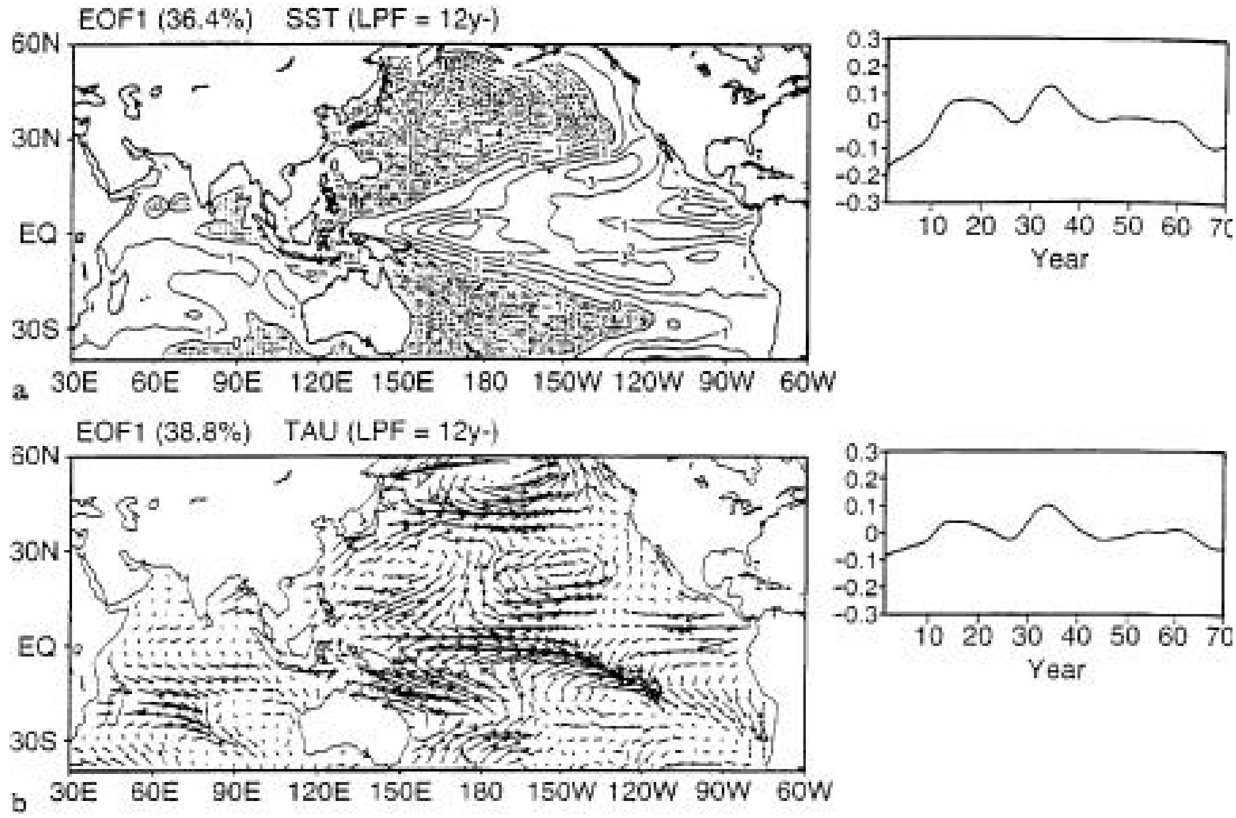
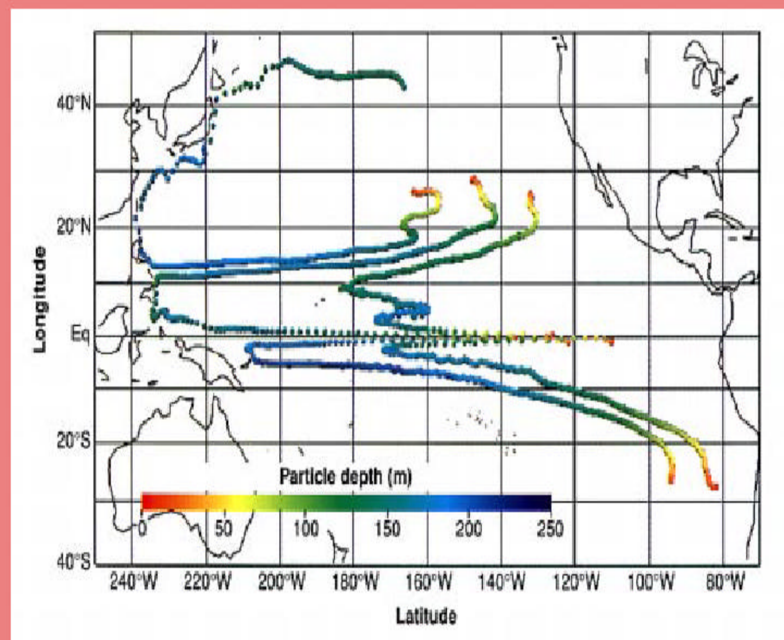


Figure 5a, b: Eigenvectors and time coefficients of the first EOFs of simulated a) SST and b) wind stress for interdecadal time scale (periods longer than 12y). Negative region is shaded in a). Values of eigenvectors are in proportion to amplitude of variance, since EOFs are calculated for variance matrix. From Yukimoto *et al.* (1996).

Decadal Variability in the North Pacific

- The Subduction Hypothesis -



The paths of water parcels over a period of 16 years after subduction off the coasts of California and Peru as simulated by means of a realistic oceanic general circulation model forced with the observed climatological winds. From the colours, which indicate the depth of the parcels, it is evident that parcels move downward, westward, and equatorward unless they start too far to the west off California, in which case they join the Kuroshio Current. Along the equator they rise to the surface while being carried eastward by the swift Equatorial Undercurrent (Gu, Philander, 1997, Science, 275, 805-807).

AV/D4/99-2

Figure 6

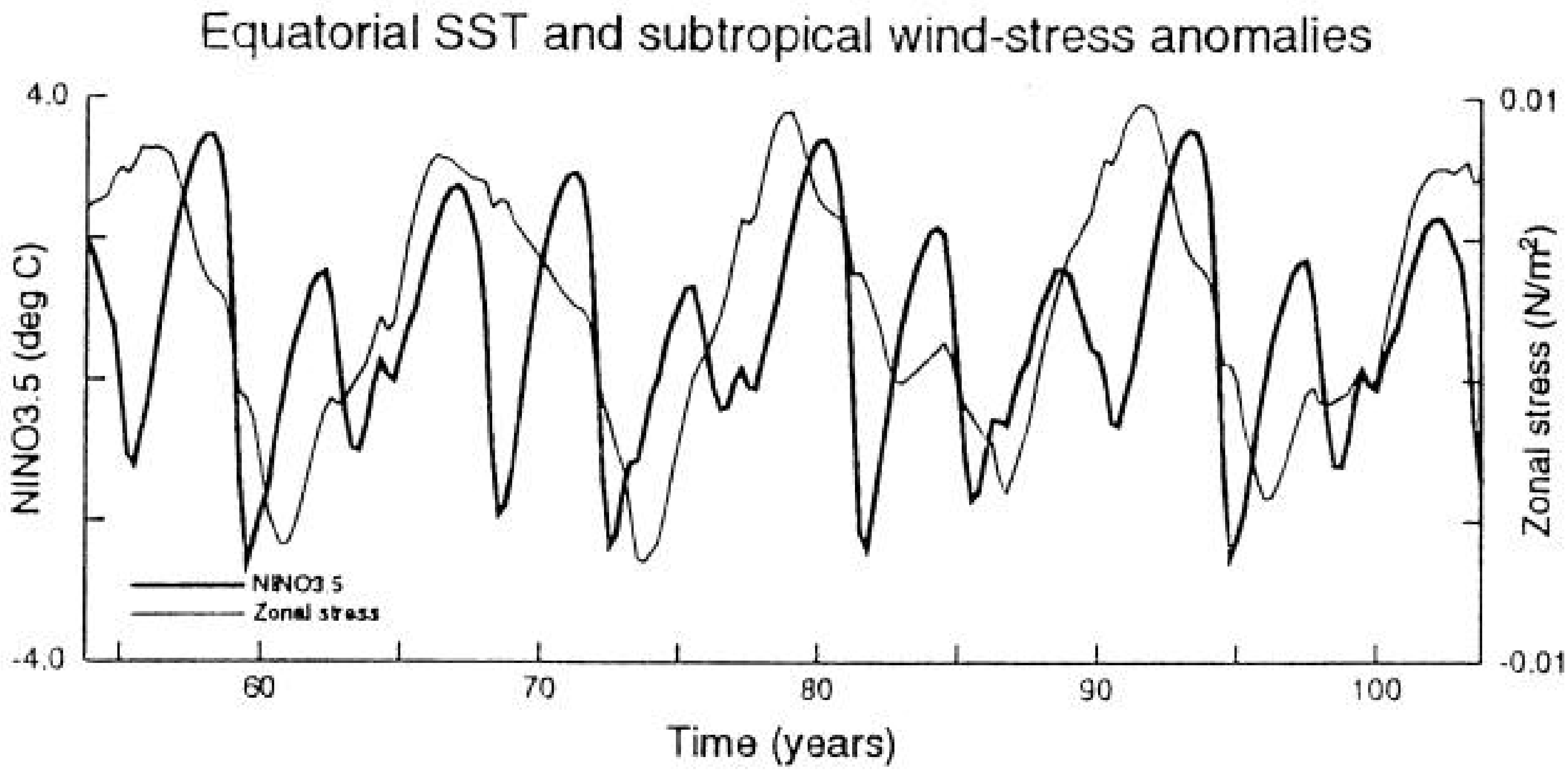


Figure 7: Time series of SST anomalies in the NINO3.5 region (thick curve) and subtropical zonal wind-stress anomalies (thin curve) for the main run. From Kleeman *et al.* (1999).

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V.IV Shallow Thermohaline Cells in the Indian Ocean: Observations *(F. Schott)*

Atmospheric Forcing Situation

The Indian Ocean differs significantly from the Pacific and Atlantic in not having a mean equatorial Ekman divergence and upwelling. Upwelling-favourable (i.e. westward) winds that cause an equatorial undercurrent, can only occur along the equator in boreal spring and are weak. Instead upwelling occurs off Somalia and Arabia, driven by the strong SW monsoon winds, and to a small extent also around India. Subduction of water masses that can in the end supply the upwelling regions occurs in the southern subtropical gyre of the Indian Ocean. Maximum density flux into the Ocean is west of Australia where winter cooling combines with haline forcing from the excess of evaporation over precipitation (Zhang and Talley, 1998).

The consequence of these forcing patterns is that the shallow thermohaline circulation of the Indian Ocean must be cross-equatorial, connecting southern subduction with northern upwelling sites.

The meridional Ekman transport is southward during the summer monsoon on both sides of the equator, providing a southward upper-layer return link for the cross-equatorial STC. During the winter, the meridional Ekman transport is northward and weaker than during the summer monsoon. As a consequence of the reversing wind fields, the cross-equatorial heat export of the northern Indian Ocean shows drastic seasonal changes, from 1.5PW southward in northern summer to .5PW northward in winter (Hastenrath and Greischar, 1993;Garternicht and Schott, 1997).

Right on the equator meridional winds in both seasons are directed against the Ekman transports, i.e. northward during the SW monsoon, southward during the NE monsoon. This forcing causes a shallow meridional cell or roll on the equator.

Large interannual anomalies of the forcing fields have been observed, the Indian Ocean Dipole Mode events (Saji *et al.*, 1999;Webster *et al.*,1999), during which anomalous westward winds along the equator cause Ekman divergence and upwelling in the eastern Indian Ocean. During these IDM events, of which particularly strong episodes occurred in 1993/4 and 1997/8, with equatorial Ekman divergences of 20-30 Sv between 3N and 3S in the eastern basin, the eastern Indian Ocean resembles more the other two oceans.

Subduction

Subduction in the Indian Ocean occurs dominantly in the southern subtropical gyre during southern winter (Zhang and Talley, 1998; Karstensen and Quadfasel, 2000). These latter authors estimated an annual mean subduction of 23 Sv, based on the method of Marshall *et al.* (1993), using the mixed-layer depths and referenced geostrophic currents derived from Levitus, and Ekman transports and pumping from the SOC climatology (Josey *et al.*, 1996). Another evaluation, limited to subtropical underwater (STUW) was by O'Connor, Fine and Olson (pers. comm., 2000), where the lateral induction term was determined using drifter velocities.

Subduction can also in the northern Arabian Sea in northern winter. Karstensen (pers. Communication, 2000) estimated this at about 2 Sv, in agreement with observations of Schott and Fischer (2000).

Upwelling

The Somali upwelling occurs in two locations, at the offshore flow of the "Southern Gyre", at 3-5°N, and of the "Great Whirl", at 9-12°N. Upwelled water can be as cold as 15°C, corresponding to densities of 26.3 kgm $^{-3}$, but typical upwelling temperatures are in the range of 21-23°C. Based on ADCP profiles and hydrography of our WOCE cruises 1995, we determined the offshore flow of upwelled waters at densities <24.5 out of the Southern Gyre at

9 Sv, concluding from sections further out that these water masses were irreversibly transformed into surface waters.

The outflow of the GW upwelling occurs along two pathways: through the Socotra Passage to the north, and by eastward recirculation within the GW. From shipboard and moored observations in the passage between Socotra nd the main land, the northward outflow of upwelled water was estimated at 2 Sv, and the GW outflow at densities <24.0 to the east and recirculation southward was estimated at 10 Sv. In summary, the Somali upwelling would yield about 20 Sv for the period of the summer monsoon.

Upwelling off Oman is weaker and associated with topographic features, feeding into squirts and filaments that can reach far offshore. Early estimates (Smith and Bottero, 1977) obtained about 2 Sv and a more recent reevaluation yielded 1.5 Sv for 1993-95 (Shi *et al.*, 2000). Upwelling around India is weak and will be neglected here. Taking the Somali and Oman estimates as valid for a four month period (June-September), yields an annual mean of about 7 Sv of upwelling from isopycnals above $\_{\theta}$ = 24.5.

Another regime where upwelling is presumed to play a role is in the 5-10°S zone, at the northern end of the Trades due to the associated Ekman divergence. This is clearly seen in a doming of isotherms and isopycnals in that region in the historical data base. At the surface, while SST does not show a minimum, a chlorophyll maximum is observed at times (Murtagudde *et al.*, 1999). This zone is potentially important because upwelling here affects SSTs near the critical temperatures of 28-29°C.

In addition, sporadic upwelling can occur along the equator during IDM events. Yu and Reinecker (2000) estimated from XBT data that the thermocline in the east during the 1997/8 event raised by about 80 m. Assuming that this applies to a regime 2°S - 2°N and linear decaying to the central Indian Ocean (80°E) yields an upward transport of about 25 Sv; this is in agreement with the Ekman divergence from the anomalous IDM easterlies for that event. However, only a fraction of this water will be irreversibly transformed into surface water.

Pathways

Water subducted in the southern subtropical gyre is advected westward with the SEC, adding to the Indonesian Throughflow. At the western end and south of about 17°S this flow follows the SE Madagascar current into an eddy and retroflection regime south of Madagascar, from there on it partially recirculates into the subtropical gyre, partially joins the Agulhas (e.g. Schott and McCreary, 2000). The northern branch of SEC inflow passes north of Madagascar and partially supplies the East African coast current, partially flows southward through the Mozambique Channel. Ganachaud *et al.* (2000) obtained a transport of about 15 Sv for this flow from a global inverse model calculation. The evidence from earlier hydrography (Swallow *et al.*, 1991) as well from recent drifters and float tracks is that the Mozambique Channel flow is supplied more by the SEC thermocline layer, while the surface waters proceed to the EACC and Somali Current.

At the equator, an annual-mean northward Somali Current transport of 10.5 Sv was determined from moored current observations (Schott *et al.*, 1990). For the summer monsoon, the transport was 21 Sv. Out of these 21 Sv, about 15 Sv fall into the depth range shallower than 200 m that can supply the Somali and Arabian upwelling regimes. Water mass properties show a continuation of the SEC into the Somali Current and thus link the southern subduction with northern upwelling.

The Equatorial Role

As stated in the beginning, meridional winds right on the equator should drive cross equatorial currents *against* the Ekman transports in both monsoon seasons. Consequently, numerical models with high vertical resolution show a shallow meridional roll on the equator in the upper 100 m (Wacogne and Pacanowski, 1996; Garternicht and Schott, 1997).

Ship sections from R/V "Meteor" in the western Indian Ocean show the existence of this roll for the summer monsoon of 1995 with northward surface flow of > 20 cm s^{-1} above 30 m and southward flow of 15 cm s^{-1} below, down to about 100 m. During 1993-94, observations with moored ADCPs were carried out across the equator at 80.5°E (Reppin *et al.*, 1999) and the meridional velocity also show the roll for the summer monsoon of 1993.

The effect of the roll is to connect the southward (northward) Ekman transports during the SW (NE) monsoon across the equator underneath the northward surface currents. The question is whether the roll involves diapycnal fluxes. If not, it is of no further importance for the consideration of Indian Ocean STCs. If there is a diapycnal effect at all, the available data suggest it to be small.

Anomalies

Very little is known on Indian Ocean anomalies and nothing on the possible role of STCs in them. The Indian Ocean Dipole Mode (Saji *et al.*, 1999; Webster *et al.*, 1999) has been shown to involve upper-layer advection by warm pool waters in the west, south of the equator, that enhance atmospheric convection over the African continent; and by equatorial and cold water advection in the eastern basin. How much watermass transformation and thus STC relevance might be involved with the large eastern equatorial upwelling during these episodes still needs to be determined.

The XBT section data from the western Indian Ocean (obtainable from the CSIRO Web site) show significant interannual variability at thermocline levels. To determine how much of this is mere isopycnal heaving and how much STC water mass variability would need better coverage of T/S profiles, to be anticipated with the buildup of ARGO.

Further, observations on the interannual variability of southern subtropical subduction and continuous measurements of transports, vertical structure and watermass composition of the Pacific to Indian Ocean Throughflow are essential for a better understanding of Indian Ocean STCs and their potential role in interannual to decadal climate variability.

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V.V Shallow Overturning Cells in the Indian Ocean *(Julian P. McCreary, Jr.)*

Two types of meridional overturning cells have been discussed in the Indian Ocean. One involves flow of deep water into the basin from the south and compensating outflow at intermediate depths. The other type involves shallower circulations that carry thermocline water from subduction areas in the southern hemisphere to upwelling regions north of the equator. This talk focuses on the latter type, which appears to be more important to the Indian Ocean's annual-mean heat budget. See Schott and McCreary (2000) for a more comprehensive discussion.

1. Two-Dimensional Structure

Annual-mean meridional streamfunction plots from GCM solutions all show the presence of a shallow, overturning circulation that extends from the southern Indian Ocean subtropics into the northern hemisphere. It consists of southward near-surface flow, downwelling (subduction) in the southern region, northward flow in a depth range from about 100-500 m, and upwelling in two regions, namely, in a band from 5-10°S and in the northern hemisphere (Wacongne and Pacanowski, 1996; Garternicht and Schott, 1997; Lee and Marotzke, 1998). There is also an intriguing "equatorial roll" within the upper 100 m, in which the southward return flow crosses the equator underneath a surface *northward* flow in the upper 25 m or so. Thus, the streamfunction maps suggest the existence of two shallow overturning cells: one confined to the southern hemisphere that is closed by upwelling in the 5-10°S band, the South Indian STC (also called the "Tropical Cell" by McCreary *et al.*, 1993); and another that extends into the northern hemisphere, the Cross-Equatorial Cell (CEC).

There are considerable differences among the solutions, which are certainly partly attributable to differences in model formulation. For example, the solution discussed by Garternicht and Schott (1997) has a very strong STC with the CEC having a strength of only about 2 Sv, whereas the Lee and Marotzke (1998) solution indicates a stronger CEC with a strength of about 6 Sv. In the McCreary *et al.* (1993) solution, the strengths of the CEC and STC are 5 Sv and 6.5 Sv, respectively. It is interesting that the models all show a significant STC with upwelling in the 5-10°S band, but observed SST does not show a zonal minimum in this region, suggesting that upwelling in the region is weak or sporadic. On the other hand, climatological mixed-layer depths show a zonally extended minimum there during both monsoons, and temperatures at 100 m have a minimum there. Further support for this upwelling is provided by satellite imagery of ocean color, which sometimes indicates the presence of phytoplankton blooms along the band (Murtugudde *et al.*, 1999).

Seasonal streamfunction maps give the impression that there is large variability in the strength and structure of the deep overturning circulations. Physically, however, the variability mostly represents an adiabatic sloshing back-and-forth of water masses, not diabatic across-isopycnal flow. Thus, it is mostly indicative of changes in heat storage, rather than cell strength.

2. Three-Dimensional Structure

The three-dimensional circulation patterns taken by water parcels in the Cross-Equatorial and Tropical Cells are much more complex than suggested in the two-dimensional streamfunction maps. Modelled flow fields generally compare well with the observed ones. For example, in the McCreary *et al.* (1993) solution lower-layer water circulates around the southern-hemisphere subtropical gyre, and some joins the EACC. It then either crosses the equator within the East African Coastal Current (EACC) to feed the upwelling regions in the northern hemisphere (off Somalia, Oman, and India), or it bends offshore just south of the equator to join the South Equatorial Countercurrent (SECC). Most of the SECC water then bends southward to provide water for the open-ocean upwelling in the 5-10°S band and to participate in the South Indian STC; the rest reverses to flow westward along the equator, and subsequently moves into the northern hemisphere. In the surface layer, upwelled water flows eastward and southward across the interior Arabian Sea, eventually crossing the equator in the interior ocean.

Haines *et al.* (1999) and Miyama (2000, pers. comm.) diagnosed similar flow patters for both annual-mean and seasonally varying circulations by tracking Lagrangian tracers and drifters in their solutions. In the latter study, the circulations were confirmed to exist in several types of ocean models, varying in complexity from the McCreary *et al.* (1993) $2\frac{1}{2}$-layer system to the JAMSTEC GCM. The two studies also show that subsurface source waters for the STC and CEC originate in the Indonesian Throughflow and also flow into the basin across its southern boundary (that is, from south of 30°S).

3. Cross-Equatorial Surface Flow

Levitus (1988) noted that the annual-mean winds were westerly just north of the equator and easterly just south of it. He therefore suggested that the southward interior flow was caused by Ekman drift, which is directed southward on both sides of the equator. A difficulty with this interpretation, of course, is that Ekman theory breaks down near the equator, so that the flows that actually carry water across the equator are not specified.

Godfrey (2000, pers. comm.) and Miyama (2000, pers. comm.) provide a theoretical explanation that circumvents this dilemma. They note that during both monsoons the zonal component of the winds nearly vanishes at the equator and that it varies roughly proportional to *y* on either side. The Ekman pumping velocity [proportional to curl (τ / f)] for such a wind is identically zero, so that isopycnals are never shifted vertically at all. It follows that this forcing generates *no geostrophic currents* but only Ekman flows. For this wind, then, the concept of Ekman flow is valid all the way to the equator: Both τ^x and f vanish as y goes to zero so that

τ^x / f remains well defined. Thus, the cross-equatorial currents that connect the off-equatorial Ekman flows can also be interpreted as Ekman currents.

An equivalent interpretation is that the interior, cross-equatorial flow is a quasi-steady Sverdrup flow driven by the wind curl associated with this special zonal wind field (proportional to *y*). Sverdrup flow is a composed of both Ekman and geostrophic components. Since this special wind field drives no geostrophic currents, the cross-equatorial flow again is equal to the Ekman drift. Mathematically, the equivalence is apparent from the relation $v = (\mathrm{curl}\,\tau^x)/\beta = -\tau^x/f$ where $\beta = f_y$, which holds only for this special wind field.

4. Equatorial Rolls

Equatorial rolls were first noted by Wacongne and Pacanowski (1996) in their GCM solution. In their solution, the roll was strongest during the summer. At this time, although the antisymmetric zonal wind drives a net southward flow across the equator in the upper 100 m, near the equator it occurs *beneath* a northward surface current associated with the equatorial roll. An oppositely directed and weaker roll exists during the winter.

Wacongne and Pacanowski (1996) hypothesized that the rolls were a response to cross-equatorial winds that are present during the monsoons, as suggested earlier by Philander and Pacanowski (1980; see McCreary, 1985, for further discussion). Similar equatorial rolls developed in Philander and Pacanowski's (1980) solutions forced by uniform meridional winds, consisting of cross-equatorial surface flow in the direction of the wind, upwelling on the upwind side, and downwelling on the downwind side. Miyama (2000, pers. comm.) confirmed that meridional winds are the driving force for the Indian-Ocean rolls, by analyzing solutions to both linear and nonlinear models forced by realistic winds with and without a meridional component.

Whether the rolls cause diapycnal fluxes depends on the mixed-layer physics of the respective model. This does not occur in the models described above because their rolls are contained in the mixed-layer and so have no effect on the heat budget. Schiller *et al.* (1998) commented on the diapycnal fluxes of the cells in their GCM. They found that when the mixed-layer depth was less than the vertical scale of the equatorial cell, which was the case in May in their simulation, there was a band of heat uptake by the upwelled water south of the equator.

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V.VI Summary of Observational Evidence for Atlantic Subtropical Cells

(Robert L. Molinari)

The canonical Subtropical Cell (STC) has been characterized as a coupled mode of air-sea variability with decadal periodicity. The memory of the mode is found in the ocean where time dependent subduction brings subtropical surface waters with variable properties to depths between 100m and 300m. These subsurface waters are then advected to the equator where upwelling transports the anomalous water mass characteristics back to the surface. Ekman divergence transports return the upwelled water to the subduction area. The anomalous equatorial SST patterns perturb the atmosphere both locally and over the subduction area re-introducing anomalies in the properties of the subducted waters.

In the Atlantic, there is some evidence for a canonical mean STC in the southern hemisphere, but not the northern hemisphere. There are few data to establish if the southern hemisphere STC has a preferred temporal signal. In addition to the lack of sufficient data, identification of STC properties is complicated in the Atlantic by the presence of other coupled modes of variability in the basin. These modes include: the annual cycle (Mitchell and Wallace, 1992); ENSO-like variability (Zebiac, 1993; Carton and Juang, 1994); ENSO-driven anomalies (Enfield and Mayer, 1997); NAO-tropical connections (Tourre *et al.*, 1999); a coupled thermodynamic mode that can explain the Atlantic's SST dipole (Chang *et al.*, 1997); and the global thermohaline circulation.

In the southern hemisphere, water mass properties and direct velocity observations along the western boundary provide evidence for a mean STC. Both at 100m, the core depth of the Equatorial Undercurrent (EUC), and the sea surface, highest salinity concentrations are observed close to the western boundary between about 10S and 20S (Levitus, and Boyer, 1994). These waters are advected to the EUC by a combination of a direct route within the North Brazil Current (NBC) and a somewhat circuitous route across the Atlantic in the South

Equatorial Countercurrent and then back to the west in the South Equatorial Current (Stramma and Schott, 1999).

Direct velocity observations indicate that about 20 Sv of NBC transport above the 26.8 sigma-theta surface retroflect north of the equator into the EUC (Schott *et al.*, 1995). The NBC transport does not only feed the EUC, but also the North Equatorial Countercurrent (NECC) and North Equatorial Undercurrent (NEUC), (Bourles *et al.*, 1999).

In the northern hemisphere, high salinity water subducted in the subtropics is transported to the western boundary in the North Equatorial Current (NEC). Synoptic cruise data indicate the NEC bifurcates near the western boundary and the southward branch feeds the NECC and NEUC (Bourles *et al.*, 1999). On the order of 5 Sv of northern hemisphere water contributes to the transport of these two eastward countercurrents.

The available literature show no direct contribution of northern hemisphere waters to the EUC west of 44W. Similarly, there is no evidence east of 44W showing that a portion of the NECC/NEUC transports move southward to become entrained in the EUC. Additional analyses is needed to determine (1) if tropical cells (Lu *et al.*, 1998) or some other mechanism can transport northern hemisphere waters to the EUC and/or (2) STC upwelling occurs off the equator in the North Atlantic (e.g., the upwelling zone off western Africa).

The intensity of the equatorial upwelling limb of the STC has been quantified by Roemmich (1983). He estimated a divergence of about 26 Sv between 8S and 8N partitioned somewhat equally across the two latitudes. Garzoli and Molinari (2000) estimate a somewhat smaller divergence between 6S and 6N, 17 Sv , using data from a July-August, 1997 cruise. The surface transport across the two latitudes is asymmetric (1 Sv north across 6N and 16 Sv south across 6S).

Evidence for equatorial Ekman transport appears in satellite tracked drifting buoy trajectories. Buoys on both sides of the equator move poleward in the trajectories of Richardson and

Reverdin (1987). In addition, many of the buoys deployed between 5S and 10S were entrained in the NBC and thus did not return directly to the southern hemisphere subduction zone. Similarly, northern hemisphere buoys typically join the NBC and do not return directly to the subduction region in the southeastern subtropical North Atlantic.

Long-time series of the oceanographic features just described are limited in the tropical Atlantic, as are time-series of atmospheric phenomena. Sea-level pressure (SLP) and SST observations from commercial vessels represent the most complete data-set available.. The Tourre *et al.* (1999) analysis of SLP/SST data indicates coupled peaks in variability at periods of 3.5 and 11.4 years. The spatial patterns of both modes indicate propagation of SST signals from the subtropics to the tropics in both hemispheres. However, it can't be discerned from this analysis if these spatial modes represent an STC or some other feature (i.e., there are no subsurface data included in their study). Furthermore, the SLP spatial patterns is NAO-like with no large signal in the equatorial Atlantic. Thus, although these modes could explain subtropical forcing of anomalous subduction they are insufficient to describe both equatorward translations of subduction anomalies and anomalous SST forcing of atmospheric features at the equator.

In summary, mean hydrographic and current distributions argue for a canonical southern hemisphere STC. That is, water subducted in the southern subtropics is advected to the equator and becomes entrained into the EUC. Equatorial upwelling and Ekman divergences then bring these waters back to the subtropics where subduction completes the oceanic limb of the STC. Similar evidence for a canonical northern hemisphere STC is not as striking. In particular, water subducted in the subtropics does reach the tropics to at least the latitudes of the NECC and NEUC. However, additional analyses and probably data are needed to determine if these waters leave the two countercurrents and become entrained into the EUC. Perhaps other mechanisms are active in the northern hemisphere to bring the subducted waters back to the surface (e.g., upwelling off west Africa).

Observations of the temporal variability of the oceanographic and atmospheric features that are part of an STC are limited to surface measurements. These observations suggest that decadal variability is characterized by an SST dipole and anomalies in the trade winds (i.e., consistent with the thermodynamic mode described by Chang *et al.*, rather than an STCmode). Additional analyses of the existing subsurface oceanic and surface atmospheric data are underway to search for STC signals

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V.VII Circulation and Warm Water Pathways in the Equatorial Atlantic: Model Simulations *(Claus Böning)*

(1) Historical Perspective

Although the issue of subtropical-tropical connections and pathways has only recently received increased attention, it is instructive to inspect also some of the earlier model studies, in particular as a prerequisite for an understanding of model assumptions and parameter sensitivities critical for the simulated behaviours. The first GCM studies taking account of the interplay between the MOC and the upper-layer equatorial currents and instability waves, specifically aiming at the seasonal mass and heat budgets of the equatorial Atlantic, were due to Philander and Pacanowski (1984, 1986a,b; hereafter PP). Using a model configuration of which many aspects have served as a standard until today (latitudinal resolution of 1/3-deg, vertical resolution 10 m, Ri-number dependent vertical mixing), the solutions began to provide a three-dimensional view of the complex, seasonally modulated tropical pathways; it included, for example, a strong asymmetry in the supply of the equatorial upwelling regime, drawing primarily subtropical water from the south, by the upper limb of the basin-scale MOC.

A closer inspection of the sources and pathways of the equatorial current system was provided by Schott and Böning (1991) based on experiments conducted as part of the WOCE "Community Modelling Effort" (CME) (Bryan and Holland, 1989; for a review of the CME see also Boening and Bryan, 1996). In general, the host of model experiments of the CME, although not specifically tuned for an investigation of equatorial dynamics (e.g., it used a vertical viscosity formulation incapable of simulating the zonal momentum budget of the EUC), helped to elucidate some important GCM sensitivities, particularly with respect to lateral mixing parameterization (e.g., the NEUC and SEUC emerging upon switching from harmonic to biharmonic formulation) and wind forcing: e.g., Bryan *et al.* (1995), examined the effect of three different wind stress climatologies, and noted a very strong dependence of the equatorial upwelling (between 2.6 S and 2.6N) on the wind stresses at the equator; a large fraction of this upwelling was associated with an intense tropical overturning cell between the equator and the NECC.

Concerning the subtropical-tropical connections, the model analysis of Schott and Böning revealed a predominant route of South Atlantic (SA) water via a western boundary undercurrent (NBUC), a feature subsequently confirmed by current measurements (Stramma *et al.*, 1995). In the northern hemisphere, a southward undercurrent off French Guyana ("Guyana Undercurrent", GUC) was shown to draw high-salinity subtropical water towards the equator; in winter it added to the SA supply of the EUC, in summer it connected to the NECC/NEUC. A similarly prominent expression of a northern hemisphere STC, has not been reproduced in later Atlantic Ocean GCMs. A particular example to be noted here is the model intercomparison study conducted in the EU-"DYNAMO" project (DYNAMO group, 1997): for all model cases differing in vertical coordinate schemes (isopycnic layers, sigma- and geopotential-levels), the zonally-integrated transport patterns revealed a much weaker shallow overturning cell than for the CME, and correspondingly, rather less prominent expressions of a GUC (Barnier *et al.*, 2001).

(2) The Role of the MOC

A leading factor responsible for the asymmetry of the STCs in the Atlantic Ocean, providing a rationale for these different GCM behaviours, was elucidated in recent, idealized model studies (Fratantoni *et al.*, 2000; Jochum and Malanotte-Rizzoli, 2000): Whereas in purely wind-driven cases the equatorial upwelling was supplied from both northern and southern subtropical thermoclines (with the off-equatorial position of the ITCZ mainly reflected in an enhanced TC-pattern in the northern hemisphere), superposition of an interhemispheric MOC transport of 15 Sv effectively suppressed the northern connection.

It hence appears that certain differences in STC patterns between various GCM realizations, e.g., between the CME and DYNAMO (and other, more recent) model studies can be rationalized in terms of differences in the strengths of the MOC: the early CME cases were characterized by a very weak (only about 5 Sv) interhemispheric transport, roughly half of the transport in more recent GCM simulations. An examination of some other (later) CME cases demonstrates, however, that an increased MOC transport alone is not sufficient to break the northern STC; e.g., Böning and Bryan (1996) show a case (exp. K13-7) where a doubling of the interhemispheric transport has relatively little effect on the equatorial upwelling: the bulk of the additional, northward flow in that case occurs at intermediate levels below 500m. The (model) factors governing the partitioning in thermocline and sub-thermocline flows thus remains an important question to be addressed.

(3) Pathways of Subtropical Waters: Particles and Tracers

An investigation of the warm water paths in the equatorial Atlantic on the basis of Lagrangian particle diagnostics was taken up by Blanke *et al.* (1999), using the method of Blanke and Raynaud (1997) for calculation of trajectories each associated with an infinitesimal fraction of the total (Eulerian) transport through a given section. The focus of the study was on the northward flow of SA water, as part of the upper limb of the MOC, between 10 S and 10 N. Of the total northward flow across 10 S in this model (37.4 Sv), 17.4 Sv were transmitted to 10 N, constituting the upper limb of the MOC. The bulk of this flow across 10 S occurred at the western boundary, but only a fraction of it (especially at sub-thermocline levels), was able to transit to 10 N without being drawn eastward and upward in the equatorial current regime, i.e. without being subjected to a basin-wide recirculation (the equatorial gyre).

Recent studies have begun to elucidate the effect of this recirculation on the spreading of passive tracers or temperature anomalies from the South Atlantic. Analysis of various passive tracers (including CFCs and an idealized tracer “age”, measuring the time since subduction in the South Atlantic) in recent, eddy-resolving Atlantic model experiments (“FLAME” - Family of Linked Atlantic Model Experiments) at IFM Kiel, suggests a rather rapid, O(5 years) spreading of signals via the NBUC towards the equator. Once in the equatorial regime, however, tracer signatures are drawn out across the zonal extent of the basin and concentrations become effectively diluted. A similar behaviour also characterizes the fate of remotely-generated thermocline anomalies: as shown by Lazar *et al.* (2000), a temperature anomaly imposed in the subduction region of the South Atlantic thermocline, largely fades away upon reaching the equator after 6-8 years.

The first dedicated analysis of water mass pathways between the subtropical and tropical Atlantic Ocean has been provided by Malanotte-Rizzoli *et al.* (2000), based on evaluations of the Bernoulli function on isopycnal surfaces of a non-eddy resolving GCM, and trajectories of floats injected along various subtropical latitudes and moving with the annual-mean velocity field. While the analysis for the South Atlantic confirmed that nearly all exchange passes through the western boundary, for the North Atlantic it suggested that some interior exchange window may exist for surfaces outcropping at 20-22 N: floats on these surfaces appeared to reach the EUC through a broad exchange pathway east of the NBC. Further sensitivity experiments by Inui *et al.* (2000) suggested a dependence of this communication window on the wind stress climatology.

A conspicuous aspect of the zonally-integrated overturning patterns in many high-resolution GCM solutions, is a northern STC cell not extending toward the equator, but comprised between a downwelling regime around 16-18 N and an upwelling limb concentrated at 10-12 N; hence at least in the zonal integrals the connection to the equatorial upwelling regime appears negligible. An inspection of various GCM results (including CME, DYNAMO, FLAME, CLIPPER, and others), reveals the latitudinal extent of this cell to be fairly robust, apart from quantitative details that appear sensitive to choices of resolution and lateral mixing. Preliminary results (J. Kröger, Kiel) of float trajectories released in the NEC at 25 W, 17-20 N, and moving with the time-dependent, eddying velocity field of a FLAME model case, suggests this shallow overturning cell to be associated with a cyclonic recirculation pattern: after reaching about 10 N over a broad interior "window", the bulk of the floats are carried eastward and, after upwelling near the eastern coast, swept back to the northwest in the surface Ekman layer. The float behaviour is consistent with tracer signatures in this model which give no indication of subduction pathways from the northern subtropics ventilating the equatorial regime south of about 5 N (i.e., the NECC).

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V.VIII Coupled Aspect of Atlantic Ocean-Atmosphere Variability

(Shang-Ping Xie)

Empirical Studies

The tropical Atlantic and Pacific share many common climatological features, such as a cold tongue and eastward shoaling of the thermocline on the equator, prevailing easterly trade winds, and a northward displaced ITCZ (Figure 1). Given that the solar radiation is both zonally uniform and symmetric about the equator, the establishment of these climatological features involves ocean-atmosphere-land interaction.

Because of the east-west asymmetry in the climatology, the Atlantic also sees an equatorial mode of SST variability much like the Pacific ENSO. This Atlantic Nino mode is captured by a rotated EOF analysis by Ruiz-Barradas *et al.* (2000). When an Atlantic warm event occurs, atmospheric deep convective heating increases near the equator, inducing anomalous westerlies in the western equatorial Atlantic (Figure 2). In response to this change in zonal wind, the thermocline deepens in the east, helping sustaining/amplifying the SST warming. This is exactly the same Bjerknes feedback that gives rise to the ENSO in the Pacific. But because of the small zonal size of the Atlantic Ocean, this equatorial coupled mode is likely to be damped, excited by stochastic forcing (Zebiak, 1993).

How to characterize the rest of tropical Atlantic SST variability is under much debate and a subject of active research. Based on EOF analysis, one school of thought contends that there is a dipole mode that is anti-symmetric about the annual-mean ITCZ (Servain, 1991; Nobre and Shukla, 1996; Figure 3). This SST pattern is known to be associated with rainfall variability in the Northeast Brazil and Sahel. The other school of thought argues that the dipole mode is unphysical and an artifact of EOF analysis (Houghton and Tourre, 1992; Enfield and Mayers, 1997; Mehta, 1998) by pointing out the fact that the SST anomalies are organized on one side of the equator, and do not significantly correlated across the equator (Figure 4). Some studies show the interhemispheric SST correlation may change sign according to time scales; SST anomalies tend to be negatively (positively) correlated between the tropical North and South Atlantic on the decadal (interannual) time scale (Tanimoto and Xie, 1999; Enfield *et al.*, 1999). But a study by Mehta (1998) finds no interhemispheric coherence at any frequency band.

Results from empirical studies agree that there are two or more modes of SST variability in the tropical Atlantic, but they disagree on what these modes are. The lack of interhemispheric correlation can be interpreted as the evidence for independent tropical North and South Atlantic modes. Equally plausible is the argument that a monopole and dipole, of equal amplitudes, interfere with each other, causing an apparent lack of correlation across the equator. Given the limited length of observation record, it is difficult to draw form conclusions as to which decomposition, northern and southern modes vs. monopole and dipole modes, is physical. Such a determination may be further complicated by the possibility that the spectra of these modes are red and do not have distinct peaks.

Coupled Modelling

On the modelling front, much progress has been made in understanding the cause of off-equatorial SST variability and the roles of ocean-atmosphere interaction. Carton *et al.* (1996) carry out a set of OGCM experiments by modifying the atmospheric forcing through wind stress and wind-induced evaporation. While equatorial variability is largely due to wind stress-forced thermocline variations, off-equatorial SST variability, particularly that in the interhemispheric SST gradient, is caused mainly by wind-induced evaporation (Figure 5).

Chang *et al.* (1997) suggest that the variability of interhemispheric SST gradient involves air-sea interaction and invoke a wind-evaporation-SST (WES) feedback to explain Carton *et al.*'s OGCM results in a coupled context. A dipole pattern of SST anomalies like the one in Figure 2 induces southerly cross-equatorial winds, which acquire an easterly (westerly) component because of the Coriolis force in the Southern (Northern) Hemisphere. These anomalous winds enhance (weaken) the background trade winds and hence the evaporative cooling south (north) of the equator, and these wind-induced changes in evaporation act to amplify the initial SST anomalies. Chang *et al.* further couple an empirical atmosphere model with two separate ocean models and show that the dipole mode is dominant in these coupled models if the thermodynamic feedback is strong enough (Figure 6). They suggest that the advection by the cross-equatorial North Brazil Current is responsible for the phase change and sets the decadal time scale for the dipole oscillation in their models.

An alternative phase-changing mechanism is proposed by Xie (1999). In addition to an in-phase relation between wind and SST anomalies that gives rise to the positive WES feedback, the maximum westerly wind anomaly is located equatorward of the maximum SST anomaly, an effect of the geostrophy. This meridional phase shift creates a tendency for the coupled anomalies to propagate toward the equator (Figure 7b). This tendency of equatorward propagation is slowed down by the mean poleward Ekman advection (Figure 7a) under the prevailing easterly trade winds.

The small amplitudes of tropical Atlantic variability (SST rms ~0.4 C) suggest that all coupled modes may be damped and thus need to be excited by external forcing. In response to mutually uncorrelated random forcing in the subtropical North and South Atlantic poleward of 25° latitude, SSTs in a coupled WES model varies at slow time scales in an anti-symmetric mode (Figure 8; Xie and Tanimoto, 1998). When the dipole mode is only weakly damped, the spectrum of the cross-equatorial SST gradient shows a peak at its intrinsic frequency. In the presence of strong damping, however, the spectrum becomes red. Even in this strongly damped case, the coupled dipole, as the least damped mode of the system, still leaves strong marks in the time-space structure of model variability. First, the spectral power starts to level off around the intrinsic frequency of the free dipole mode. Second, the coupled mode gives rise to the characteristic dipole structure even if the northern and southern forcing is mutually uncorrelated.

In a general sense, there are two modes of ocean-atmosphere interaction in the tropical oceans. Within the equatorial upwelling zone where the coupling between the thermocline depth and SST is strong, the Bjerknes feedback involving zonal interaction of equatorial wind, thermocline depth and SST dominates, giving rise to ENSO in the Pacific and Atlantic. Off the equator, on the other hand, the general downwelling renders the ocean dynamics ineffective in changing SST, leaving surface heat flux to play an important role in SST variability. In a large ocean basin like the Pacific, the ENSO mode that is strongly trapped on the equator dominates. The Bjerknes feedback weakens in a small domain like the Atlantic (Zebiak, 1993), allowing a weak WES dipole mode to become the leading mode (Figure 9; Xie *et al.*, 1999). In the particular calculation shown in Figure 9, the equatorial Bjerknes mode has a slightly smaller than but comparable growth rate with the dipole mode, suggesting that they may coexist.

Atmospheric Response

In the simple and intermediate coupled models, the existence of a dipole mode can be confirmed by suppressing ocean dynamic effects and hence the Bjerknes feedback. In fully coupled GCMs, such sensitivity experiments are much more difficult to carry out if not impossible. The lack of interhemispheric SST correlation in some coupled GCMs has led to a speculation that the observed wind anomaly pattern in Figure 3 may act as a one-way forcing to the ocean, but is not a response to the SST dipole (Dommenget and Latif, 2000). If true, the dipole is merely a passive response to some fortuitous arrangement of winds.

Attempts to determine the atmospheric response to a prescribed SST dipole with atmospheric GCMs have yielded mixed results. Dommenget and Latif (2000) claim that the ECHAM3 does not show significant response to a SST dipole. Sutton *et al.* (2000) obtain similar results in the HadAM1 while noting a significant response in the ITCZ and cross-equatorial wind. Chang *et al.* (2000) report significant response in both meridional and zonal wind components within a latitudinal band between 10S and 10N in the CCM3. The apparent disagreement among the GCMs (Figure 10) can arise from differences in model physics, but may also be partially due to the difficulty in obtaining stable statistics in the presence of strong internal chaotic variability of the atmosphere.

Okumura *et al.* (2000) prescribe a dipole pattern of SST anomalies (SSTAs) in the tropical Atlantic that does not change with time, and carry out long integrations of an atmospheric GCM until stable statistics of model anomaly fields are obtained. They find that the Atlantic ITCZ shifts into the warmer hemisphere, with the trade winds relaxing (intensifying) on a tropical-wide scale over the warm (cold) SSTAs, much as in observations and in support of the WES feedback (Figure 11a).

The existence of this WES feedback is further confirmed by results from a coupled model with the same AGCM as its component. A clear north-south seesaw due to the WES feedback is clearly visible in a latitude-time section of SSTAs (Figure 12; Xie and Saito, 2000). The dominance of the dipole mode in this coupled model is partially due to the suppression of the thermocline feedback in the intermediate ocean model.

The atmospheric response to a tropical SST dipole is not limited to the tropics, but also include a barotropic component that extends into the extratropics. When coupled with an interactive ocean mixed layer of 50 m deep, the AGCM response closely resembles a composite of observational data based on an index of cross-equatorial SST gradient (Figure 11), both having a projection onto the NAO. In response to a negative tropical SST dipole, the Azores high centered at 40N strengthens, and so does the Icelandic low albeit with a smaller amplitude. The winter cold surges in the eastern US and Canada weaken substantially. A pair of SSTA centers appear off the US and Canadian coast, respectively, associated with changes in the speed of prevailing westerlies and forming the extratropical portion of the North Atlantic tripole.

Astride the trade and westerly wind regimes, the Azores high appears to be an important bridge between the tropical and extratropical North Atlantic. Okumura *et al.*'s results indicate that the tropical SSTAs can affect the extratropical North Atlantic through atmospheric teleconnection. Watanabe and Kimoto (1999) and Robertson *et al.* (2000) obtain similar results, albeit each based on one single integration forced by global observed SSTs. This Azores bridge appears not to be a one-way process, and can act the other way around. When forced by observed wind poleward of 20° latitude, a coupled WES model can reproduce the observed time series of interhemispheric differences in SST and zonal wind (Figure 13). Thus, the extratropical and tropical Atlantic may be fully interactive through the Azores variability.

It is even less clear whether the extratropical North Atlantic supports coupled ocean-atmosphere modes. The NAO appears to show a significant response to the North Atlantic SST tripole (Rodwell *et al.*, 1999; Mehta *et al.*, 2000), and may thus provide a weak feedback onto the SSTAs. On the other hand, other studies suggest that the tripole pattern is mainly determined by the dominant atmospheric stationary eddy arrangement or NAO, and results from the passive ocean response to stochastic forcing by the NAO (Delworth and Mehta, 1998; Seager *et al.*, 2000). On the other hand, the ocean gyre adjustment (Grotzner *et al.*, 1998; Marshall and Czaja, 2000), advection by the North Atlantic Current (Sutton and Allan, 1997), and thermohaline circulation (Delworth *et al.*, 1993) have also been implicated as important in generating low-frequency North Atlantic SST variability.

Implications for STC Studies

The study of the Atlantic STCs is just beginning and its role in the climate and its variability needs to be determined. A recent simulation of the Atlantic circulation indicates that the bifurcation latitudes of the western boundary currents are 12N and 15S, respectively (Inui *et al.*, 2000; Figure 14). Changes in trade winds associated with off-equatorial modes of SST variability (e.g., Figure 3) can affect the strength of the STCs and thus the transport of the cold thermocline water into the tropics. Kleeman *et al.* (1999) suggest that this $v'\bar{T}$ mechanism leads to decadal ENSO modulation in the Pacific.

The $\bar{v}T'$ mechanism of Gu and Philander (1997) may also operate in the Atlantic where SST variability in the subtropics is as strong as on the equator. Ragu (2000) show that the subducted thermal anomalies are strongly dissipated in the western boundary region before reaching the equator much as in the Pacific (Schneider *et al.*, 1999; Nonaka and Xie, 2000).

There are a number of noteworthy differences in the STCs between the Pacific and Atlantic. First, the northernmost latitude where the subducted water feeds into the equatorial Atlantic is much more southward than in the North Pacific because of the MOC. This determines that the subducted thermal anomalies that can affect the equatorial SSTs through the Gu and Philander mechanism must be of tropical origin in the Atlantic. Second, whereas the tropical dipole mode appears ideal to supply SSTAs for subduction and force the STC strength change through its wind variability, it is anti-symmetric about the ITCZ. Thus, STC-related anomalies from the two hemispheres may cancel each other when they reach the equator. However, the Atlantic climatology in general and its ITCZ in particular are not symmetric about the equator, leaving room for the STCs to play a role, but this role appears to be more complicated than in the Pacific.

Summary

a. There is an equatorial ENSO-like mode in the Atlantic due to the Bjerknes feedback.

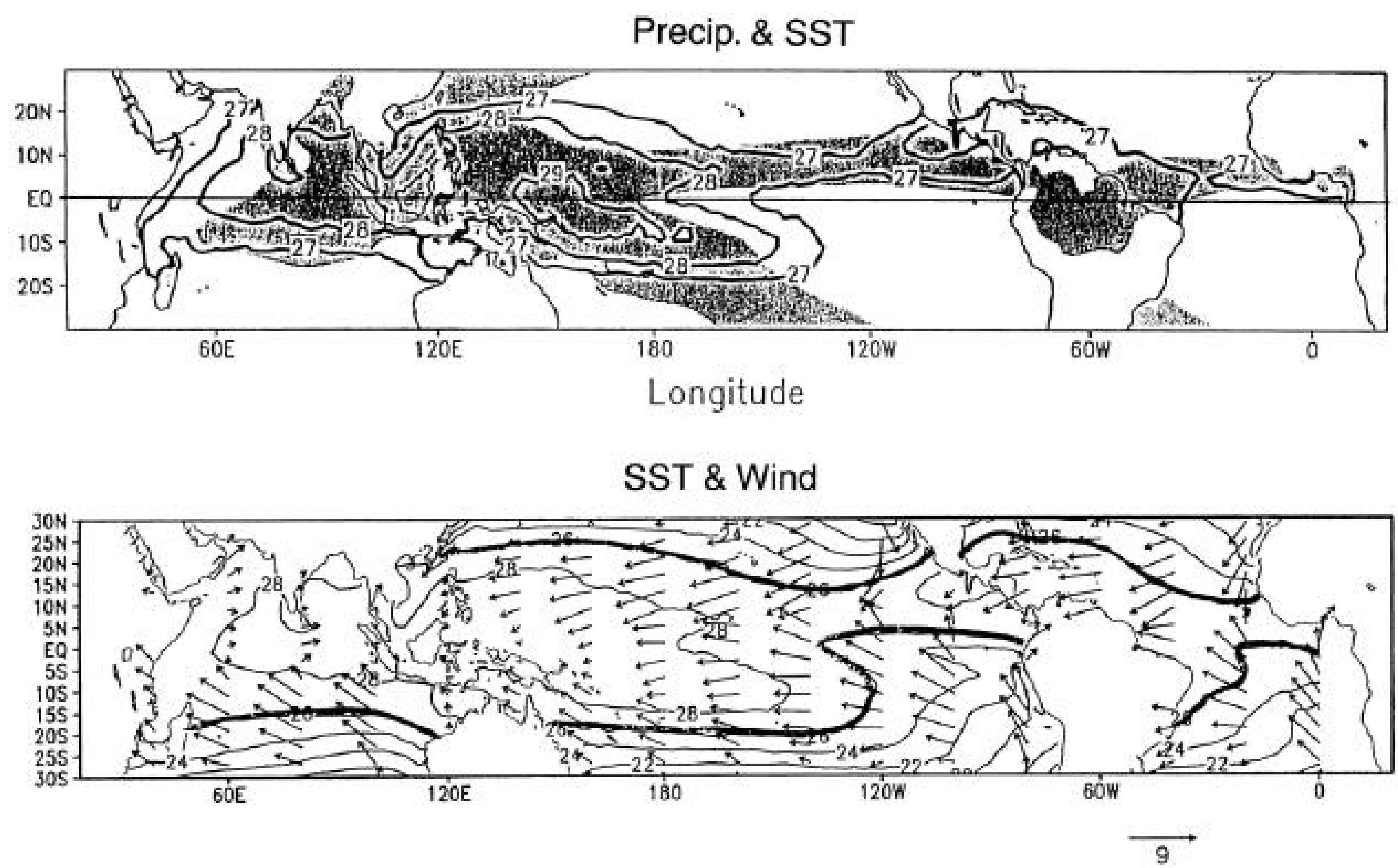
b. Empirical analysis of observational data does not yet give conclusive answers as to how to characterize the tropical Atlantic SST variability.

c. Some coupled models suggest that the cross-equatorial SST gradient is controlled by a coupled mode of ocean-atmosphere interaction that is anti-symmetric about the equator.

d. Atmospheric response to a prescribed SST dipole disagrees among AGCMs, although they agree on the response to ENSO-like equatorial SST anomalies. This disagreement calls for a better understanding of atmospheric response to SST anomalies over cold ocean surface, where the planetary boundary layer processes are likely important.

e. The Azores high can potentially act as a two-way bridge that allows the tropical and extratropical North Atlantic to interact.

f. The roles of ocean dynamics including the STCs are poorly understood in the tropical Atlantic and need further investigations.



Precip. & SST
20N
10N
EQ
10S
20S
60E
120E
180
120W
60W
0
Longitude
SST & Wind
30N
25N
20N
15N
10N
5N
EQ
5S
10S
15S
20S
25S
30S
9

Figure 1

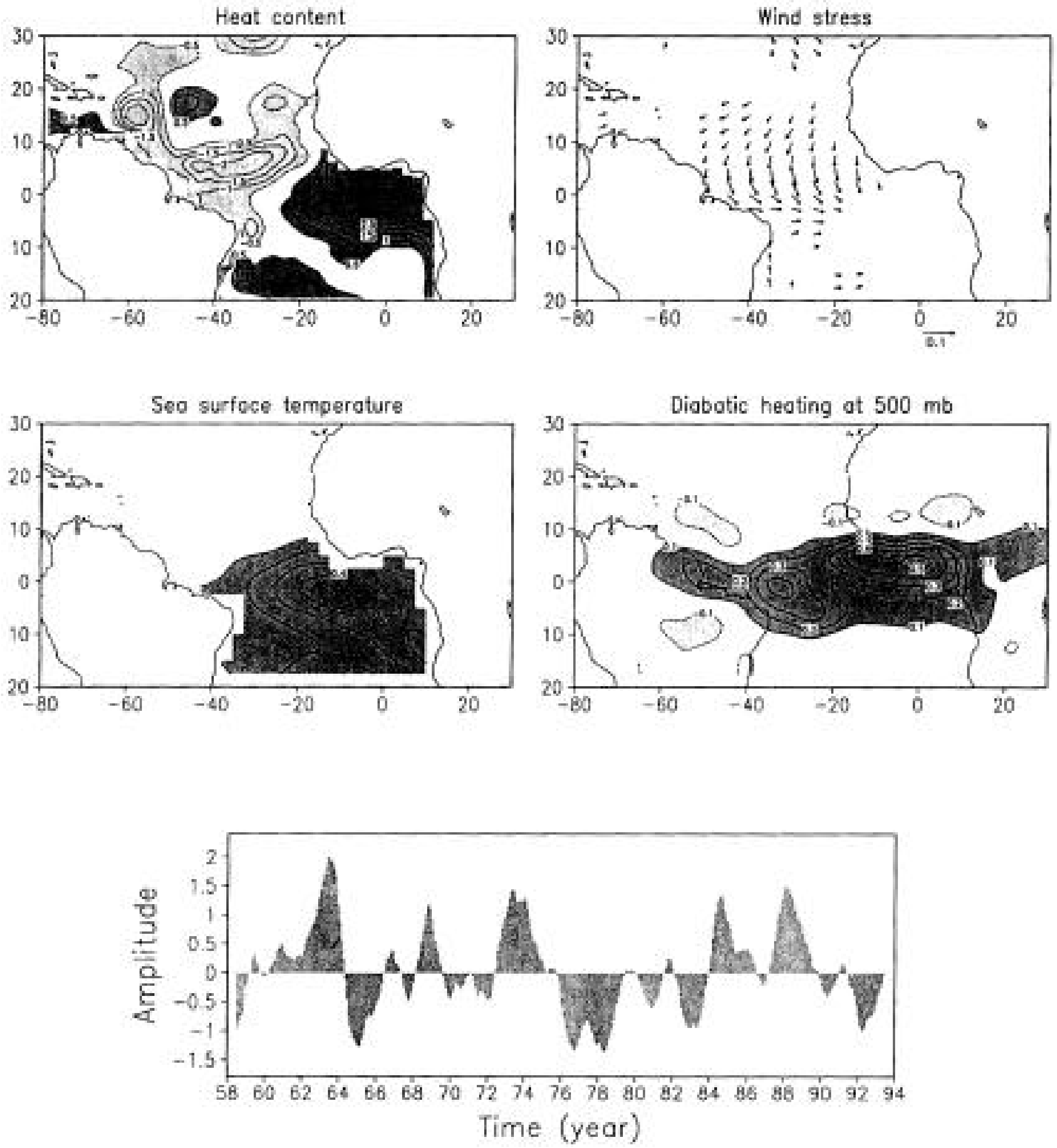


Figure 2: Atlantic Niño mode and its associated principal component coefficient (PCC) from a 5-variable rotated combined analysis. Uppper left: hc’(10^8 x J/m^2), upper right: τ’(dyn/cm^2); middle left: sst’(°C), middle right: q’ at 500 mb (°C/day); lower: PCC smoothed with a 12-month running mean. Only wind stress anomalies larger than 0.02 dyn/cm^2 are shown. Explained variance is 4.9%. Dark/light shading denotes positive/negative anomalies.

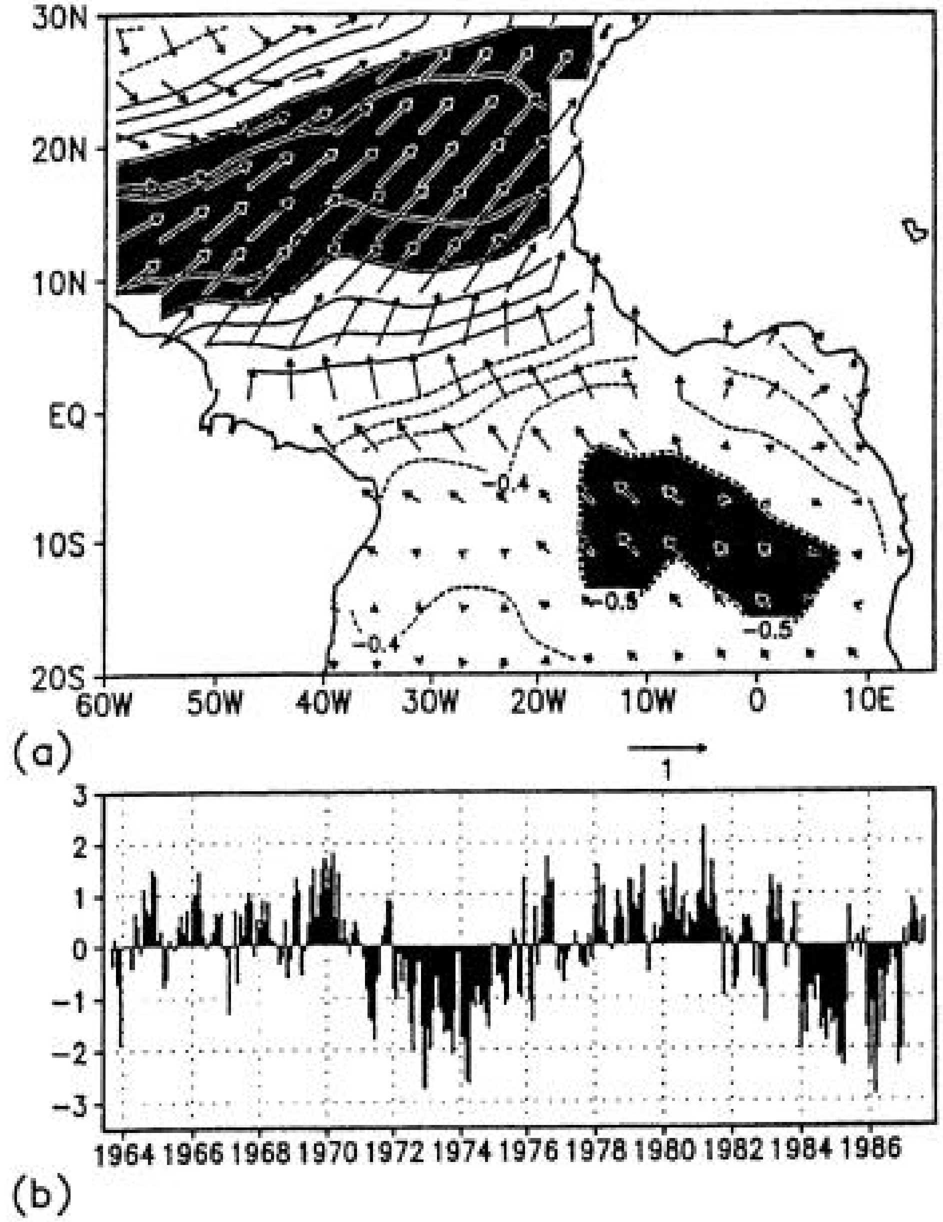


Figure 3: The first joint EOF of SST, τ^x, and τ^y monthly anomalies over the tropical Atlantic from September 1963 to August 1987; (a) the spatial pattern and (b) the associated coefficient time series. The contours represent the SST loadings; contour interval is 0.1; negative contours are dashed; values greater than 0.4 or lower than -0.5 are shaded; the zero contour is not drawn. The arrows represent the vectorial sum of τ^x and τ^y loadings; the vectors are scaled accordingly to the arrow plotted at the lower right side of the upper panel.

c) T Correlation

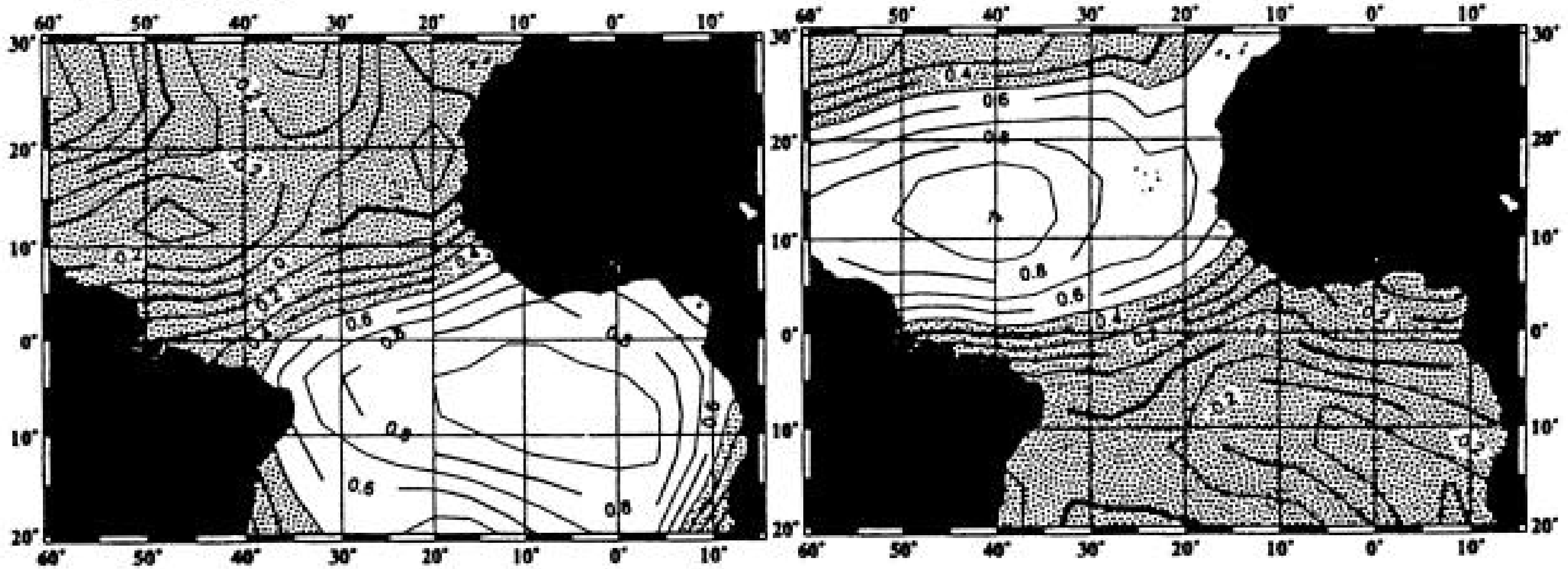


Figure 4: (a) Percent of SST variance explained by unrotated PC modes 1 and 2. (b) Percent of SST variance explained by varimax rotated modes 1 and 2. (c) Point correlation map of SST with repect to 7°S, 7°W (left) and 13°N, 39°W (right). Stippling indicates areas not statistically significant at the 95% confidence level.

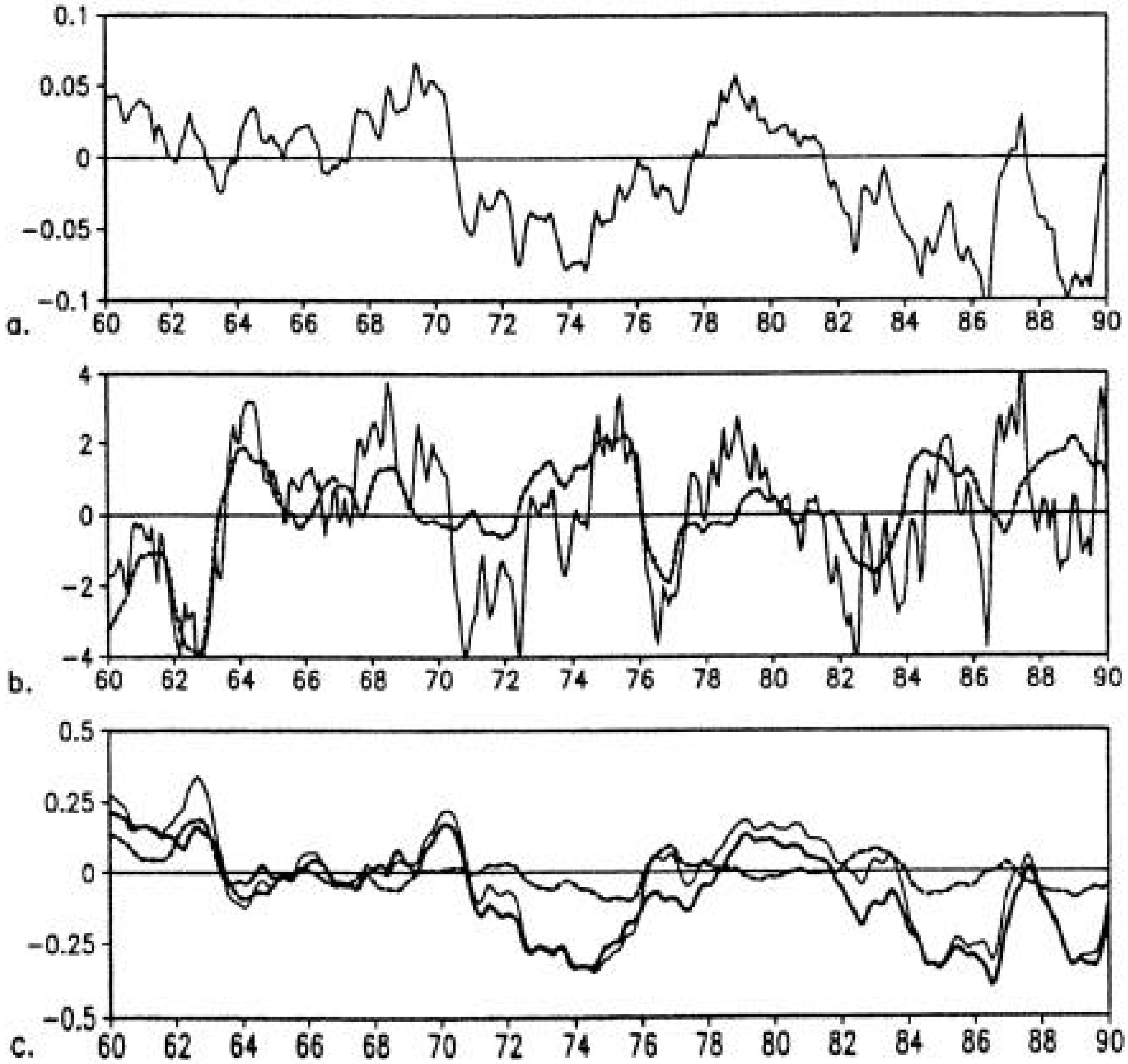
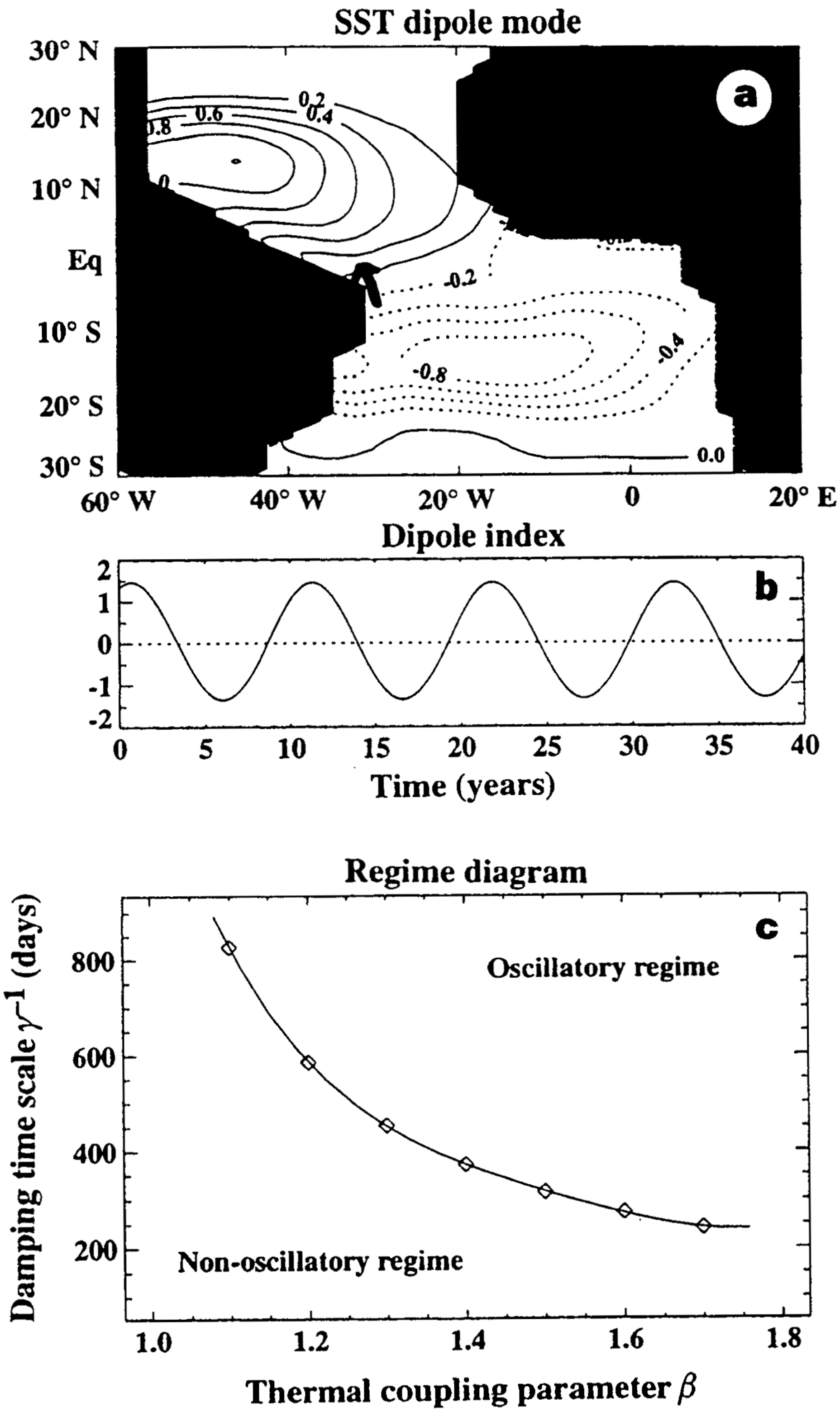


Figure 5: Time series of 12-month smoothed interhemispheric differences of variables from the simulation. (a) Surface wind stress anomaly, (b) heat content anomaly, and (c) SST anomaly. Results from the simulation are shown with a solid line. Results from experiment 3 and experiment 4 are also displayed. In experiment 3 the effect of interannual variations in latent cooling due to interannual fluctuations in wind speed are eliminated (short dashed line). In experiment 4 interannual variations in latent cooling are retained, but the contribution of the interannual variations in the winds in driving anomalous ocean circulation is eliminated (long dashed line). For all other calculations climatological winds are used.



SST dipole mode
a
30° N
20° N
10° N
Eq
10° S
20° S
30° S
60° W
40° W
20° W
0
20° E
0.2
0.4
0.6
-0.2
-0.4
-0.8
0.0
Dipole index
b
2
1
0
-1
-2
0
5
10
15
20
25
30
35
40
Time (years)
Regime diagram
c
Oscillatory regime
Non-oscillatory regime
Damping time scale γ^{-1} (days)
800
600
400
200
1.0
1.2
1.4
1.6
1.8
Thermal coupling parameter β

Figure 6

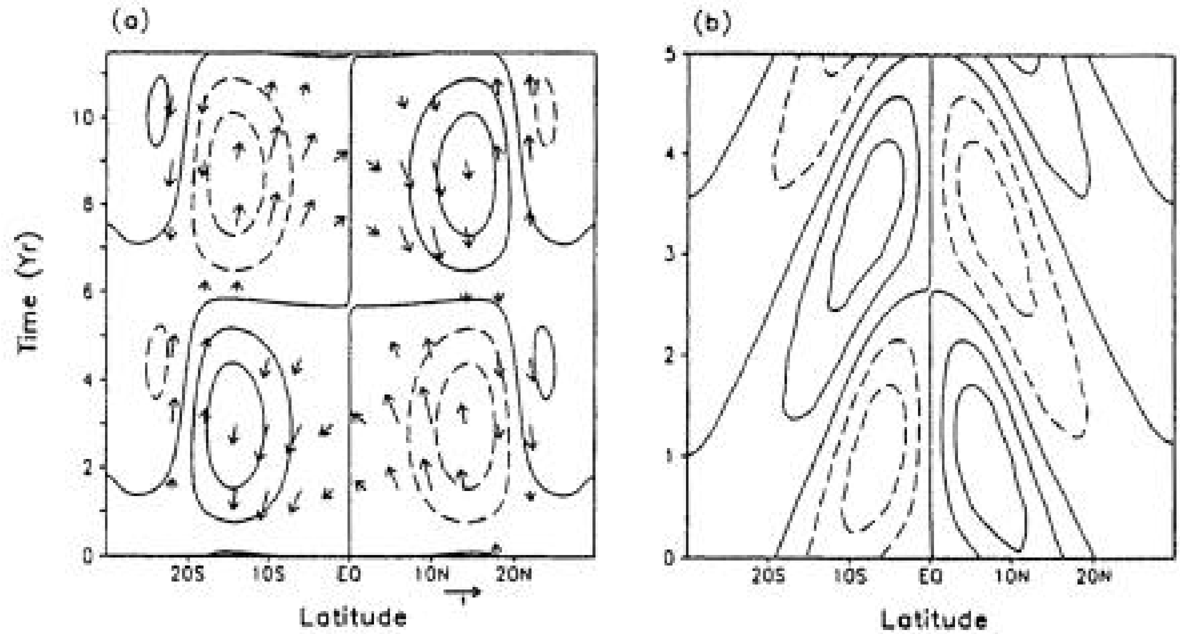


Figure 7: Latitude-time section of sea surface temperature (contours with the negative ones dashed) and surface wind velocity (vectors) associated with the free WES mode in models (a) with and (b) without the mean Ekman advection. The exponential growth trend is removed. Easterly wind appears as an upward vector.

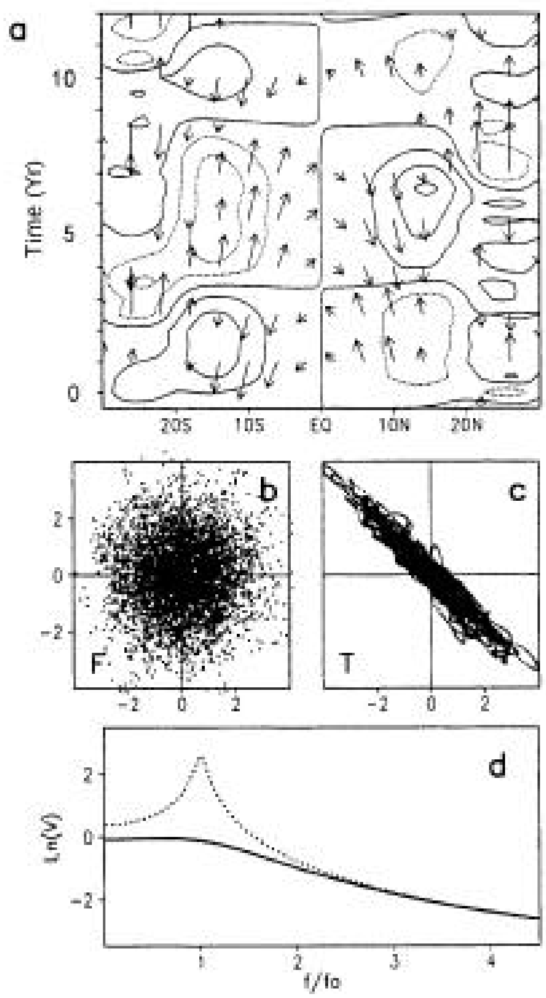


Figure 8: **a**, Latitude-time section of SST anomaly (negative dashed; contour interval; 0.75°C) in a coupled ocean-atmosphere model ($b=b_0$) forced by extratropical white-noise wind disturbances. The forcing has a fixed spatial structure reaching a maximum at 30° and vanishing at 20°. Anomalous wind velocity (ms^{-1}) is plotted with an upleft vector denoting southeasterly wind. Scatter plots of **b**, white-noise forcings applied to the northern and southern extratropics and **c**, model SSTs at 15N and 15S (all data are normalized with their variances). **d**, Spectra of meridional wind velocity (V) at the equator calculated from 500-year integrations with weak ($b=0.6b_0$; dotted) and realistic ($b=b_0$; solid line) damping rates. Wind velocity is normalized by forcing amplitude and frequency by that of the leading WES mode, f_0 with $2\pi/f_0 = 11.5$ years.

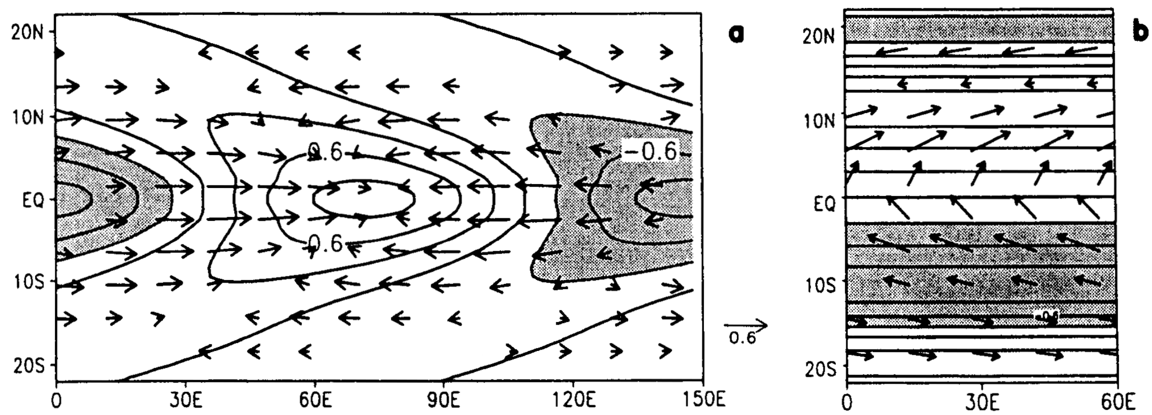


Figure 9: Dominant modes at **a**, the Pacific and **b**, the Atlantic wavelengths in the coupled model. SST in contours (<-0.3°C shaded) and surface wind velocity (m/s) in vectors.

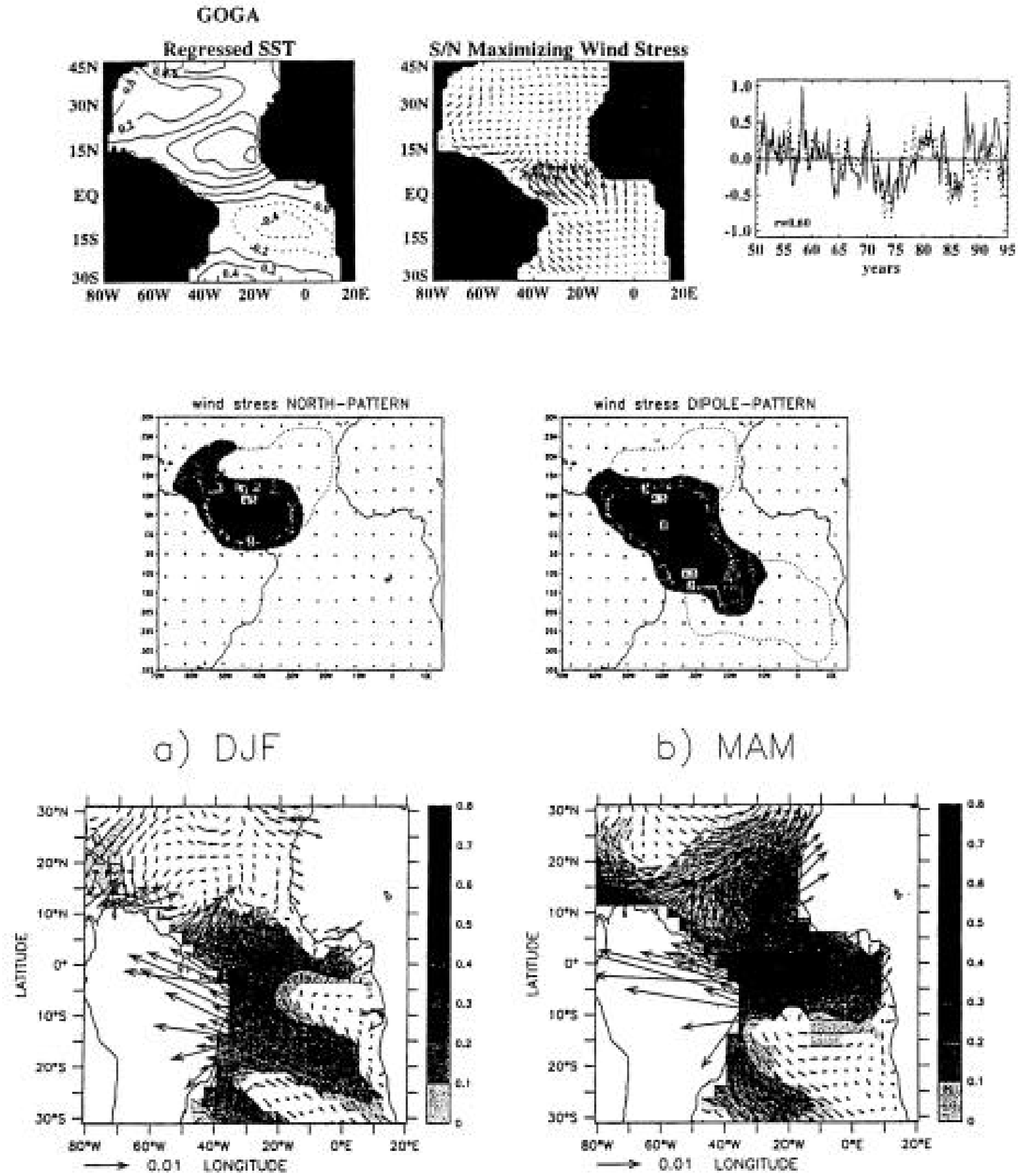
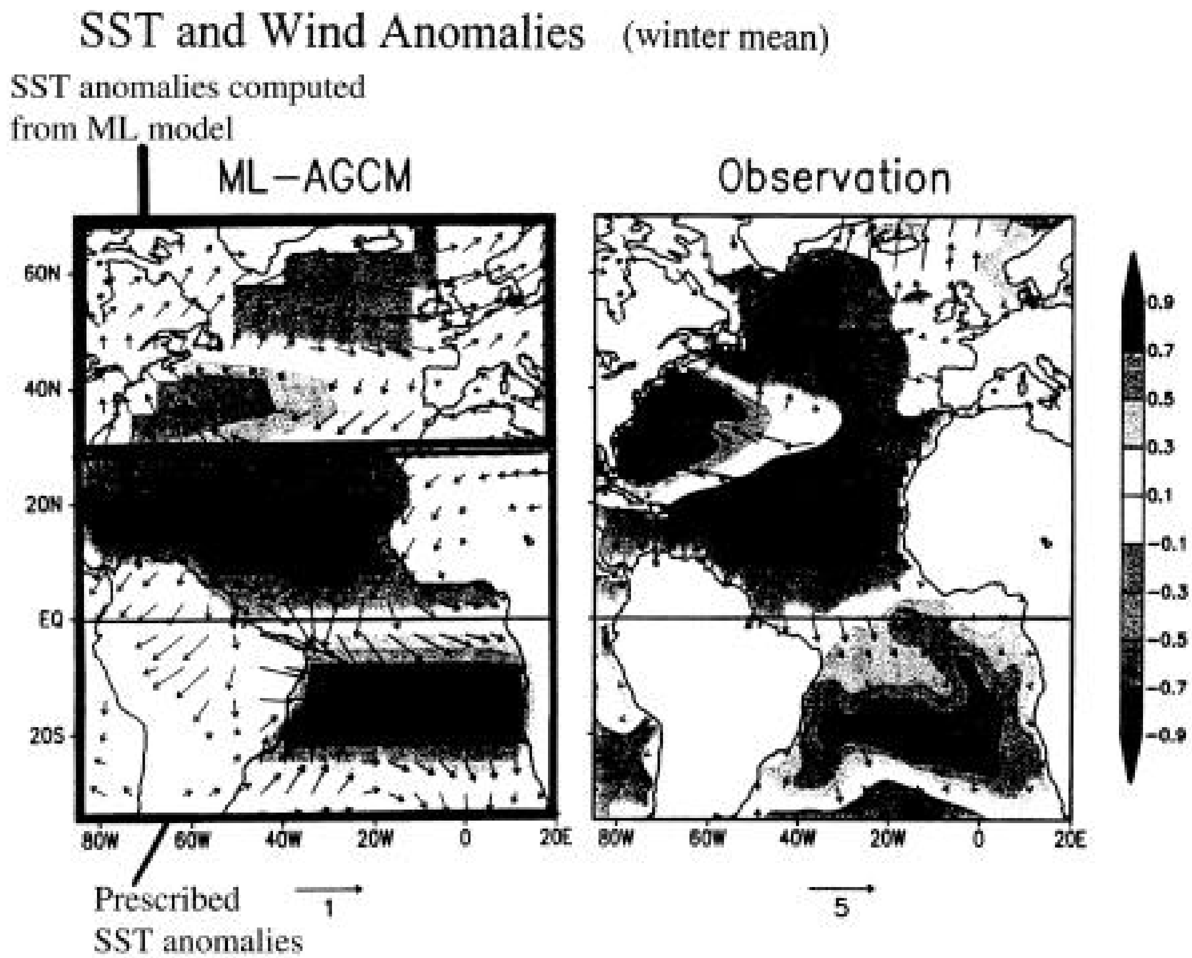

GOGA
Regressed SST
S/N Maximizing Wind Stress
45N
30N
15N
EQ
15S
30S
80W 60W 40W 20W 0 20E
1.0
0.5
0.0
-0.5
-1.0
50 55 60 65 70 75 80 85 90 95
years
wind stress NORTH-PATTERN
wind stress DIPOLE-PATTERN
a) DJF
b) MAM
30°N
20°N
10°N
0°
10°S
20°S
30°S
LATITUDE
80°W 60°W 40°W 20°W 0°E 20°E
0.01 LONGITUDE
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0

Figure 10



SST and Wind Anomalies (winter mean)
SST anomalies computed
from ML model
ML-AGCM
Observation
60N
40N
20N
EQ
20S
80W
60W
40W
20W
0
20E
0.9
0.7
0.5
0.3
0.1
-0.1
-0.3
-0.5
-0.7
-0.9
Prescribed
SST anomalies
1
5

Figure 11

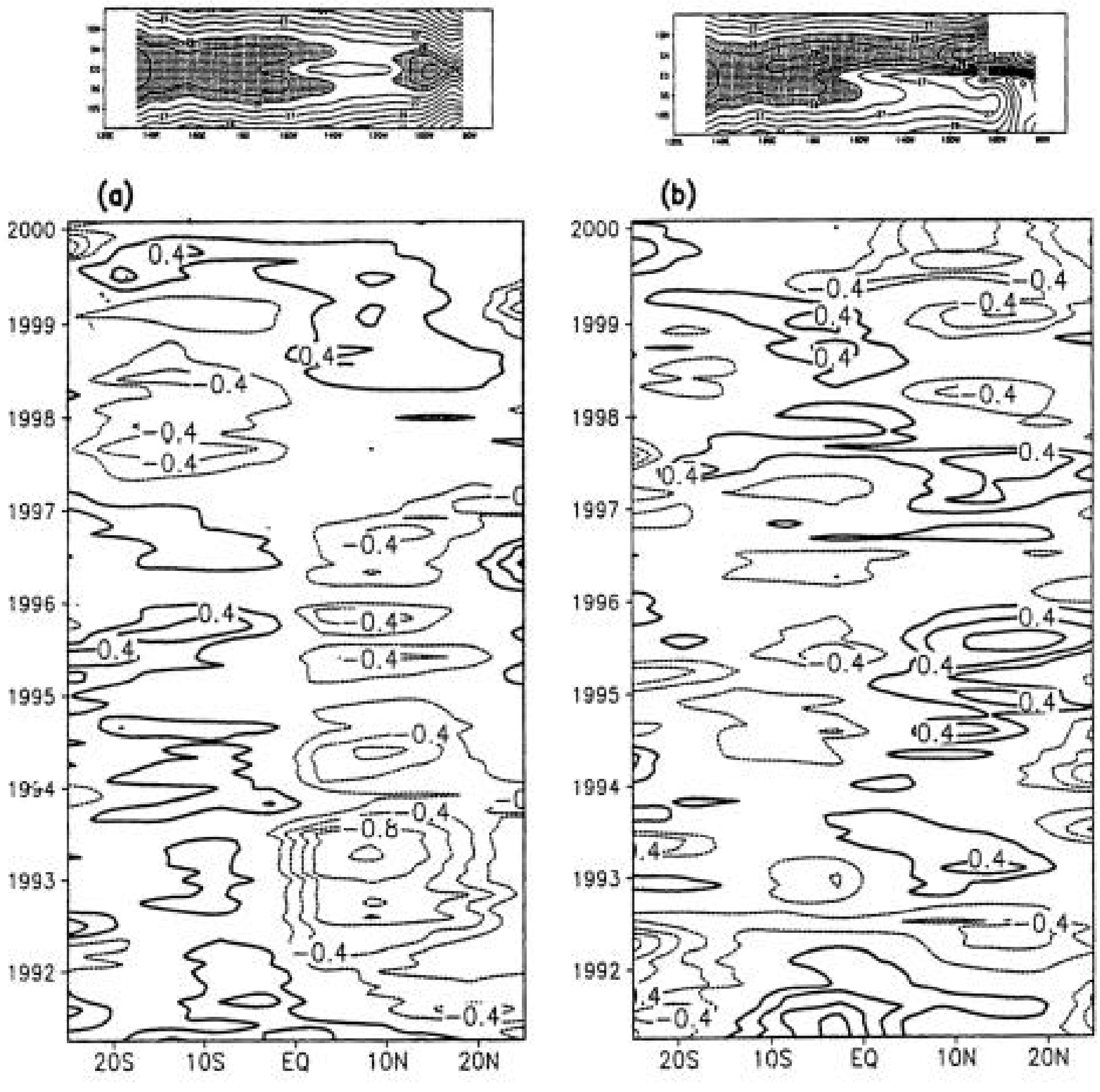

(a)
(b)
2000
1999
1998
1997
1996
1995
1994
1993
1992
20S
10S
EQ
10N
20N
0.4
-0.4
-0.8

Figure 12

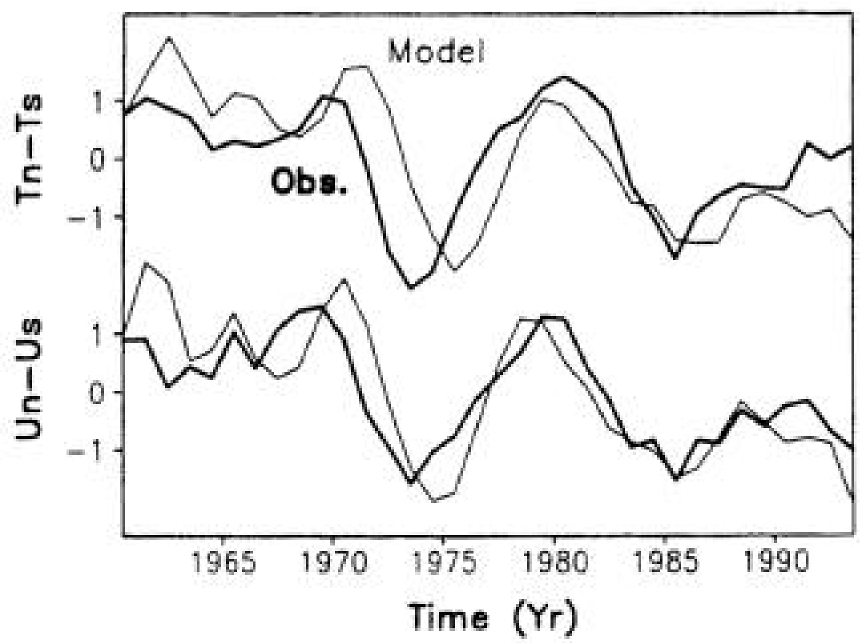


Figure 13: Observed (thick lines) and simulated (thin lines) hemispheric differences in SST (upper panel) and zonal wind speed (lower panel) between 10N-20N and 20S-10S. (All the curves are normalized with their respective variances and a three-year running mean is applied to the observed data.) Note that the model data are taken from regions free of external forcing.

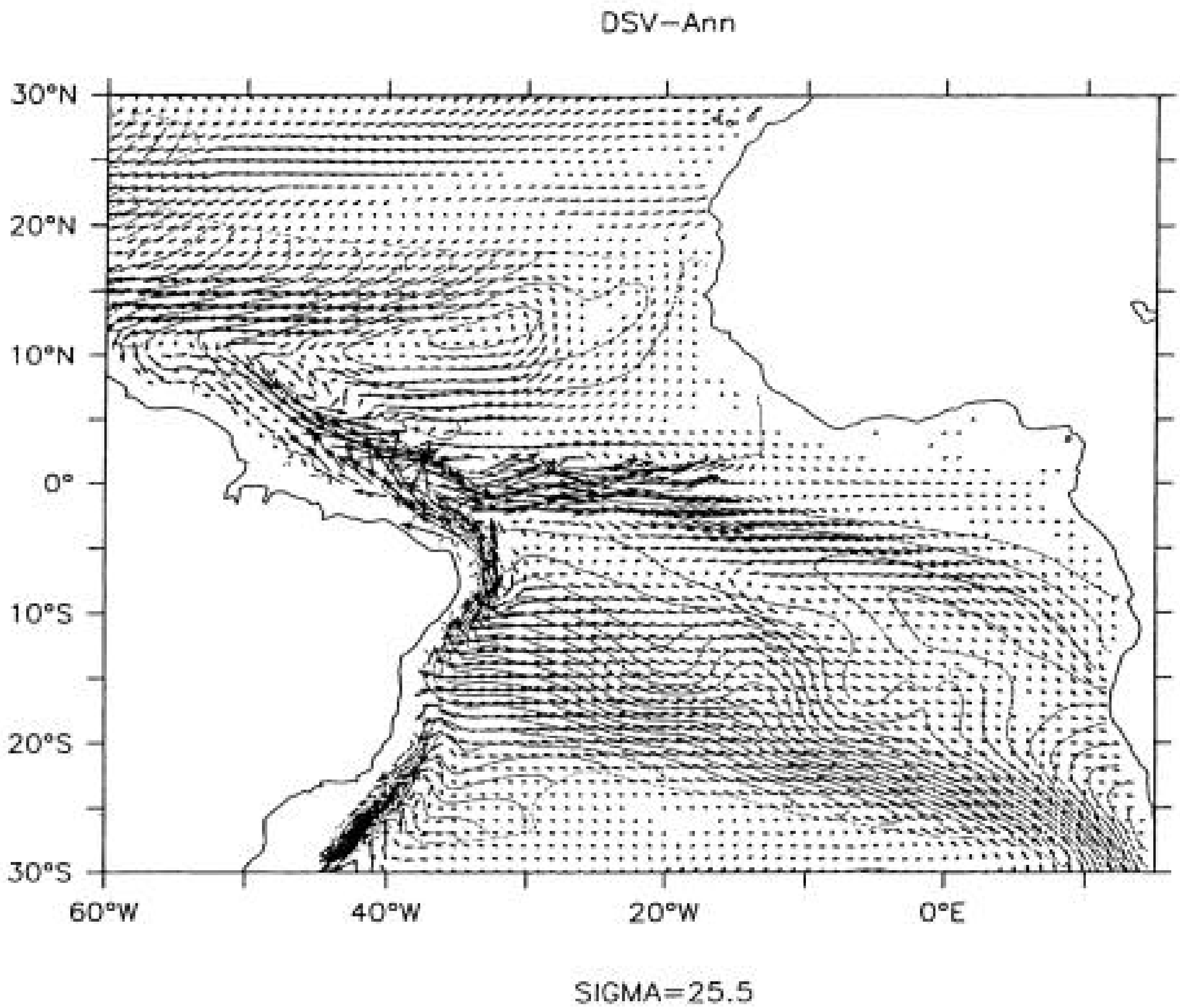


Figure 14: Bernoulli function and current vectors on the σ_θ (a) 24.5, (b) 25.0, and (c) 25.5 iso-surface with an interval of 0.5 m for DSV driven experiment.

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VI. Abstracts of Contributed Posters

The Origin of Waters Observed Along 137°E

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Using the World Ocean Atlas data set, we examine the origins and flow paths of waters observed along 137°E section in the western North Pacific. A method is developed to trace waters from 137°E back through the subtropical gyre to their outcrops. We divide the water masses observed along this section into several zones. (1) There is a zone of waters that connect directly to areas of surface subduction in the subtropical gyre. This is chiefly the water mass known as North Pacific Tropical Water, but includes most of the waters of the mid-latitude thermocline. Working backward along geostrophic streamlines, we find the locations of the outcrops of these waters. Subducted waters are aged using this technique and found to be between 0.5 and 35 years old by the time they reach 137°E. For this subducted regime, waters on a given isopycnal observed along 137°E increase in age with decreasing latitude, with waters at the southern end of the section being 2-3 times older than waters at the northern end. It is found that subducted water masses are homogenized by the time they reach 137°E. That is, the originally subducted waters have a wide variation in _-S characteristics. Whereas by the time they reach 137°E, they form a coherent water mass with a tight _-S relation. It is shown that isopycnal mixing is not the cause of this homogenization, leaving diapycnal mixing as the major player. (2) There is a recirculating zone, where waters do not have any direct contact with the surface. Some of these recirculating waters, particularly the North Pacific Intermediate Water are strongly influenced by surface processes outside the subtropical gyre, but are not subducted by Ekman pumping. (3) There is North Pacific Subtropical Mode Water, which has

direct contact with the surface, but the subduction process is enhanced by buoyancy forcing. (4) There is a seasonal thermocline, where waters are strongly influenced by surface heating and cooling, and isopycnals can disappear during the winter. (5) There are a variety of tropical upper and intermediate waters interleaving in complex ways and having varied origins and flow paths. We use these generalizations to characterize water mass variability observed in the PR-1 hydrographic section along 137°E.

The Equalant Cruises in the Equatorial Atlantic (1999-2000)

B. Bourles, C. Andrie, Y. Gouriou, Y. DuPenhoat, G. Eldin and S. Arnault

In the framework of Clivar, the Equalant cruises are part of the French ECLAT (Etudes Climatiques dans l'Atlantique Tropical) program, and mainly consist in the realization of five meridional sections across the equatorial Atlantic. These cruises have been carried out on board R/V Thalassa from July, 13th to August, 21st, 1999 for the western and central sections (350W, 230W and 100W), and from July,24th to August, 21st, 2000 for the eastern sections (100°W, 0°E, 60°E). The main scientific objectives of the EQUALANT project are to study:

1) the equatorial dynamic variability in the surface and subsurface layers (and the fate of the countercurrents in the east),

2) the variability of the large scale and low frequency thermohaline circulation in the equatorial belt , and

3) the deep circulation in the equatorial band (deep equatorial jets, Intermediate Counter-Currents, Deep Western Boundary Current).

During these cruises, hydrological (CTD-O2 casts with 24 bottles) and current (Lowered Acoustic Doppler Current Profiler) profiles were carried out from the surface down to the bottom (103 profiles in 1999 and 89 in 2000), along with tracer measurements (CFCs and nutrients).

XBT and XCTD probes were launched every 20’ during zonal transits, and every 15’ between CTD casts along the meridional sections to improve sampling resolution in the thermostad. Two VM-ADCPs allowed to get current profiles between the surface down to 700 m depth all along the trackline.

Very first results are shown and discussed:

1) Salinity and temperature vertical sections obtained from XCTD/XBT profiles along the transits between meridional sections, that show meandering structures;

2) Vertical sections of the zonal component of the current velocity and salinity at 350°W, 230°W, 100°W in 1999 and 100°W in 2000, that show the eastward evolution of undercurrents and the difference at one year interval between current structures at 100W, and

3) Vertical profiles of the zonal component of the current velocity along with CFC-11 concentration, measured along the equator at 350°W, 230°W and 100W in 1999, showing striking correlation between deep equatorial jets and CFCs vertical distribution.

Rossby Waves in the Tropical North Pacific and Their Role in Decadal Thermocline Variability

A. Capotondi and M. A. Alexander, NOAA/CIRES Climate Diagnostics Center, Boulder, USA

1. Introduction

Thermocline variability in the Pacific has a characteristic spatial structure, as shown in Figure 1, where temperature variance at 200 m depth from both Levitus (Figure 1a) and a numerical model simulation (Figure 1b) are compared. The model is the NCAR ocean model, forced with observed atmospheric forcing over the period 1958-1997. Both modelled and observed patterns in Figure 1 indicate the presence of three areas of enhanced variance, one in the Kuroshio extension around 40°N, and the other two in the tropics, quasi-symmetric with respect to the equator, at 10°-15°N, and around 10°S, respectively. While enhanced variability in the Kuroshio extension area is expected, the areas of enhanced variance in the tropics are more puzzling. In this study we use the results from the numerical simulation in Figure 1b to examine the variability at 10°-15°N. In particular, we are interested in understanding the nature of the variability in this band, including: what forces it, is it associated with propagating disturbances and if so, do those disturbances influence other parts of the ocean?

Figure 1. Temperature standard deviation (in °C) at 200 m depth in the Pacific using data from Levitus (a), and from a numerical simulation performed with the NCAR ocean model forced with observed surface forcing over the period 1958-1997 (b). Contour interval is 0.1°C.

2. The Ocean Signal

Zonal sections of temperature and salinity anomalies along 13°N show that the largest amplitudes are found in the thermocline, and are associated with large changes in isopycnal depth, suggesting that the variability along 13°N results from adiabatic vertical displacements of the thermocline. Power spectra of the 25.5 depth anomalies are dominated by the low-frequencies (periods longer than ~7 years). The propagation characteristics have been studied using cross-spectral analysis. In the decadal range (7-10 years) the thermocline displacements along 13°N remain significantly coherent (at the 90% significant level) with the thermocline displacements in the eastern end of the basin all the way to the western boundary. Significant coherence is also found along part of the western boundary, both to the north and the south of 13°N, suggesting that after reaching the boundary the propagating signal continues along the boundary. In the same spectral band (7-10 years), phase lags increase monotonically from east to west. Phase speeds can be estimated by relating the phase lags between the signal at different points to the distance between those points. It is found that in the eastern half of the basin the phase speed is ~13 cm/s, while west of the dateline the phase speed increases to ~22 cm/s. To rationalize the results in the context of Rossby wave dynamics, the phase speeds of the first three baroclinic Rossby wave modes have been estimated using the WKB approximation, for the given model stratification along 13°N. For the first baroclinic mode, the phase speed predicted by linear theory for the model increases from about 15 cm/s close to the eastern boundary to ~20 cm/s in the western half of the basin, where the thermocline deepens. Thus, the ocean signal is slower than the first baroclinic Rossby wave mode east of the dateline, while west of the dateline the ocean signal propagates slightly faster. The different propagation characteristics in the eastern and western halves of the basin will be rationalized by examining the nature of the forcing.

3. Forcing

Ekman pumping is the component of the surface forcing which can be expected to produce adiabatic thermocline displacements. For this model simulation, Ekman pumping is derived from the NCEP/NCAR reanalyses. Correlations as large as 0.8 are found between Ekman pumping and thermocline displacements east of ~180°, while west of the dateline correlations drop to values smaller than ~0.5. Also, cross-spectral analysis shows that east of the dateline Ekman pumping anomalies in the 7-10 year spectral band propagate westward with a phase speed of ~9 cm/s. Thus, we can interpret the ocean model signal east of the dateline as forced Rossby waves. The westward propagation of “decadal” Ekman pumping anomalies with a phase speed which is very close to the phase speed of free, baroclinic Rossby waves can explain the “decadal” timescale of the waves as a “quasi-resonant” response to the forcing. West of the dateline, on the other hand, thermocline variability mainly consists of free, first-mode baroclinic Rossby waves.

4. Area of Influence

After reaching the western boundary around 13°N, the ocean signal can be traced along the western boundary all the way to the equator, and along the equator. The amplitude of the isopycnal displacements along the equator is much reduced with respect to the amplitude along 13°N, with average depth variations of 5-10 m. However, these changes in thermocline depth appear to persist for several years, and have the potential of modulating ENSO at decadal timescales.

5. Discussion

Given the surface Ekman pumping, we have been able to explain the nature of the ocean signal, and its long (decadal) timescale. However, we do not know what determines the westward propagation of the Ekman pumping anomalies at decadal timescales, whether due to local coupled Rossby waves, or linked to the basin-scale decadal variability. We have also not explained why the ocean response is particularly large in the 10°-15°N latitude band. Maybe the answer is to be found in the thermocline structure at these latitudes. These issues will be analyzed in future studies.

Victoria Coles and Michele Rienecker

Two isopycnal models of the Pacific Ocean forced with climatological surface momentum and buoyancy forcing show exchange between the subtropical and tropical gyres in the thermocline. Both western boundary and interior basin pathways exist, and have strong seasonal cycles in transport magnitude. Seasonal variability in the interior basin pathway is linked to changes in the potential vorticity or thickness of the thermocline ridge located under the ITCZ at roughly

10N. TAO moorings near the ITCZ show a seasonal cycle in the thickness of the permanent thermocline consistent with the models. In the models, the thermocline shoals in fall as the wind stress curl in the central Pacific intensifies Ekman pumping and NECC transport. Simultaneously, freshwater associated with the ITCZ stabilizes and shoals the mixed layer. Because the base of the thermocline is less upwelled than shallower isopycnals, the net impact on the thermocline is a thickening. As the wind stress curl relaxes in spring and early summer, the upper layers of the thermocline deepen, and the mixed layer also deepens, compressing the thermocline.

The exchange and seasonal variability can be traced through the injection and advection of an ideal tracer with source function north of the ITCZ. The tracer fields show convergence of subtropical water onto the equator in the western basin and at 140W, consistent with observations of tritium concentration. The interior pathway is controlled by surface wind stress curl at the ITCZ, and the tracer fields show seasonal variations in the equatorward transport over the entire pathway to the equator. The injection of subtropical water occurs over a limited spatial area, and during a short time span, and may have a more significant impact on the equatorial thermocline than the annually averaged net meridional heat flux might suggest.

Simulations with SSM/I derived wind forcing show interannual variability in the interior pathway consistent with variations in the Ekman pumping at 10N, and with the variations in thermocline potential vorticity. Interannual variability is dominated by the ENSO cycle.

STC in the Indian Ocean From an Assimilation of WOCE Hydrographic Data in a Primitive Equation Model

Bruno Ferron and Jochem Marotzke

An assimilation experiment using a primitive equation model and its adjoint is examined to study the Indian Ocean circulation of year 1995 where most of WOCE CTDs were collected. Mean sea surface height and hydrography are assimilated in this regional coarse resolution (1°x1° horizontally) model. The circulation is forced with NCEP fields. The annual mean STC and its seasonal variations are investigated.

North Brazil Current Experiment: First Results

Silvia L. Garzoli, Gustavo J. Goni and Bill Johns

The North Brazil Current is a major western boundary current in the tropical Atlantic that transports upper ocean waters northward across the equator. It plays a dual role, first in closing the wind-driven equatorial gyre bounded on the south by the South Equatorial Current (SEC) and on the north by the North Equatorial Countercurrent (NECC), and second in providing a conduit for cross-equatorial transport of South Atlantic upper ocean waters as part of the Atlantic meridional overturning cell (MOC). The NBC separates sharply from the coast at 6-7N and curves back on itself (retroflects) to feed NECC. During this retroflection phase the pinch off large anticyclonic current rings. These features then move northwestward toward the Caribbean, roughly paralleling the South American coastline. These features then move northwestward toward the Caribbean, roughly paralleling the South American coastline. NBC ring shedding is thought to account for as much as one-third of the net warm water transport across the equatorial-tropical gyre boundary into the North Atlantic in compensation for the southward export of North Atlantic Deep Water.

In order to study the precise mechanisms which contribute to NBC ring formation, the structure and dynamics of the rings themselves, and the role that they play in the inter-ocean exchange of heat and salt, and extensive field program started in November 1998 and ended in June 2000. Four hydrographic/velocity cruises took place during that period. An extensive array of moored instruments (14 inverted echo sounders, 1 current meter mooring and 1 CMM/CTD mooring) was deployed and recovered. Surface drifters were satellite tracked. RAFOS floats at different depths were launched and tracked acoustically. Altimeter data was analyzed to complement these observations. The poster will present examples of the data sets collected and the first results from the experiment. Initial results indicate that the rings may account for more than one third of the MOC.

The Role of Seasonal Stress Variations in Generating Long-Term Mean Heat Flux into the Indian Ocean

J. S. Godfrey, Y.-L. Zhang, A. Schiller, L. J. Waterman and R. Fiedler

In boreal summer, more water is upwelled and warmed north of the equator in the Indian Ocean than can be removed by the (well-defined) annual mean Ekman transport across the equator; this latter is the annual mean of $[d(\tau_x)/dy]/(\rho^*\beta)$, integrated across the Indian Ocean at the equator. A numerical model experiment has been designed to explore the fate of the summer excess upwelling. In a control run, a global ocean circulation model is forced with observed seasonal shortwave radiation, wind speeds, air-sea humidity ratios and wind stresses. In the experimental run, the wind stresses (only) are replaced by their 12-month running mean (12MRM). 12MRM inter-run differences in downward surface heat flux and SST have a ratio of about -10 W/m^2/°C in the tropics, much as expected from the fact that radiation, surface wind speeds and humidity ratios are the same in the two runs. SST rises as much as 2.5°C off Somalia compared to the control, due to reduced coastal upwelling. 0-100 m velocity differences might be expected to be in thermal wind balance with the 0-100 m average temperature; in fact they flow down a strong gradient of this average temperature, to cross the equator near Africa in a shallowing of the cross-equatorial "Southern Gyre". They then join a strong eastward (difference) equatorial jet, thus removing the warmer Somali SST eastwards.

The 12MRM net heat flux into the Indian Ocean north of 70S reduces (in the model) by 6 W/m^2. This area-mean flux reduction demands changes in the mean overturning circulation to balance it. The mixed layer thins, averaged over area north of 70°S and on 12MRM, and near-surface steric heights therefore decrease north of 70°S. This causes the near-surface inflow from Indonesia to increase, and the deeper inflow to decrease - i.e. the Throughflow shallows. The enhancement in the near-surface Indonesian Throughflow flow joins a complex (difference) circulation, in which changes to the Southern Gyre and to mean zonal flow along the equator are prominent. These changes mainly cause the western boundary inflow to be shallower and warmer, and the (southward) Sverdrup outflow across 70°S to be deeper and colder in the experiment compared to the control run. The net effect is a reduction of the overturning cell, resulting in the 6 W/m^2 reduction in surface heat flux. At least some of the change in vertical mass transport (and water mass change) seems to occur in the eastern equatorial Indian Ocean, perhaps due to removal from the experimental run of low Richardson numbers associated with the transient Wyrtki Jets. If so, it suggests that the causes of change in area-average SST and heat flux in the northern Indian Ocean may lie as much in mixing in the eastern Indian Ocean as in the Somali Current system.

Shallow Tropical Cells in the Atlantic: Observations of Pathways, Transports and Variability

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As part of the German CLIVAR program new measurements with moored current meter stations, profiling floats (APEX) and shipboard observations are being carried out in the equatorward flow of the South Atlantic STC. The evaluation builds on our previous WOCE measurements and on studying historical hydrographic data.

Special focii of the study are:

(1) Variability of the transport of NBUC/NBC by moored array at 11S.

(2) Transports and water mass variability on standard sections (10S, 5S, 35W).

(3) STC pathways by shallow (200 m or 400 m) trajectories of profiling floats.

(4) Comparison with GCMs.

Overturning Cells in the Upper Pacific

Wilco Hazeleger, Pedro de Vries and Geert Jan van Oldenborgh, KNMI, Oceanographic Research Dept.

Source waters of the equatorial thermocline in the Pacific are studied with a high-resolution ocean model (OCCAM). Using annual mean transports a tropical overturning cell (TC) is found that consists of downwelling 5 degrees poleward of the equator and upwelling at the equator. Also a subtropical cell (STC) is found that has its downwelling branch around 20 degrees and upwelling at the equator. When the annual mean overturning in density coordinates is considered the TC is much weaker while the STC is hardly effected. When also high-frequency isopycnal mass fluxes are included the TC is completely compensated by an eddy-induced overturning. Seasonal variations in the STC

And tropical instability waves are responsible for the compensation. So the Lagrangian mean circulation shows that the equatorial thermocline is ventilated by the STCs only. It appears that the largest contribution comes from the South Pacific.

Decadal Upper Ocean Temperature Variability in the Tropical Pacific

Wilco Hazeleger, Martin Visbeck, Mark Cane, Alicia Karspeck and Naomi Naik

Decadal variability in upper ocean temperature in the Pacific is studied using observations and results from model experiments. Especially propagation of upper ocean thermal anomalies from the midlatitudes to the tropics is studied as a possible source for decadal equatorial thermocline variability. In the observations propagation along the subtropical gyre of the North Pacific is clear. However, no propagation into the equatorial region is found. Model experiments with an ocean model forced with observed monthly wind and wind stress anomalies are performed to study the apparent propagation. Distinct propagation of thermal anomalies in the subtropics is found in the model, although the amplitude of the anomalies is small. The anomalies clearly propagate into the tropics but they do not reach the equatorial region. The small response at the equator to extratropical variability consists of a change in the mean depth of the thermocline. It appears that most variability in the subtropics and tropics is generated by local wind stress anomalies. The results are discussed using results from a linear shallow water model in which similar features are found.

Decadal Variability of the Interior Communication Window

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The communication from the subtropical gyre interior to the tropics is examined, using wind stress datasets and results from an ocean data assimilation system. It is shown that the interior communication can be clarified by a simple Interior Mass Communication Index (IMCR), which can be easily calculated from the Sverdrup function. For the northern (southern) hemisphere the IMCR can be defined as the meridional minimum (maximum) of the Sverdrup function maximum (minimum) at each latitude. The interior communication is closely related to the ENSO cycle, and its rate and pathway have strong interannual-decadal variability.

Wind Stress Effects on the Atlantic Subtropical - Tropical Circulation

Tomoko Inui, Alban Lazar, Paola Malanotte-Rizzoli and Antonio Busalacchi

A reduced-gravity, primitive equation, upper-ocean GCM is used to study circulation pathways in the Atlantic subtropical and tropical gyres. The model has a variable-depth oceanic mixed layer coupled to an advective atmospheric mixed-layer. The model calculates the heat and salinity flux internally by using wind speed and cloud cover. The Hellerman and Rosenstein (HR) and DaSilva (DSV) climatological annual-mean and monthly wind stress forcings are used to force the model. Bernoulli function, trajectory, and transport analyses are performed to characterize the circulation, and to isolate subsurface pathways between the subtropical and tropical gyres.

A comparison between the annual--mean forced experiment of HR and DSV shows two results for the North Atlantic:(1) the communication window between the subtropical and tropical gyres is similar in width, (2) the interior exchange window width is substantially larger in the DSV than

HR experiment, accompanied by larger transport as well in the DSV experiment. The South Atlantic exhibits a similar communication between the subtropics and tropics for either wind data set. The annual-mean of the seasonally varying forcing also supports these results. A two layer ventilated thermocline model shows that the communication window for subsurface pathways is approximately a function of the east-west gradient of the Ekman pumping at outcrop lines divided by the one at the subtropical-tropical gyre boundary. This solution is validated using three additional GCM experiments. It is proved that the communication windows are primarily controlled by the wind stress effects. Within the communication window, the interior exchange window is expected to provide a characteristic-conserved water to the equatorial region compared to the western boundary exchange window. The interior exchange window is widened by two factors: (1) eliminating part of the positive Ekman pumping region in the eastern North Atlantic, (2) weakening the Ekman pumping over the whole region.

Pathways in the Tropical Pacific Related to the 1997-99 El Niño-La Nina

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El Niño-La Niña events have increased in frequency and intensity during the last decades. The El Niño event of 1997 was the strongest on record and the transition to La Niña was very rapid. Studying the sources and the pathways of the equatorial water masses during these events should improve the understanding of the short-term mechanisms and the long-term variability of El Niño-La Niña. In particular, it must be tested how much of the long-term variability is influenced by the shallow tropical-subtropical overturning cells. The ocean global circulation model used for this study is the ORCA 8.2 version of the OPA model developed at LODYC in Paris. It is forced by the ERS wind stress and NCEP heat and fresh water fluxes with a relaxation to the Reynolds' SST. A software, calculating the 3D trajectories of virtual floats backward or forward, is applied over the period 1994-99 in the tropical Pacific. The Lagrangian diagnostics done with that software provide useful information on the origin and evolution of the equatorial water masses. In order to validate the model, the simulated currents are compared to the current meter measurements of the TAO moorings. The validation results in a good correlation at the surface, as well as at the level of the equatorial undercurrent. Interesting pathways in relation to the 1997 El Niño are first highlighted, even if there is no typical scheme in view of the strong variability of the currents. Backward trajectories of floats, launched at the beginning of El Niño at the dateline around the equator and near the surface, indicates that the water masses have stayed in the equatorial band over the last 3-4 years. This accumulation of water in the equatorial band is in agreement with the buildup theory of Wyrtki. Forward trajectories from the same locations and date evidence a discharge of these water masses during the mature phase of El Niño, at first southward and then northward. The rapid transition to La Niña is due to the sudden arrival in May 1998 of upwelled water at the surface around 0°-130°W. Interestingly, the backward trajectories issued from the patch of cold water at the surface indicate that nearly all of the cold water came from the extra equatorial regions. With only four years of backward integration, the water masses can only be tracked to about 15° of latitude and 200 meters depth. The water masses were brought in both hemispheres by the subtropical gyres into the western boundary currents in about two years. After several months in these currents, they moved into the equatorial undercurrent. This current finally brought them to the surface near 0°-130°W after one year and a half. The influence of the long-term oceanic circulation from the subtropics to the equatorial upwelling region is thus important for the development of La Niña in 1998. Forward trajectories, launched from the same locations and date, show the Ekman divergence and associated poleward transport of water during the mature phase of La Niña. This preliminary study indicates that the pathways between the equatorial and subtropical regions vary strongly in relation with the 1997-99 El Niño-La Niña. However, a clear pathway connecting the northern and southern subtropical gyres to the central-eastern equatorial Pacific appears important in the development of La Niña. Different forcing and models with and without data assimilation will be used to determine the realism of such pathways, and quantify their importance in the long-term variability of El Niño-La Niña.

Does the Meridional Overturning Circulation Shut Off the Northern STC in the Atlantic?

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Results from twin numerical simulations of the Atlantic using the U.S. Navy's Layered Ocean Model are used to diagnose the impact of the Alantic Meridional Overturning Circulation (MOC) on the subtropical cells. One simulation is forced by seasonal winds alone, and another is forced by the same winds plus a 14 Sv MOC imposed by inflow/outflow ports at the model's northern and southern boundaries. The simulation forced by climatological winds alone shows an essentially symmetric shallow overturning cell about the equator with nearly equal amounts (~8 Sv) of thermocline water from both hemispheres feeding the EUC and associated equatorial upwelling. The addition of a realistic MOC alters this to a situation in which most of the supply of thermocline water to the equator is from the South Atlantic (14 Sv) while only about 2 Sv comes from the North Atlantic. In both simulations, despite the different magnitudes of the exchanges, the equatorward thermocline flow is concentrated in western boundary currents within 10 deg. of the equator, indicating a dominant western boundary ventilation pathway in both hemispheres. In the wind forced experiment, the northern hemisphere STC exhibits a significant seasonal cycle about its annual mean, with maximum equatorward transports during October-March and a minimum (nearly zero) transport in May-June. This seasonality is obscured in the MOC-driven model by nonlinear effects related to superposition of the MOC and wind-driven flows, most notably the shedding of rings from the North Brazil Current.

Subtropical Cell Pathways in the Atlantic Inferred From Climatological Hydrographic Data

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An analysis of the available hydrographic data base in the Atlantic between 1950-2000 is carried out to determine the interior pathways of thermocline water from the shallow subtropical subduction regions to the equator. The geostrophic velocity streamfunction relative to 1200 m is computed on density surfaces and integrated vertically and zonally to estimate the total transport flowing equatorward below the surface Ekman layer. In the North Atlantic the interior ventilating trajectories are confined to densities between about 24.0 to 26.0 sigma-theta and only about 2 Sv of water reaches the equator via the interior. Flow on the shallower of these density surfaces (24.4 sigma) originates from the central Atlantic near 40°W between 15-20°N while flow on the deeper surfaces (25.4 sigma) originates from near 20°W just off the coast of Africa. The ventilating trajectories reach their westernmost location at about 10°N, where there appears to be a concentration of equatorward flow near 40°W and again near the western boundary. In the South Atlantic, about 4 Sv of thermocline water reaches the equator through the interior in the same range of densities, but weighted toward a slightly higher mean density than the North Atlantic. The shallower layers originate from the central part of the basin (along 10-30°W at 10-15°S) while the deeper layers originate at higher latitudes (~20°S) from the eastern part of the basin, similar to the North Atlantic. The ventilation pathways from the South

Atlantic are spread over a wide interior window, which at 6°S extends from 10°W to the western boundary. The contributions of western boundary currents to the equatorial thermocline ventilation are not accounted for in this analysis.

Observations of Interior Equatorward Flow and Equatorial Divergence in the Pacific Ocean

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Interior ocean circulation pathways from the subtropics to the equator are quite different in the northern and southern hemispheres of the Pacific Ocean. These contrasts are delineated, and an equatorial upwelling estimate is made, using salinity, planetary potential vorticity, and acceleration potential distributions, as well as direct velocity data. In the North Pacific the pycnocline shoals and strengthens under the Intertropical Convergence Zone, separating the North Equatorial Current from the North Equatorial Countercurrent. The resulting high potential vorticity suggests inhibition of meridional water-property exchange between the subtopics and the equator, and the mean interior southward transport of pycnocline water is estimated at only $5 \pm 1 \times 10^6$ m^3 s^{-1}. In contrast, the southern branch of the South Equatorial Current carries $15 \pm 1 \times 10^6$ m^3 s^{-1} of pycnocline water from the southern subtropics northwestward directly to the equator in the South Pacific through an interior region of low and relatively uniform potential vorticity. In both hemispheres, these interior pathways extend downward as far as the lightest waters of the equatorial pycnostad. The investigation is continued equatorward using direct estimates of upper ocean horizontal velocity and divergence from shipboard observations between 170°W and 95°W. Mean meridional currents for this longitude range include equatorward flow within the pyncocline reaching 0.05 m s^{-1} in the south and 0.04 m s^{-1} in the

north near 23°C (85 m) as well as poleward surface flows reaching 0.09 m s^{-1} in the south and 0.13 m s^{-1} in the north. Again, -an asymmetry in the meridional transports is found, suggesting that on the order of 10 x 10^6 m^3 s^{-1} of pycnocline water from the southern hemisphere is upwelled at the equator and moves into the northern hemisphere as surface water. This interhemispheric exchange path could be part of the route for water from the southern hemisphere to supply the Indonesian Throughflow. Vertical velocity is diagnosed by integrating regional horizontal divergence

down from the surface. Equatorial upwelling velocities peak at 1.9 (±0.9) x 10-5 m s^{-1} at 50 m. The upwelling volume transport in the area bounded by 3.6°S, 5.2°N, 170°W, and 95°W is 62 (±18) o 10^6 m^3 s^{-1} at 50 m. Strong downwelling is apparent within the North Equatorial Countercurrent.

Generation of North Brazil Rings - A New Hypothesis

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A high resolution OGCM is used to study the dynamics of crosshemispheric flow in the Atlantic. It is found, that eddies are created at all depths and latitudes along the western boundary to change the sign of the flow's potential vorticity by friction. This gives rise not only to the often observed surface intensified North Brazil Rings (NBR) but also to thermocline intensified EUC-Rings (shedded in the NBC/EUC retroflection as suggested by Fratantoni) and intermediate eddies which have yet to be observed. The latter two usually won't have a surface signal. Those three types of eddies can, but need not merge and thus give rise to a large variety in the vertical structure of NBRs - consistent with observations. In addition to a pure description of those Rings we will briefly explore their dynamics and their mutual vertical interaction.

Interannual Variability of the Subsurface High-Salinity Tongue South of the Equator at 165°E

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A time serie of salinity at thermocline level was constructed from repeated meridional CTD sections along 165°E during 1984-2000. A tongue of high-salinity water extends along the isopycnal sigma-t=24.5 from its surface outcrop in the southeast Pacific to 175 m depth near 5-10°S along the section at 165°E, and is the source of high-salinity water that forms the south flank of the equatorial undercurrent. In the west, the entire tongue moves vertically with the thermocline, mostly as part of the ENSO cycle, but salinity in the tongue also varies interannually over a range of 0.4 PSU, and this was only partly associated with ENSO.

Most of the salinity variability on the isopycnal can be accounted for as a result of zonal advection, partly due to changes in the zonal current and partly to changes of the zonal gradient. These suggest that better observations of E-P and winter mixed layer formation in the southeast Pacific outcrop region would be important in understanding slow variations in the ventilation of the equatorial thermocline.

The Relationship Between Oscillating Subtropical Wind and Equatorial Temperature

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An earlier study showed that an atmosphere-ocean model of the Pacific develops a midlatitude oscillation that produces decadal sea-surface-temperature (SST) variability on the equator. We use the ocean component of this model to understand better how subtropical wind stress oscillations can cause such SST variability. The model ocean consists of a motionless abyss and three active layers which roughly correspond to the mixed layer, thermocline, and intermediate water. Both temperature and density can vary horizontally within each layer.

For a steady wind, the model develops a Subtropical Cell (STC), in which northward surface Ekman transport subducts, flows equatorward in the thermocline, and returns to the surface at the equator. A prescribed subtropical wind-stress anomaly perturbs the strength of the STC, which in turn modifies equatorial upwelling and equatorial SST. A midlatitude wind-stress anomaly has a much smaller, indirect effect on equatorial SST.

We use the transient response to a switched-on wind perturbation to predict the ocean response to an oscillating wind. This method correctly predicts the results of several numerical experiments and extends these results to a wide range of forcing periods. The transient and oscillating cases have a more complicated relationship (compared to the steady case) between perturbations to equatorial SST and the various branches of the STC (i.e., equatorial upwelling, poleward surface flow, and equatorward thermocline flow). The thermocline branch anomalies

are generally much weaker than those in the surface branch and equatorial upwelling. The thermocline branch anomalies always lag the surface branch and lag the equatorial upwelling for timescales of several decades. It is the upwelling strength that most directly affects equatorial SST. Thus, geostrophic flow in the thermocline may not be as good an indicator of tropical-subtropical interaction as suggested in several recent studies.

On the Ventilation of the Tropical Atlantic Thermocline

Alban Lazar, T. Inui, P. Malanotte-Rizzoli, A. J. Busalacchi and L. Wang

This study is an attempt to assess some of the physics of the thermocline branches of the subtropical-tropical cells (STC) in the Atlantic Ocean. We carried out a diagnostic and sensitivity study of the seasonal cycle of subduction, entrainment and subsurface circulation in an upper Atlantic ocean GCM. Whereas previous large scale studies of tropical water formation where limited to portions of the regions at play, our results provide a global picture of the surface sources and sinks of the subsurface branches of the STCs. We present subduction rate, epoch and duration. These quantities, poorly known for the Southern basin, are essential since they quantify the main source of water of the upper equatorial thermocline. Then the subsurface pathways are characterized using conservative quantities as well as releases of Lagrangian floats. Some elementary comparisons with some available observations are also presented. In term of mechanisms at play, we emphasize the importance of the wind in the timing and intensity of subduction

Dynamics of Interdecadal Thermocline Variability in the Tropical-Extratropical Ocean

Zheng Liu

We will examine two closely related topics on interdecadal thermocline variability: the evolution of the thermocline anomaly in the extratropical ocean (Stephens *et al.*, 2000) and the penetration of the influence of extratropical forcing on the equatorial thermocline (Shin and Liu, 2000).

First, the evolution of decadal subduction temperature anomalies in the subtropical North Pacific is studied using a simple and a complex ocean model. It is found that the amplitude of the temperature anomaly decays faster than a passive tracer by about 30- 50%. The faster decay is caused by the divergence of group velocity of the subduction planetary wave, which is contributed to, significantly, by the divergent Sverdrup flow in the subtropical gyre. The temperature anomaly also seems to propagate southward slower than the passive tracer, or mean ventilation flow. This occurs because the mean PV gradient in the ventilated zone directs eastward; the associated general beta effect produces a northward propagation for the temperature anomaly, partially canceling the southward advection by the ventilation flow.

Second, OGCMs are used to investigate the response of the equatorial thermocline to extratropical buoyancy forcing. Passive tracers and analytical theories are also used to shed light on the dynamics of the thermocline response. The major findings are the following. a) The midlatitude region seems to be the optimal region for surface buoyancy forcing to affect the equatorial thermocline. This occurs because, first, thermocline anomalies in the midlatitudes can penetrate into the equator very efficiently; second, buoyancy forcing generates a strong local response in the midlatitudes. b) Dynamic waves as well as thermocline ventilation contribute to the response in the equatorial thermocline. Consequently, equatorward penetration is substantially greater for a temperature anomaly than for a passive tracer. c) Midlatitude forcing generate significant temperature response in the equatorial thermocline for forcing periods longer than decadal. d) For a low latitude (10°-20°) buoyancy forcing, the equatorial thermocline could be dominated by a temperature anomaly that has the opposite sign to the surface forcing, because of the strong higher mode baroclinic response in the ventilated thermocline. Finally, the relevance of our work to observations and climate variability is also discussed.

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Variability of the Upper Tropical Atlantic in the Coupled Model ECHAM4/OPYC

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We investigate the variability of the upper tropical Atlantic Ocean using a 300 year integration of the coupled atmosphere/ocean general circulation model ECHAM4/OPYC.

The model's shallow tropical meridional cell extends over the upper 100 meters. It is strongest in the western part of the tropical Atlantic where upwelling velocities of up to 2*10-5 m/s occur.

Looking at the horizontal current system in the subsurface layers the model reproduces the EUC (Equatorial Undercurrent) as well as the NBUC (North Brasil Undercurrent), with the current cores located at a depth of about 100 to 150 meters.

The transport of the NBUC is about 20 Sv (1 Sv = 10^6 m^3/s) at 5°S. It shows a pronounced semi-annual period in contrast to the transport at 10°S where an annual cycle is predominant. The annual mean values at 5°S exhibit variations of +/- 2.5 Sv or about 10% of the mean transport.

Questions of interest (in cooperation with the observational group at IfM Kiel, Germany) are relations between the variations of the NBUC transport, the equatorial divergence (by means of the upwelling velocity) and the divergence of the meridional Ekman transport as well as the temperature/salinity anomalies in the NBUC and in the equatorial upper layer

Observations of the Time Varying Shallow Tropical/Subtropical Overturning Circulation in the Pacific Ocean

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This study investigates the shallow subtropical/tropical overturning circulation (STC) in the Pacific using the hydrographic data formed by the combination of NODC98 and PMEL data sets covering the last 50 years. The inclusion of PMEL data, in particular the CTD surveys during the service of TAO moorings, significantly improved the data coverage of the 1990s in the tropical region. The subtropical/tropical communication pathway is delineated by the spatial patterns of the salinity, potential vorticity (PV), and geostrophic flows on isopycnal surfaces. In the northern hemisphere, the mean interior pathway takes a zigzag detour around the high PV associated with the ITCZ, with a mean equatorward transport of 7 Sv. The interior communication between the tropical and subtropical gyre is more direct in the South Pacific, with a mean equatorward transport of 14 Sv. The data allow for an examination of decadal time scale variations about the mean. The most significant change occurred in the northern hemisphere STC in the 1990s when the interior flow was reduced by one half in the thermocline. These results are consistent with Sverdrup solutions forced by various wind products from COADS, FSU, ECMWF, NCEP, and satellite measurements. The period of the 1990s was one of unusual warmth in the tropical Pacific, suggesting that decadal changes in the shallow thermocline circulation may play a role in regulating the background state of the ocean in this region.

Structure and Dynamics of the Indian-Ocean Cross-Equatorial Cell

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In contrast to the Atlantic and Pacific Oceans, significant upwelling does not occur along the equator in the Indian Ocean, but rather in the northern hemisphere off Somalia, Oman, and the southern tip India. The annual-mean Cross-equatorial Cell (CEC) in the Indian Ocean is a shallow (z > -500 m) meridional overturning circulation, consisting of this northern-hemisphere upwelling, southward flow of surface water, subduction at southern-hemisphere midlatitudes, and a return northward flow of thermocline water. In this study, several types of ocean models are utilized to investigate CEC structure and dynamics.

(1) We first discuss the driving force of the cross-equatorial flows associated with the CEC. Levitus (1987) and others have suggested that the surface cross-equatorial transport results

from the Ekman drift being directed southward on both sides of the equator during the summer monsoon, a consequence of the zonal winds being westerly (easterly) north (south) of the equator. During the winter monsoon, the Ekman drift is directed oppositely. In itself, this explanation is unsatisfying because Ekman theory breaks down at the equator. Here, we show that it is the wind curl that drives the required surface cross-equatorial flow, consistent with Sverdrup theory. Interestingly, the Sverdrup transport is equal to the Ekman drift in a special case, being well defined even on the equator.

(2) We also show that the cross-equatorial flow occurs just below the surface typically beneath surface opposite flow, so that there is a shallow, cross-equatorial "roll." The roll is clearly seen in the JAMSTEC OGCM data. This roll is not an artifact of zonal averaging, but rather occurs across much of the interior ocean. We can reproduce the cross-equatorial flow using a linear continuously stratified modal model very well, which indicates that its basic dynamics are linear. We demonstrate that the roll is a direct (local) response to the southerly cross-equatorial component of the winds during the Southwest Monsoon.

(3) Finally, we illustrate the 3-dimensional pathways of CEC by tracking model drifters from the northern-hemisphere upwelling regions, forwards in time to follow the surface pathways and backwards in time to follow the subsurface flows. The subsurface branch of the CEC crosses the equator along the western boundary. We show that some of water parcels that upwell off Somalia and Oman come from the Indonesian Throughflow; others come from the subduction region in the southern Indian Ocean. After upwelling, surface drifters in the OGCM solution move to the equator, but do not directly cross it because of the equatorial roll. Instead, they drift eastward across the basin just north of the equator, and only then cross the equator to move into the South Indian Ocean.

Decadal Variations of the Strength of the Pacific Subtropical Cells and Their Effect on the Tropical Heat Balance

Masami Nonaka (IPRC/FRSGC), Shang-Ping Xie and Julian P. McCreary, Jr. (IPRC)

While it has been documented in many studies that equatorial sea surface temperature (SST) exhibits decadal variations, the mechanisms that cause these variations are not yet known. Because of its long adjustment time scale, the midlatitude ocean has been suggested to be important, in contrast to the interannual variations of ENSO for which the tropical ocean is primarily involved. In a simple ocean-atmosphere coupled model, Kleeman *et al.* (1999; KMK) suggested that variations in the strength of the Pacific Subtropical Cells (STCs) could cause equatorial SST anomalies. In this scenario, strengthened (weakened) STCs transport more (less) subtropical water into the tropical Pacific by the STCs, and consequently the water that upwells in the eastern, equatorial ocean is cooler (warmer), thereby cooling and expanding (warming and shrinking) the equatorial cold tongue. The KMK mechanism, however, has been shown to operate only in their simple coupled model. It is not yet known if the STC in the real Pacific Ocean exhibits decadal variations, or if it influences equatorial SST.

In this study, we examine STC and equatorial SST variations in a Pacific OGCM forced by realistic winds, the NCEP reanalysis from January 1958 to December 1997. Decadal variations of upper-layer temperature in the eastern equatorial ocean are present in this “standard” solution. We also performed an additional test experiment in which equatorial wind-stress variations were suppressed near the equator, weakening from full strength to zero from 10 degrees to 3 degrees of the equator.

In the test solution, eastern equatorial SST has decadal variations more than half their amplitude in the standard case, whereas the amplitudes of interannual variations are reduced to about one forth. This difference suggests that the mechanisms that influence equatorial SST are different for decadal and interannual variations; in particular, more than half of the decadal variations are caused by off-equatorial wind-stress variations. In the standard run, temporal variations of STC strength also vary decadally, so that the heat transport from the subtropics to the tropics by the STCs exhibits decadal variations that coincide well with eastern-equatorial SST variations. A heat-budget analysis of the tropical ocean shows that on decadal time scales, variations of net heat transport via the STC almost balance the surface heat flux. These results suggest that realistic variations of the Pacific wind-stress field can lead to decadal variations of STC strength, and that their effect on the tropical heat budget is consistent with the KMK mechanism.

Effects of Mixed Layer Variability on the Subduction of Subtropical Underwater in the North Atlantic

Bridgette M. O'Connor, Rana A. Fine, Donald B. Olson and Robert L. Molinari

Subtropical Underwater (STUW) is a component of Subtropical-tropical Overturning Cells (STC), and understanding variability in its formation is a key toward understanding STC variability. Mixed layer effects on the subduction of STUW have been examined in the North Atlantic. This involved calculating a STUW subduction rate of 36 m/yr from WOCE/TOGA drifter array over the period 1989-1998, and 44 m/yr from CFCs. The rate of formation for both methods is about 2 Sv. Drifter velocities are used to examine the contribution to the subduction rate from lateral induction and vertical pumping corrected for vorticity stretching. The lateral induction is calculated from the ambient mean velocity components and the mixed layer depth gradient. The mixed layer was first calculated from the Levitus climatology based on a sigma-t criterion of 0.125. To examine the effect of mixed layer variability in the calculation, the mixed layer was calculated over 1989-1997 using XBT data and a temperature criterion of 0.005. The XBT data were further subdivided into high NAO and low NAO years: 1989-1994 and 1995-1997, respectively. The XBT data gave consistently higher subduction rates than Levitus climatology. The largest effect of mixed layer variability was on the lateral induction term, which is negligible using climatology and increases to 10 m/yr using the XBT data. The lower rate from climatology is probably due to smoothing, as well as the fact that the climatology spans a long time period. The effect on the other terms is less significant. The rate increased by about 60% over 1989-1997; the increase during the high NAO years was about 50%, and during the low NAO years, as much as 80%. The large increase during the low NAO years maybe expected due to southward shift in winds. Mixed layer depth changes have a one to one effect on the pumping terms. However, mixed layer effects are not straightforward, because it is the mixed layer configuration that affects the lateral induction term and not mixed layer depth. Small changes in mixed layer configuration can result in major changes in the lateral induction term, whereas mixed layer depth changes have no effect on lateral induction once the configuration is the same. A part of the task of modelling variability of STCs is to represent STUW formation. Thus, processes affecting the intensity and configuration of STCs need to be understood.

Interannual Variability and Intergyre Exchange Pathways for the Pacific Ocean

Keith Rodgers, Bruno Blanke, Christophe Menkes and Gurvan Madec

The ORCA incarnation of the OPA model (Madec *et al.*, 1998) is used to study intergyre exchange for the subtropical cell of the Pacific Ocean. The meridional resolution of this model in the tropics is 0.5 degrees, and its zonal resolution is 2 degrees, and thus it captures the critical scales of the tropical ocean currents. The model has been initialized with Levitus temperature and salinity data, and forced with NCEP reanalysis fields over the period 1948-1999. Eulerian and Lagrangian diagnostics have been used to analyze the mean circulation, as well as its variability. One point of focus of this study has been to better describe the changes in ocean circulation associated with the 1976 “climate shift” in the Pacific. The main finding is that after 1975 the transport of the equatorial undercurrent diminishes by approximately 25 per cent, and that the core of the undercurrent shifts to a slightly shallower isopycnal horizon. The sensitivity of the subtropical cell to the parameterization of horizontal mixing chosen for the model is also investigated.

The Role of Off-Equatorial Subsurface Anomalies and NECC Subsurface Pathways in Influencing the 1991-1992 El Nino

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A combination of observations and numerical model results are combined to formulate a plausible scenario about how off-equatorial subsurface thermal anomalies, propagating from subtropical gyre subduction sites, can influence tropical Pacific sea surface temperature (SST) via subsurface pathways of the North Equatorial Counter Current (NECC).

Based upon NCEP ocean re-analysis data, the three dimensional co-evolution of the tropical Pacific climate system is first examined to explain the onset of the 1991-92 El Nino event. A logical sequence is found that links subsurface and surface temperature anomalies on and off the equator in the western and central Pacific. Along the NECC path, subsurface temperature anomalies propagate coherently eastward from the western boundary in mid-1989 to the date line in mid-1990. As the thermocline shoals eastward along the NECC, these subsurface anomalies outcropped in regions near the date line, initiating warm SST anomalies which further amplified while advectively extending into the equatorial wave guide. These subsurface-induced SST anomalies could plausibly trigger local coupled air-sea interactions producing

atmospheric-oceanic anomalies that developed and evolved in 1991, thus setting onset conditions for the 1991-92 El Nino.

A reduced-gravity, primitive-equation, upper-ocean general circulation model is then used to tie these tropical observations to the larger scale subtropical/tropical circulation. Velocity fields, and isopycnal and trajectory analyses help to understand the mean flow of mixed layer and thermocline waters between the subtropics and tropics. Thermocline waters can be traced to subtropical subduction sites and are shown to move toward the tropics almost zonally across the basin, succeeding in flowing toward the equator only along relatively narrow north-south conduits. The low-latitude western boundary currents serve as the main southward circuit for the subducted subtropical thermocline water, with an important confluence of these pathways occurring in the sub-thermocline regions of the NECC in the far western tropical Pacific. These flows are shown in the model to be swept eastward by the deeper branches of the North Equatorial Countercurrent, consistent with the NCEP re-analysis data, finally penetrating to the equator in the central and eastern Pacific.

These results differ markedly from the delayed oscillator physics in that a major role can be played by the eastward advection of off-equatorial subsurface thermal anomalies and their outcropping along the NECC subsurface pathways, not necessarily involving wave reflections along the western boundary for triggering El Nino. This mechanism explains the observed, otherwise unaccounted for, surface warming near the date line during late 1989 and early 1990.

Mean Circulation and Upwelling in the Equatorial Pacific Ocean Determined by Inverse Methods

Bernadette Sloyan, Gregory Johnson and Billy Kessler, PMEL

An inverse model for the central and eastern equatorial Pacific Ocean (180E - 95W, 8S - 8N) is developed. The region is divided into 18 boxes with zonal boundaries at 8S, 2S, 2N and 8S and meridional boundaries at 15 degree intervals between 180E and 95W. Seventeen isopycnal layers are chosen to span the water masses and the equatorial currents. Mass, temperature and salt are conserved in all layers. The zonal and meridional geostrophic velocities are estimated from a combination of ADCP velocities and the density field. Ekman transports are also added to the initial zonal and meridional fluxes. COADS air-sea climatologies are used to compute the air-sea transformation across outcropping isopycnal surfaces. Levitus climatology is used to calculated the mean area, temperature and salinity of isopycnal layers within each box. The subsequent system of simultaneous equation is solved for the unknown reference velocities, diapycnal fluxes and corrections to the air-sea climatologies.

The broad scale circulation of the model is consistent with previous studies. The eastward transport of the Equatorial Under Current (EUC) increases from 26 Sv at 170W to 31 Sv at 125W. Between 125 W and 95 W the mass transport of the EUC decreases reaching a minimum of 14 Sv at 95W. Coincident with the eastward increase and decrease of the EUC is a westward increase and decrease of the mass transport associated with the South Equatorial Current (SEC). At 95W the SEC transport is -14 Sv, increasing to -40 Sv at 140W and then decreasing to -20 Sv by 180E. The mass transport associated with the Southern and Northern Surface and Subsurface Counter Currents also shows some longitudinal variation. The interior and air-sea diapycnal fluxes indicate how mass, heat and salt are exchanged between the different isopycnal layers and the equatorial and subtropical Pacific Ocean.

Dynamical and Thermal Responses of the Ocean Surface Layers to the Atmospheric Variability Associated With the Pan-Atlantic Decadal Oscillation

Youichi Tanimoto, Hokkaido University, Japan

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To investigate the thermal and dynamical responses of the upper ocean to the decadal atmospheric variations over the Pan-Atlantic region that contains the North Atlantic Oscillation (NAO) and the tropical dipole, we examined the subsurface temperature field from the hydrographic observations and the atmospheric forcing terms from the NCEP (National Center for Environmental Prediction) reanalysis dataset.

While a northward gradient of sea surface temperature anomalies is large (the tropical dipole appears), the reduced (enhanced) wind speed over the northern tropics and south of Greenland (off east coast of the United States) induces a suppressed (an active) turbulent surface heat release in the same regions. The Ekman heat convergence, calculated from SST and wind stress fields, works as an additional thermal forcing in most regions of the North Atlantic. Ekman pumping velocity calculated as the dynamical response to the anomalous wind stress field indicates reduced downwelling (upwelling) in the region of the subtropical gyre (the subpolar gyre and the northern tropics).

The thermal structure of the upper ocean was discussed with the viewpoint of causal consistency. Although the temperature anomalies of the tropical ocean showed the dipole structure not only at the surface but also in the near-surface layers, the weaker magnitude of wind stress anomalies and the deeper thermocline depth seemed to interfere the predominance of a center of action south of the equator. In the interior region of the subtropical gyre, temperature anomalies have different polarities between in the near-surface layers to about

150-200 m depth and beneath it, which is associated with the dominant terms: the total heat flux and Ekman pumping, respectively. In contrast, the significant anomalies in the interior region of the subpolar gyre have a deeper structure with the same polarity because all examined forcing terms work in a same sense.

Air-Sea Interaction Over the Indian Ocean Induced by the Indonesian Throughflow

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The effects of the Indonesian throughflow on the upper thermocline circulation and surface heat flux over the Indian Ocean are presented for a 3-D ocean model forced by two different monthly climatologies, as they show interesting differences, which could have implications for long-term variability in the Indian and Australasian monsoons. In the ECWMF-forced model, there is little impact on the annual mean surface heat flux in the region surrounding the throughflow exit straits, whereas in the SSM/I-forced model, a modest throughflow of less than 5 Sv. Over the upper 300 m induces an extra 10-50 W m^{-2} output. In the SSM/I-forced model, there is insignificant penetration of the throughflow into the northern Indian Ocean. However, in the ECMWF-forced model, the throughflow induces a 5-10 W m^{-2} increase in output over the Somali Current in the annual mean. These differences are attributed to differences in the strength and direction of the Ekman transport of the ambient flow, and the vertical structure of the transport and temperature anomalies associated with the throughflow. In both models, the throughflow induces 5-30 W m^{-2} increases in net output over a broad swathe of the southern Indian Ocean, with increases reaching up to 100 W m^{-2} over the Agulhas Current retroflection.

In general, seasonal variations in the throughflow’s effect on the net surface heat flux are attributed to seasonal variations in the ambient circulation of the Indian Ocean, specifically in coastal upwelling along the south Javan, west Australian, and Somalian coast, and in the depth of convective overturning in the band for 40°S to 50°S, and its sensing of the mean throughflow's thermal anomaly. The seasonal differences plus annual mean differences yield maximum differences in net surface heat output in boreal summer. Values exceed 40 W m^{-2} over the southern Indian Ocean interior, 60 W m^{-2} over the Agulhas retroflection and immediate vicinity of the exit channels, and reach 30 W m^{-2} over the Somali Jet.

Upper Ocean Pathways and Decadal Variability in the Pacific

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Analysis of sea surface temperature (SST) data indicates that SST in both the equatorial and mid-latitude North Pacific has strong decadal variability with a characteristic time scale of ~14 years, as well as distinct interdecadal variability. The decadal variability in the mid-latitude North Pacific is shown to be out of phase with that in the equatorial Pacific. Moreover, the so-called climate regime shift in the 1970s is found to be the result of the constructive interference of the decadal and interdecadal signals in SST.

The Levitus climatology is used to construct the three-dimensional mid-latitude-equatorial subduction pathways that link the extratropics to the equatorial Pacific. The extratropical and equatorial pathways in the North and South Pacific are asymmetric with respect to the equator, because the ITCZ lies north of the equator and manifests itself as a potential vorticity barrier. As a result, the subduction exchange window of relevance to decadal variability lies much further away from the equator in the extratropical North Pacific than in the South Pacific. In addition, the subduction exchange window is much larger in the North Pacific versus the South. Taken together with the characteristics of the decadal SST variability, this suggests that the extratropical and equatorial coupling on decadal time scales is mainly between the extratropical North Pacific and the equatorial Pacific.

Trajectory analyses along the mid-latitude-equatorial subduction pathways indicate that the time scale of this oceanic teleconnection is ~ six years via the mid-latitude subduction exchange window. This characteristic time scale and the relative phases of the observed decadal variability between the mid-latitude North Pacific and the equatorial Pacific are consistent with a negative delayed action oscillator for the observed decadal variability. Although this mechanism was originally proposed by Gu and Philander (1997) to explain the interdecadal variability, here we have developed a self-sustaining oscillator that is consistent with the observed decadal variability. However, our results suggest a self-sustained internal mode of this type is not consistent with lower frequency variability because there is a mismatch in the time scale of the interdecadal variability and the subduction exchange time scale.

VII. Appendix A: Workshop Agenda

Monday October 9

9 - 9.30 a.m.	Arrival and registration ($80 fee)
9.30 - 10 a.m.	Welcome by the Director of the Istituto Veneto Scienze Lettere ed Arti. *Alessandro Franchini*
	Introduction of participants and presentation of workshop objectives. *Paola Malanotte-Rizzoli*
	CLIVAR perspective and charge to participants. *Antonio Busalacchi*

Pacific Ocean - Overview presentations

Chairperson: William Kessler

10 - 11 a.m.	Observations. *Gregory Johnson*
11 - 11.30 a.m.	Coffee break
11.30 - 12.30 a.m.	Ocean Models. *Jay McCreary*
12.30 - 1 p.m.	General discussion
1 - 2. 30 p.m.	Lunch
2.30 - 3.30 p.m.	Coupled systems. *Antonio Busalacchi*
3.30 - 4 p.m.	Coffee break
4 - 4.30 p.m.	General discussion
4.30 - 6 p.m.	Poster session

Wine will be served during the poster session.

Tuesday October 10

Indian Ocean - Overview presentations

Chairperson: Roxana Wajsowicz

9.15 - 10.15 a.m	Observations. *Fritz Schott*
10.15 -10.50 a.m.	Coffee break
10.50 - 11.50 a.m.	Ocean Models. *Jay McCreary*
11.50 - 12.40 p.m.	General discussion
12.40 - 1 p.m.	Summary report of the workshop on decadal variability, Joint CLIVAR-JSC Working Group on Coupled Modelling. *Vikram Mehta*
1 - 2.30 p.m.	Lunch
2.30 - 3.30 p.m.	Coupled systems. *Weiqing Han (Peter Webster)*
3.30 - 4 p.m.	Coffee break
4 - 4. 30 p.m.	General discussion
4.30 - 6 p.m.	Poster session

Wine will be served during the poster session.

Wednesday October 11

Atlantic Ocean - Overview presentations

Chairperson: William Johns

9.15 - 10.15 a.m.	Observations. *Robert Molinari*
10.15 - 10.50 a.m.	Coffee break
10.50 - 11.50 a.m.	Ocean models. *Claus Boening*
11.50 - 12.30 p.m.	General discussion
12.30 - 2 p.m.	Lunch
2 - 3 p.m.	Coupled systems. *Shang-Ping Xie*
3 - 3.30 p.m.	Coffee break
3.30 - 4.30 p.m.	Plenary session on differences/similarities among the three oceans with relative summaries presented for: · observations *(Fritz Schott)* · ocean models *(Jay McCreary)* · coupled systems *(Tony Busalacchi)*
4.30 - 5. p.m.	Formation of three working groups (WG) · Pacific WG; Co-chairs: *Michael McPhaden and Lewis Rothstein* · Indian WG; Co-chairs: *Rana Fine and Rui-Xin Huang* · Atlantic WG; Co-chairs: *Silvia Garzoli and Barry Klinger* with charges (issues, future needs, recommendations) overall (flexible) composition and assignment of chairpersons.

8 p.m.	Social dinner

Thursday October 12

(Coffee will be available at 10.30 a.m. and 3 p.m. without formal coffee breaks)

9.15 - 12 a.m.	WGs meet
12 - 1p.m.	WG preliminary reports and general discussion
1 - 2.30 p.m.	Lunch
2.30 - 4 p.m.	WG meet
4 - 5 p.m.	WG summary reports and general discussion

Friday October 13

9.15 - 10.30 a.m.	Plenary session
10.30 - 11 a.m.	Coffee break
11 a.m. - 1 p.m.	Plenary session with final summary statements
	Discussion of report structure, timeline and individual assignments

Closure of workshop

VIII. Appendix B: List of Participants

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