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Ocean modeling for ENSO

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Abstract. Tropical ocean modeling has played a major part in the development of our knowledge of El Niño-Southern Oscillation (ENSO) during the Tropical Ocean-Global Atmosphere (TOGA) decade. Although the foundations had already been laid, it was only with the impetus from TOGA that tropical ocean modeling was able to develop so extensively. In this paper we discuss the development of the wide range of tropical ocean models in use today, from the simple to the complex; the ways in which their abilities to reproduce different phenomena have been assessed; and the ways in which they are being used to better understand and predict the behavior of the coupled ocean-atmosphere system in the tropics. Ocean model development is far from finished, however. Outstanding issues such as forcing fields, model improvements, testing strategies, and applications are also discussed. There is at least as much work to be done in the next decade as was achieved in the last.

1. Introduction

Tropical Ocean-Global Atmosphere (TOGA) was a 10-year international research program designed to study the principal cause of interannual variability in the Earth's climate, namely, the interactions between the tropical oceans and the atmosphere. El Niño-Southern Oscillation (ENSO) is the most visible example of such an interaction. Two of the principal aims of TOGA were to describe and understand those aspects of the coupled ocean-atmosphere system that give rise to predictability and to study the feasibility of modeling the system so as to be able to make predictions. Numerical modeling of the ocean provides a major tool for helping to understand phenomena such as ENSO, either through studies using ocean models alone or as part of a coupled ocean atmosphere system. Further, it is of course necessary to use an ocean model in any model-based prediction scheme for ENSO. Oceanic modeling has therefore been a vital component of the TOGA program, and in this paper we review what progress has been made and some of the outstanding issues that remain.

We start in section 2 with a brief overview of the situation before TOGA. Section 3 describes the development of tropical ocean models during the TOGA period. This is broken down into a consideration of simple, intermediate, and general circulation models, with an additional subsection on the parameterization of physical processes, chiefly mixing. Section 4 concerns the testing or validation of models; this is done by looking critically at how well models are able to simulate the tropical ocean in terms of the mean seasonal cycle, interannual variability, and also higher-frequency phenomena. Section 5 is about the application of ocean models to TOGA issues, with a look at some of the areas in which the models have been used. The issues covered include oceanic processes within ENSO, surface fluxes, and data assimilation and coupled model initialization. We conclude in section 6 with a summary, which restates five of the major issues in ENSO ocean modeling.

There are two strategies by which a multiauthor review paper could be written: a consensus document, uniform in style and always espousing a mutually agreed perspective, or a multisection paper, which retains some of the diversity of outlook and writing of the authors. We have followed the latter approach, though, of course, the final paper has been formed into a coherent whole. We believe we have covered most of the ground relevant for ENSO ocean modeling in a reasonably ordered fashion, but the reader is encouraged to read the whole paper rather than pick out particular sections of interest. A fuller perspective and more interesting sidelines are likely to be obtained in this way, and we hope that the paper as a whole will be informative and stimulating.

2. Ocean Modeling Prior to TOGA

The International Union of Geodesy and Geophysics (IUGG) quadrennial report on equatorial oceanography by *Cane and Sarachik* [1983] provides a very useful summary of the state of ocean modeling just before the TOGA program. There were little ocean subsurface data available for any of the tropical oceans, so thermocline depth variability was the topic of primary interest. The importance of SST variability for the role of the ocean in the coupled air-sea system was understood, but model studies to investigate the processes responsible for its variations were in their earliest stages. Considerable theoretical work on the forced linear equatorial circulation, both the wave response and the steady circulations that are possible when friction is included, had been done. Wave propagation and

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reflection phenomena (particularly Kelvin/Rossby forcing and reflection) held the interest of many. The emphasis had been on process studies rather than on simulation studies (although *Busalacchi and O'Brien* [1981] had published the first of their efforts to simulate tropical Pacific thermocline variability by imposing monthly mean winds on a reduced gravity layer model). *Rasmussen and Carpenter* [1982] (hereinafter referred to as RC) had recently published their important study of the ocean surface composite structure of the previous five ENSO periods, but the modeling community was just beginning to think about how to model the phenomena described in RC. Most ocean thinking about ENSO still was based on Wyrtki-esque ideas of relaxing trade winds, the associated eastward propagating Kelvin waves, and (sometimes) their subsequent reflected coastal and Rossby waves.

The models in use were much the same as employed now: the shallow water equations provided the theoretical basis for much of our understanding of the linear limit of equatorial dynamics, reduced gravity models were in use to study the large-scale patterns of thermocline/sea level variation, intermediate models were being used to bring mixed layer physics to the study of upper ocean currents and thermal variations, and *Bryan* [1969] / *Semtner* [1974] / *Cox* [1984] primitive equation general circulation models were in use for studies in which the full nonlinearity of the momentum and heat equations were deemed necessary. The ocean was being studied as a forced fluid, with atmospheric forcing imposed, rather than as part of the coupled ocean-atmosphere system.

Knowledge of the atmospheric forcing was a real problem. Theory made clear that the equatorial oceans were very sensitive to the space and time variations of the wind stress, and early efforts to evaluate different analyses of the monthly mean winds over the tropical Pacific revealed that the uncertainties were of the same magnitude as the winds themselves. *Cane and Sarachik* [1983 p.1139] comment on the need for accurate wind data sets and contemporaneous ocean observations, "Until both of these data sets are available, the confrontation between theory and experiment must necessarily be incomplete". As this volume indicates, these needs were taken to heart by the TOGA community, and the most widespread, serious data collection effort in the history of basin-scale oceanography was undertaken.

3. Tropical Ocean Model Development

3.1. Simple Models

For a variety of reasons, simple models such as reduced-gravity models or related models incorporating several low-order baroclinic modes have proven quite adept at advancing our understanding of tropical ocean dynamics. Since most of the variability in the tropical oceans is above the thermocline, the upper ocean circulation can be described efficiently by a few baroclinic modes and is often characterized as a two-layer system where deep subthermocline fluctuations are neglected within the context of the reduced-gravity formulation. The relatively simple physics of these models has meant that the more difficult issues to be confronted in general circulation models, such as mixing parameterizations and problems with a diffusive thermocline, could be overlooked since only the deviations about a specified pycnocline are simulated. Going into the TOGA decade, there was ample evidence that the dynamic response of the tropical ocean circulation (e.g., changes to vertically integrated quantities such as sea level, dynamic height, and upper ocean heat content) could be simulated with a fair degree of accuracy by simple models (compare *Busalacchi and O'Brien* [1980, 1981] and *Cane* [1984] for the seasonal and interannual response of the tropical Pacific; *Busalacchi and Picaut* [1983], for the seasonal response of the tropical Atlantic Ocean; and *Gent et al.* [1983] for the semiannual variability in the Indian Ocean). These studies suggested that the large-scale dynamic topography and off-equatorial currents in the tropics were essentially wind driven, linear, and hence deterministic. Equatorial wave dynamics were shown to be important for transmitting the effects of wind energy from one end of a tropical ocean basin to the other. Such propagation would take place on a rapid timescale (of order months) with significant implications for the redistribution of mass and heat at low latitudes.

During the TOGA decade the availability of sea level observations from tide gauges, dynamic height information from expendable bathythermograph (XBT) and TOGA Tropical Atmosphere-Ocean (TAO) observations, and basin-scale sea surface topography information from satellite altimeters spurred the continued use and refinement of reduced-gravity models in support of simulation studies. Atlantic and Pacific Ocean studies such as those described in works by *du Penhoat and Treguier* [1985], *Busalacchi and Cane* [1985], and *Bigg and Blundell* [1989] demonstrated that the amplitude and phase of such simulations could be improved by including more than just the first baroclinic mode, although *Kessler and McPhaden* [1995] showed how suitably chosen single-mode models can behave comparably to multimode models in certain circumstances. In the Indian Ocean, climatological wind data were first used by *Luther and O'Brien* [1985] and *Woodberry et al.* [1989] to study the wind-forced response to the full seasonal cycle. As discussed in section 4.2, hindcasts of the interannual variability cf the wind-forced dynamics continued for the Pacific Ocean during TOGA and began for the tropical Atlantic and Indian Oceans. In addition to hindcast experiments, this class of model was also extended for the study of equatorial wave processes, for evaluating deficiencies in surface wind data, for observing system simulation experiments (OSSE), and for ocean data assimilation studies. In view of the central importance of equatorial wave dynamics to theories of ENSO such as the delayed action oscillator [*Suarez and Schopf*, 1988], analytical and numerical reduced-gravity solutions were used to investigate the behavior of equatorial waves with regards to nonlinearities [*Greatbatch*, 1985; *Ma*, 1992], the eastern boundary response [*Bigg and Gill*, 1986], the western boundary response [*Clarke*, 1991], wave-mean flow interactions [*McPhaden et al.*, 1986], topographic scattering [*McPhaden and Gill*, 1987], and zonally varying background stratification [*Busalacchi and Cane*, 1988]. The fact that these models were computationally efficient also meant that multiple simulations could be performed and used to evaluate the quality of various wind products on seasonal and interannual timescales [*McPhaden et al.*, 1988b; *Busalacchi et al.*, 1990]. Similarly, reduced-gravity models have been used to evaluate in situ sampling as part of OSSEs that assessed potential aliasing by the zonal spread between ship track XBT observations [*McPhaden et al.*, 1988a], the potential redundancy of individual XBT transects [*Bennett*, 1990], and the placement of TOGA TAO moorings [*Miller*, 1990]. As a result of the two-layer approximation, sea level and thermocline depth observations are easily projected onto the vertical. This has permitted some of the first oceanographic applications of advanced data assimilation techniques, such as the Kalman filter [*Miller and Cane*, 1989] and the adjoint [*Sheinbaum and Anderson*, 1990a, b], to take place in tropical ocean models.

3.2. Intermediate Models

The term "intermediate model" is used to denote a model that is more complex than the simple process models but still falls short of the complexity of a general circulation model (GCM). The pursuit of an understanding of ENSO led quite naturally to the development of intermediate models. In the 1970s and 1980s it was noticed that linear, shallow water theory provided an explanation for the observed changes in sea level height and thermocline depth [*Wyrtki*, 1975; *Busalacchi and O'Brien*, 1980]. This observation suggested that, to leading order, an ocean model of ENSO only needed to contain linear, wind-driven motions that could be modeled in terms of a limited number of vertical modes of a stratified ocean. The complexity of an ocean GCM could be avoided.

The determination of sea surface temperature (SST) was, however, not so simple. The correlation between thermocline depth and SST, which is used in the simple models, can explain only part of the observed SST variability. Other processes, including horizontal advection, surface heat fluxes, and entrainment into the ocean mixed layer, were also known to be important. The natural way to address the problem of determination of SST was to combine shallow water theory for the dynamics with a relatively complete equation for the SST.

An early attempt to formulate an intermediate model was made by *Anderson and McCreary* [1985]. The ocean component of their coupled model included fully nonlinear shallow water dynamics and a nonlinear equation for the total upper layer temperature (as opposed to temperature anomalies relative to climatology). The temperature equation accounted for horizontal mixing, relaxation to prescribed meridionally varying temperatures to account for surface fluxes, and entrainment from below, analogous to *Kraus and Turner* [1967] style mixing. Coupled to a simple atmosphere model, the model produced low-frequency oscillations that propagated slowly eastward.

A quite different model, and still the one most widely used, is that of *Zebiak and Cane* [1987] hereinafter referred to as ZC). In this model the currents are given by linear shallow water theory and a single vertical mode. However, the currents determined in this way cannot represent the surface intensification of the wind-driven circulation and hence could not correctly simulate their effect on SST. ZC dealt with this by assuming the directly wind-driven surface currents could be described by a linear, frictional, Ekman balance. They then separated this component from the complete equations and placed it in a fixed depth mixed layer. The total currents within the mixed layer are then given by the wind-driven component plus a generally weaker geostrophic component related to the surface topography.

The temperature of the mixed layer, which is the SST, is determined with a fully three-dimensional, nonlinear, thermodynamic equation. The mixed layer has a fixed depth, so the entrainment at its base is given by the velocity divergence in the layer. The temperature of the water entrained is determined in terms of the thermocline depth through the use of simple empirical relations. These are based on the observed correlation between the temperature just below the surface and the depth of the thermocline and bring in the mechanism identified in the simple model of *Hirst* [1986]. The surface flux is assumed to damp SST anomalies on a timescale of about 125 days. In the coupled model used to simulate and predict ENSO this ocean model is coupled to a simple atmosphere model in which the winds are driven by convective heating, which is parameterized in terms of the SST and the surface wind convergence.

An important simplification of the ZC model is that it is an anomaly model that simulates the SST departure from a specified annual cycle. The problem of correctly simulating the seasonal cycle of SST, thermocline depth, surface winds, and so on is sidestepped.

All other intermediate models have followed the lead of the ZC model. For example, *Kleeman* [1993] has developed a model that falls somewhere in between the ZC model and the simple models. His model also uses shallow water theory (but with fewer horizontal modes) and a frictional surface layer to determine thermocline displacements and surface currents. Kleeman simplifies the SST equation by assuming a fixed meridional structure to the SST anomalies. Further, the influence of entrainment on the SST, which is parameterized in ZC with a single term, is split in two by Kleeman. He has a term that describes in an empirical manner the influence of thermocline displacements on SST and another that describes the effects of Ekman upwelling. The latter term is done in an analogous way to ZC.

The most recent example of a ZC type model designed for simulation of ENSO anomalies is that of *Y-Q. Chen et al.* [1995]. While the models mentioned above use a single vertical mode, Chen et al use two. Previous work had shown that two modes were needed to adequately simulate thermocline variability [*Cane*, 1984]. The Chen et al model also includes a parameterization of subsurface temperature, which follows from that of *Seager et al.* [1988] (see below) and is more empirically based than the original scheme of ZC. These two modifications allowed them to use a more physically reasonable parameter set than ZC. For example, the unreasonably large drag coefficient which ZC use to compensate for the absence of higher vertical modes, is not necessary in the Chen et al model. The Chen et al model, run in hindcast mode, produces a very impressive simulation of the last few decades of SST variability. Despite this, when coupled to the atmosphere component of the coupled ZC model, the coupled oscillation it produced was less realistic than with the original ZC model. Presumably, this result emphasizes just how much the intermediate models have been finely tuned (often unconsciously) to reproduce reality. Improvements in either component always require a thorough retuning of the complete model.

Intermediate models have also been used in simulations of the tropical climatology. The first such effort was that of *Seager et al.* [1988] (hereinafter referred to as SZC) using the ZC model. This required replacing the model's parameterization of subsurface temperature anomalies with one that related the total subsurface temperature to the simulated climatological thermocline depth. Another intermediate model designed to simulate both climatology and interannual variability is that of *Wang et al.* [1995]. This model differs from other intermediate models through use of the primitive equations in reduced gravity form, a parameterization of subsurface temperatures that is more internally consistent, and replacement of a fixed mixed layer depth with one determined using Kraus-Turner type physics [see also *Chang*, 1994]. The use of more realistic mixed layer physics in particular is a potentially considerable advantage over the other intermediate models and may aid in simulation of the seasonal cycle.

3.3. Ocean GCMs

At the top of the heirarchy of tropical ocean models lie the ocean general circulation models (OGCMs). Essentially, these are models based on the primitive equations, and they include the basic physics required to reproduce the large scale ocean circulation. They are often formulated with a number of fixed vertical levels (e.g. the Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model (MOM) code, but can also be written to use isopycnic levels as a vertical coordinate system (e.g. the OPYC model), [*Oberhuber*, 1993]. The boundary between OGCMs and other sophisticated ocean models is not always well defined, and the coverage of this section is appropriately broad.

Early work on high-resolution GCM modeling of the equatorial oceans took place in the United States, with pioneering studies by *Semtner and Holland* [1980] and *Philander and Pacanowski* [1980]. These were idealized studies but were made at a resolution sufficient to resolve transient and turbulent activity. The computer code used was the GFDL general circulation model that originated with work by *Bryan* [1969] and had been rewritten by *Semtner* [1974]. Subsequently, the code has been transformed by *Cox* [1984] and then by others (*Pacanowski et al.* [1991], *Pacanowski* [1995]) into the GFDL MOM code that is available today.

Work continued at GFDL to develop a realistic model that could be used to simulate the Pacific and Atlantic Oceans [*Philander and Seigel*, 1985; *Philander and Pacanowski*, 1986]. It was already evident that the structure of the Equatorial Undercurrent was sensitive to vertical mixing processes, and the very simple specifications common at that time were inadequate. As part of their model development, therefore, *Pacanowski and Philander* [1981] (hereinafter referred to as PP) introduced a Richardson-number dependent mixing scheme of a particular form, with coefficients motivated by fitting the model to observations in three case study situations. This mixing scheme, which will be referred to as PP mixing, was widely adopted by other modelers and is still in common use today.

The configuration for the tropical Pacific settled on by *Philander et al.* [1987] has a 33 km meridional resolution between 10°N and 10°S (i.e. about 1/3°), increasing gradually to 200 km by 25°N. Zonal resolution is 100 km, and 27 vertical levels are employed, with a spacing of 10 m in the upper 100. Horizontal mixing is in the usual harmonic form, but vertical mixing uses the PP specification. This model was used not only to simulate the Pacific in some detail [*Philander et al.*, 1986, 1987] but also for coupling to an atmosphere model to investigate tropical variability [*Philander et al.*, 1992]. An important finding was that the ENSO behavior of the high-resolution system differed in character from the interannual variability produced by the same atmosphere coupled to a lower-resolution ocean (*Lau et al.*, [1992], who used an ocean resolution of 4.5° by 3.75°), providing the first demonstration of the importance of ocean model resolution in a coupled GCM system.

A differently configured OGCM was also developed at GFDL, this time by *Rosati and Miyakoda* [1988]. The model used the same basic code [*Cox*, 1984], but was set up as a global 1° by 1° model, with 1/3° meridional spacing in the equatorial region. The model was specifically designed for possible coupling to an atmosphere and careful attention was paid to the mixing physics. A Mellor-Yamada 2.5-level turbulence closure scheme was included [*Mellor and Yamada*, 1982], and also, nonlinear horizontal viscosity was included [*Smagorinsky*, 1963]. It was found that high-frequency forcing (every 12 hours) from the atmosphere was needed to drive the turbulent mixing effectively. The model produced better SST, surface currents, and mixed layer depths than the same model using constant vertical mixing coefficients. The subsurface simulation at the equator was unsatisfactory, however, with a diffuse thermocline and a slow Equatorial Undercurrent.

The *Rosati and Miyakoda* [1988] OGCM was used as the basis for a global oceanic data assimilation system by *Derber and Rosati* [1989]. An important change was a modification to the *Mellor and Yamada* [1982] mixing by imposing a limit on the turbulent length scale dependent on the local static stability, following *Galperin et al.* [1988]. This reduces the mixing through the thermocline, giving a more realistic vertical structure to the upper ocean, and it was also found to produce smoother horizontal fields.

The Pacific basin ocean model of *Philander et al.* [1987] was combined with the analysis scheme of *Derber and Rosati* [1989] to create an operational tropical Pacific analysis system at the National Centers for Environmental Prediction (NCEP) [*Leetmaa and Ji*, 1989; *Ji et al.*, 1995]. Subsequently, the ocean model was coupled to a version of the NCEP medium-range forecast model, which enabled the production of real time seasonal forecasts of tropical Pacific SST [*Ji et al.*, 1994a, b]. This system is in operational use at the present time. The ocean model itself is still basically identical to that developed by Philander et al in the early 1980s.

A number of other groups in the United States used the GFDL code for modeling the equatorial Pacific. Harrison and coworkers at the Pacific Marine Environmental Laboratory (PMEL) used the *Philander et al.* [1987] version of the model in numerous investigations into the sensitivity of ocean simulations to different aspects of the forcing [e.g., *Harrison et al.*, 1989; *Harrison*, 1989, 1991; *Harrison and Craig*, 1993]. The Philander-configured version of the model has also been used at University of California Los Angeles (UCLA) in the development of a coupled GCM. It was found that in coupled mode the *Mellor and Yamada* [1982] mixing gave superior results in the subsurface equatorial structure when compared to the PP scheme [(*Ma et al.*, 1994]. In uncoupled mode, however, the PP scheme performed adequately [*Halpern et al.*, 1995]. The GFDL model has also been implemented at the Center for Ocean-Land-Atmosphere Studies (COLA), where a 1.5° global model has a 0.5° equatorial resolution to allow ENSO studies [*Huang and Schneider*, 1995]; the same model is also used in a Pacific basin only version [*Kirtman and Schneider*, 1996].

A separate group to use the basic GFDL code (in its *Cox* [1984] version) to develop a high resolution Pacific model was based at the U.K. Meteorological Office (UKMO) [*Gordon and Corry*, 1991]. This also had a 1/3° meridional resolution near to the equator, but used a flat bottom topography and neglected the barotropic mode. Zonal resolution was 1.5°, reducing to 0.5° near the eastern and western boundaries, and 16 vertical levels were used. Vertical mixing was specified using PP mixing plus penetration of solar radiation (which helps mix the surface layers). Later versions of the model included an additional *Kraus and Turner* [1967] mixing scheme to allow for the wind-turbulence mixing of the surface layers [see *Davey et al.*, 1994; *Kraus and Turner*, 1967]. This model was also coupled to an atmosphere GCM (AGCM), and although early integrations gave a good mean seasonal cycle, interannual variability was very weak [*Gordon*, 1989]. More recent versions of the coupled system show better variability [*Ineson and Davey*, 1997], even though the ocean component is little changed.

The GFDL code has also been adopted by tropical modelers in Japan. As described in *Stockdale et al.* [1993], Kitamura et al implemented a 2.5° by 1° model at the Meteorological Research Institute (MRI), using *Mellor and Yamada* [1982] vertical mixing and 19 levels with a fine spacing near the surface (5 levels in the top 20 m). Also in *Stockdale et al.* [1993], Yamagata and Masumoto describe a 0.5° by 0.5° model of the tropical Pacific, which uses PP mixing, and standard Laplacian mixing in the horizontal. This model is notable for very strong Legeckis wave activity. The philosophy of much of Yamagata and Masumoto's work has been to investigate regional features with relatively high resolution models (either 0.25° by 0.25° or 0.5° by 0.5°), partly as a test of the models but also to learn more about possible areas of ocean-atmosphere coupled interaction [*Umatani and Yamagata*, 1991; *Masumoto and Yamagata*, 1991]. More recently, the model has been extended to the Indian Ocean and used to study the flows around Indonesia [*Masumoto and Yamagata*, 1993]. An Atlantic version has also been used to study the seasonal cycle of the Guinea and Atlantic domes [*Yamagata and Iizuka*, 1995].

By no means are all ocean GCMs related to the code developed over the years by GFDL. One very successful OGCM is that developed at the Laboratoire d'Océanographie Dynamique et de Climatologie (LODYC) in Paris, under the leadership of Pascale Delecluse [*Andrich et al.*, 1988; *Blanke and Delecluse*, 1993; *Delecluse et al.*, 1993]. The model uses a rigid lid assumption (following *Bryan* [1969]) but is discretized on an Arakawa C grid, uses a tensorial formalism to ensure second-order accuracy on arbitrary curvilinear orthogonal grids [*Marti et al.*, 1992], and uses a discretization of the vorticity term which ensures conservation of potential enstrophy for non divergent horizontal flows [*Sadourny*, 1975]. Although global versions of the model exist, most work has concentrated on high-resolution tropical studies, primarily of the Atlantic and Pacific [*Reverdin et al.*, 1991; *Stockdale et al.*, 1993]. The usual configuration has an equatorial resolution of 1/3° by 3/4° and 30 levels and uses a turbulent kinetic energy (TKE) parameterization with a 1.5-level closure for vertical mixing, following *Gaspar et al.* [1990]. Much work has been done on model tuning and testing in an attempt to produce a realistic tropical simulation [*Blanke and Delecluse*, 1993; *Braconnot and Frankignoul*, 1994].

Another major center of ocean model development is the Max-Planck-Institut (MPI) für Meteorologie in Hamburg, Germany. The two major OGCMs there are OPYC, an isopycnic model developed by *Oberhuber* [1993]; and HOPE, a *z* coordinate model developed by Maier-Reimer and exploited for tropical studies by Latif and coworkers. OPYC has primarily been used for midlatitude and global modeling studies, but it has also been tested for its ability to simulate El Niño [*Miller et al.*, 1993], and in coupled mode it is able to produce ENSO-like variability [*Lunkeit et al.*, 1994]. The HOPE model has been used extensively for tropical studies. An early version was employed by *Latif* [1987] and *Barnett et al.* [1991], but this was then reworked and renamed HOPE in 1992. The standard version of the model has a global domain, an equatorial resolution of 0.5° by 2.8°, and 20 vertical levels. Discretization is on an Arakawa E grid. A free surface is included, with the barotropic mode calculated by a direct matrix solver and a 2-hour time step. The baroclinic mode solution uses direct time stepping and an iterative semi-implicit method. Horizontal momentum mixing is partly harmonic and partly biharmonic, but the most important term in the Equatorial Undercurrent region is a shear dependent mixing term, similar but not identical to that introduced by *Smagorinsky* [1963]. Vertical mixing is primarily Richardson number dependent (in the form of *Pacanowski and Philander* [1981]), but there is an additional mixing term over the depth of the diagnosed surface mixed layer. Penetrative solar radiation is also included. HOPE has been used in coupled mode at MPI [*Latif et al.*, 1994], at the European Centre for Medium-range Weather Forecasts (ECMWF) [*Stockdale et al.*, 1994], and at Scripps Institute of Oceanography, California [*Schneider et al.*, 1996]. The original version in these studies suffered from a diffuse equatorial thermocline, but a change to the numerics (from upstream to centered differencing in the tracer advection equation) and retuning of the horizontal and vertical mixing produced a big improvement. The improved version (HOPE2) has been used for ENSO forecast experiments at ECMWF [*Stockdale*, 1997], for coupled seasonal cycle simulations at MPI [*Frey et al.*, 1997], and is also the basis of the latest hybrid coupled model forecasting system at Scripps (see the National Oceanic and Atmospheric Administration's (NOAA) long-lead forecast bulletin).

A somewhat different model designed for simulation of the equatorial oceans is that of *Gent and Cane* [1989]. This is a primitive equation, reduced-gravity model, and while not strictly speaking an OGCM seems appropriate for consideration here because it does seek to simulate the general circulation of the upper tropical ocean. This model has been used for studies of the heat budget and circulation of the tropical Pacific [e.g. *Gent*, 1991; *Brady and Gent*, 1994]. A typical configuration uses 0.25° meridional resolution at the equator (1.0° in the subtropics) and 1.0° zonal resolution. In the vertical, seven sigma layers overlie an inert deep ocean, and a surface mixed layer is also included. Horizontal viscosity is minimized by using only a sixteenth-order Shapiro filter to suppress grid noise. An improved version of the model was developed by *Chen et al.* [1994b], who incorporated a hybrid vertical mixing scheme based jointly on *Kraus and Turner* [1967] type mixing and Price's dynamical instability model [*Price et al.*, 1986]. This model was used by *Murtugudde et al.* [1996] for simulating the seasonal variation of SST in the three tropical oceans. It should not be confused with the isopycnal primitive equation model described by *Murtugudde et al.* [1995]. This latter model has also been used for studies in the tropical oceans, but little has yet been published, and the emphasis appears to be on developing the model for global applications.

Another model in the same class is the Poseidon model, developed at NASA Goddard Space Flight Center [*Schopf and Loughe*, 1995]. This is also a primitive equation, reduced-gravity model, but it uses an isopycnic vertical coordinate rather than sigma coordinates and has a global domain. The model has been shown to reproduce ENSO variability when forced with AGCM winds and has also been used in coupled ocean-atmosphere experiments [*Mechoso et al.*, 1995].

There has also been significant development of global OGCMs during the TOGA period, and although many of these models are designed for looking at longer timescales, they are often able to represent some sort of tropical interannual variability. Examples of coupled variability in coarse- resolution models are discussed by *Sperber et al.* [1987], *Philander et al.* [1989], *Meehl* [1990], and *Tett* [1995]. The variability is typically not very realistic, probably due in part to the low resolution affecting both the ocean mean state and the dynamics of equatorial transients. The latest generation of climate models is being developed with grid sizes that will allow reasonable resolution of equatorial ocean dynamics and thus has a better chance of capturing ENSO variability. Obtaining quantatively correct model ENSO variability will probably still be difficult. Nonetheless, it is indicative of the success of TOGA that what was once a brand new field of research is now considered a basic component of any reputable climate system model.

As can be seen, there is quite some variety in the tropical ocean GCMs that have been developed during the TOGA decade, as different groups have made different numerical and physical choices in the configuration of their models. The consequences of these differences are not well known, since few controlled comparisons have been carried out. A study instigated by the TOGA NEG [*Stockdale et al.*, 1993] demonstrated a fairly large range of behavior of different OGCMs run with identical forcing, although certain problems appeared to be common to all models. An assessment of the general performance of tropical OGCMs is given in section 4. The two aspects of an OGCM that are most important in determining its tropical performance are resolution (particularly horizontal resolution) and the treatment of mixing processes. The latter of these is a topic sufficiently difficult to merit separate discussion.

3.4. Parameterizations of Mixing

Mixing is important both in the horizontal and vertical (or isopycnal and diapycnal). The equatorial ocean includes strong current shears and temperature gradients in both orientations, and so it is clear that both types of mixing should be well parameterized. Horizontal mixing has generally been given the least attention so far, partly because it is often intimately tied up with issues of resolution and partly because it does not have such an immediate impact on SST. Many models run with the smallest horizontal viscosity consistent with numeric stability. Effective viscosity coefficients for high-resolution models are typically 1000 $m^2 s^{-1}$, and at this point, reasonable simulations seem possible. Some models use more complex types of mixing than Laplacian, e.g. the HOPE model which uses mixing proportional to the modulus of the horizontal strain rate tensor. Spatial plots of the effective mixing coefficient look plausible, but how important such refinements are for modeling the equatorial flow remains unclear. From a physical point of view the amount of horizontal mixing to be parameterized depends on the extent (if any) to which eddies are produced by the model. The largest equatorial eddies (known as tropical instability waves; see section 4.3) are already resolved by many models, although they are still sensitive to the sub grid scale parameterizations.

A systematic study with the LODYC model [*Maes et al.*, 1997] showed that the dynamics of the equatorial ocean changed substantially as the horizontal mixing of both momentum and heat were varied together between 10^4 and 10^2 $m^2 s^{-1}$. Despite the large changes in the character of the model simulation, the impact on SST was small. This was in contradiction to some earlier results [see *Stockdale et al.*, 1993], which varied only the mixing of momentum; in this case a modest but noticeable sensitivity of SST to horizontal momentum mixing was found. The difference in result may be due either to the role of horizontal diffusion (fixed in one case and varying in the other) or to the different character of vertical mixing in the models (relatively weak PP mixing in one and a strongly mixing TKE scheme in the LODYC model). Certainly Maes et al. point out the strong non linearities in the interaction between horizontal and vertical mixing and the equatorial circulation. The two results taken together suggest that the treatment of horizontal mixing in the equatorial ocean is an important part of a complex whole. It is undoubtedly an area where much more work is needed, but the interactions with the (equally tricky) vertical mixing make progress difficult.

There are two main approaches that have been taken in trying to come up with a good vertical mixing parameterization for the tropical ocean. The first is exemplified by *Peters et al.* [1988], and uses in situ micro turbulence measurements to try to relate mixing to large scale conditions. The second is represented by *Pacanowski and Philander* [1981], who chose the parameters for an empirical formula on the basis of the fit between model simulation and large-scale observations. Both approaches have serious limitations; observational data are inadequate to accurately define a comprehensive treatment of ocean mixing, and finding and accurately inverting a fit between large-scale model and observed fields are extremely difficult.

In situ data on turbulent mixing are sparse and patchy. Only a handful of measurement studies have been carried out, typically lasting only a few days and representing a limited area in a specific "synoptic" situation of large scale ocean behavior. Early indications that turbulent dissipation has a strong equatorial peak [*Crawford*, 1982] were contradicted by later results [*Moum et al.*, 1986]. Observations in the eastern Pacific suggested that the equatorial momentum balance can only be closed by invoking breaking internal gravity waves or some other such process [*Dillon et al.*, 1989]. It is now known that low Richardson number regions above the Equatorial Undercurrent core support intermittent and diurnally varying bursts of turbulence, and it is thought that these are triggered by internal waves. Observational evidence exists [*Hebert et al.*, 1992], but the precise nature of these waves and their effect on the equatorial zonal momentum balance remains uncertain [*Moum et al.*, 1992]. Away from the equator, observations of how momentum is distributed within (and below) the mixed layer have also been somewhat contradictory [*Price et al.*, 1987; *Chereskin and Roemmich*, 1991]. A further problem with observational studies is that they are generally indirect measurements, and hence the interpretation of the results is not beyond question. The intermittent nature of much mixing also makes suitable sampling and statistical treatment difficult, resulting in large error bars on any derived values. But despite all these caveats, it must be said that trying to invent or improve mixing schemes for an ocean model without any observational guidance is very difficult too.

Many OGCM modelers have at least played with coefficients in their mixing schemes, while others have tried out a number of different parameterizations. Several fairly basic facts emerge from such experiments. Foremost of these is that to maintain a sharp thermocline in the tropical oceans, vertical mixing is required to be small across the thermocline, while to keep a reasonable mixed layer structure, relatively large mixing is needed near the surface. Strong momentum mixing near the surface is also required to prevent excessively strong shallow surface currents being driven by the wind. A particularly sensitive oceanic feature is the Equatorial Undercurrent (EUC). This is strongly influenced by viscosity (both vertical and horizontal), as well as the wind stress and thermal structure of the upper equatorial ocean (which between them largely determine the pressure gradients which drive the undercurrent). Because the thermal structure is sensitive to both local and non local mixing and the structure of the EUC itself, the whole problem becomes very difficult to untangle. The relationship between zonal advection in the EUC, the position and strength of meridional upwelling/overturning circulations, and the temperature of the water in the equatorial cold tongue is an example of what might need to be sorted out.

The schemes used for vertical mixing in today's OGCMs can easily be summarized. *Pacanowski and Philander*'s [1981] Richardson number dependent scheme (PP mixing) ensures substantial mixing in parts of the EUC, while giving only small mixing through the tropical thermocline. The PP scheme can also give substantial mixing in the surface layers of the ocean, where wind variability creates appreciable vertical shears in velocity. *Peters et al.* [1988] (hereinafter referred to as PGT) proposed a different formulation, in which the functional relationship between Richardson number and mixing showed a much sharper transition between strong and weak mixing. The effect of this sharper transition on equatorial simulations seems fairly modest.

Schemes such as PP (or PGT) are often combined with a specific model of surface mixing, and in many models this is necessary to produce good simulations. The requirements of models for surface mixing can differ, depending on vertical resolution and the surface heat fluxes which are applied. One model of surface mixing used is that of *Kraus and Turner* [1967] (hereinafter referred to as KT). This KT mixing is basically a balance between surface input of turbulent kinetic energy (TKE) and its dissipation by entraining deeper water. A rather different strategy is taken in the HOPE model, where a surface mixed layer depth is simply diagnosed from the existing temperature structure (a criterion of 0.5°C difference from the SST is used), and then enhanced mixing is applied over this depth. The mixing added decays with depth much as the TKE decays with depth in the KT scheme, and this prevents excessive mixing in the presence of deep surface layers. The diagnostic nature of the scheme is not very satisfactory from a physical point of view, but it seems to work quite effectively.

A more ambitious approach to vertical mixing is to calculate the evolution of the turbulence directly, via the use of turbulence closure schemes. Equations can be written for the evolution of TKE, mixing length, etc. Each level of equations refers to a higher-order level, and a closure hypothesis must be made at some point. *Mellor and Yamada* [1982] introduced a 2.5-level turbulence closure scheme into oceanography, and this has been used e.g. by *Rosati and Miyakoda* [1988]. A 1.5-level turbulence closure scheme has been developed at LODYC, and it has been shown that this gives better results than PP mixing in simulations of the Atlantic [*Blanke and Delecluse*, 1993].

Most of the mixing schemes mentioned so far use the concept of local diffusive mixing and reduce to a specification of an eddy mixing coefficient. Other formalisms are possible, for example, the transilient mixing of *Stull and Kraus* [1987]. Here the transport of momentum and heat is represented by a matrix multiplication, which is able to represent non local processes. As discussed above, observational evidence suggests that such processes are sometimes important. The problem of reliably specifying what mixing takes place under which conditions remains.

Another recent parameterization is that proposed by *Large et al.* [1994]. This consists of separate but matched schemes for the surface mixed layer and the ocean interior. Physical processes considered include some that are often neglected in other schemes (non local transport in the boundary layer and double diffusion in the ocean interior). Internal wave mixing is represented as a constant diffusivity, however, and no account is made for specifically equatorial or other processes which may enhance this.

Vertical mixing in the surface layers is much helped in the models by the inclusion of the penetration of solar radiation. This is especially important in areas such as the west Pacific where net heat flux is small, and so even the small amount of sunlight at a depth of 50-100 m can be significant in driving convection. One interesting area that has not yet been included in TOGA ocean models is the influence of biology. Plankton can change the optical properties of water substantially, and this can influence SST by several degrees in extreme cases, although the biggest effects are probably in midlatitudes.

A final comment on vertical mixing is its relationship to isopycnal or layer type ocean models. There are physical arguments as to why oceanic mixing is a largely isopycnic process, with only relatively small amounts of diapycnic mixing. For a z coordinate model this can cause problems in areas of steeply sloping isotherms, although the mixing tensor can be rotated to allow for this [*Redi*, 1982; *Gent and McWilliams*, 1990]. For an isopynic model the diapycnal flux is explicitly controlled, and in strong upwelling regions such as the equator the diapycnal flux will often take the form of entrainment. In global scale modeling, it is often a problem to limit diapycnal mixing in boundary currents and flows over topography, and isopycnal models have some natural advantages over their z coordinate cousins in this regard. In equatorial modeling, vertical/diapycnal mixing is an integral part of the problem, and whichever type of model is used, appropriate parameterizations must be developed and used. It has not been demonstrated whether one model formulation has an advantage over the other in tackling this problem.

4. Model Simulations: Successes and Failures

4.1. Simulation of the Mean Seasonal Cycle

Until the beginning of the TOGA decade, SST received very little attention from ocean modelers. A handful of studies had presented SST fields [*e.g. Han,* 1984], but it was normally considered allowable to specify the observed air temperature and humidity in the heat flux boundary conditions. As discussed below, this approach is of limited interest if the desire is to model and understand the processes that determine SST. During the TOGA decade both intermediate models and GCMs have been used to simulate the seasonal cycle. Further, during this decade the attention has gradually moved toward the seasonal cycle of SST.

The first attempt to use an intermediate model to simulate the seasonal cycle was that of *Seager et al.* [(1988], who used a modified version of the *Zebiak and Cane* [1987] model. They paid close attention to the surface heat flux, adopting a method that rests on the observation that the marine boundary layer rapidly adjusts to the underlying SST. While this means that low-level atmospheric properties cannot justifiably be specified in the heat flux calculations, it also means that it should be possible to parameterize them in terms of the SST. SZC used a particularly simple marine boundary layer model (an assumption that the air humidity is a fixed proportion of the saturation humidity evaluated at the SST) to derive the air humidity from the SST. This allowed calculation of the latent heat flux. The solar flux was given by a formula related to observed cloud cover, and the sensible and longwave heat losses were combined into a simple term proportional to SST. Only winds and cloud cover needed to be externally specified.

The relatively simple SZC model provided a quite realistic simulation of tropical Pacific SST. Off the equator and in the western equatorial Pacific the seasonal cycle of SST was controlled by the seasonal cycle of heat flux. On the equator and in the east the seasonal cycle of SST was also influenced by equatorial upwelling and horizontal advection. SZC explained the warm eastern equatorial Pacific temperatures in northern spring and cold temperatures in northern fall as being the result of the seasonal cycle in the strength of the trade winds. These are strongest in fall, when the Intertropical Convergence Zone (ITCZ) is at its farthest point north, and weakest in spring. SZC claim this drives a seasonal cycle in thermocline depth and upwelling in the east Pacific, which in turn creates the seasonal cycle in SST. Whether it was the seasonal variation of upwelling or of the temperature of upwelled water, which mattered most was left unanswered. Either way, this coupling between the eastern Pacific SST, the zonal winds, and the ocean dynamics has recently been questioned, as will be discussed later in this section. Nonetheless, the surprising aspect of this simulation was how well the seasonal cycle could be simulated with a relatively simple dynamical ocean model and a minimum of externally specified information.

SZC considered that the bulk formula estimate of solar radiation was a potentially large source of error in their SST calculation. Indeed, *Blumenthal and Cane* [1989], using a statistical technique that located the optimal values of uncertain parameters and an assessment of the errors introduced by the data, concluded that it was not possible to demonstrate unambiguously that the model was in error. Uncertainties in the forcing data were large enough to account for model-data discrepancies.

Since that early attempt, satellite-derived estimates of the surface solar radiation have become available. *Seager and Blumenthal* [1994] used two of these data sets and a more general heat flux formulation to redo the calculations of SZC.

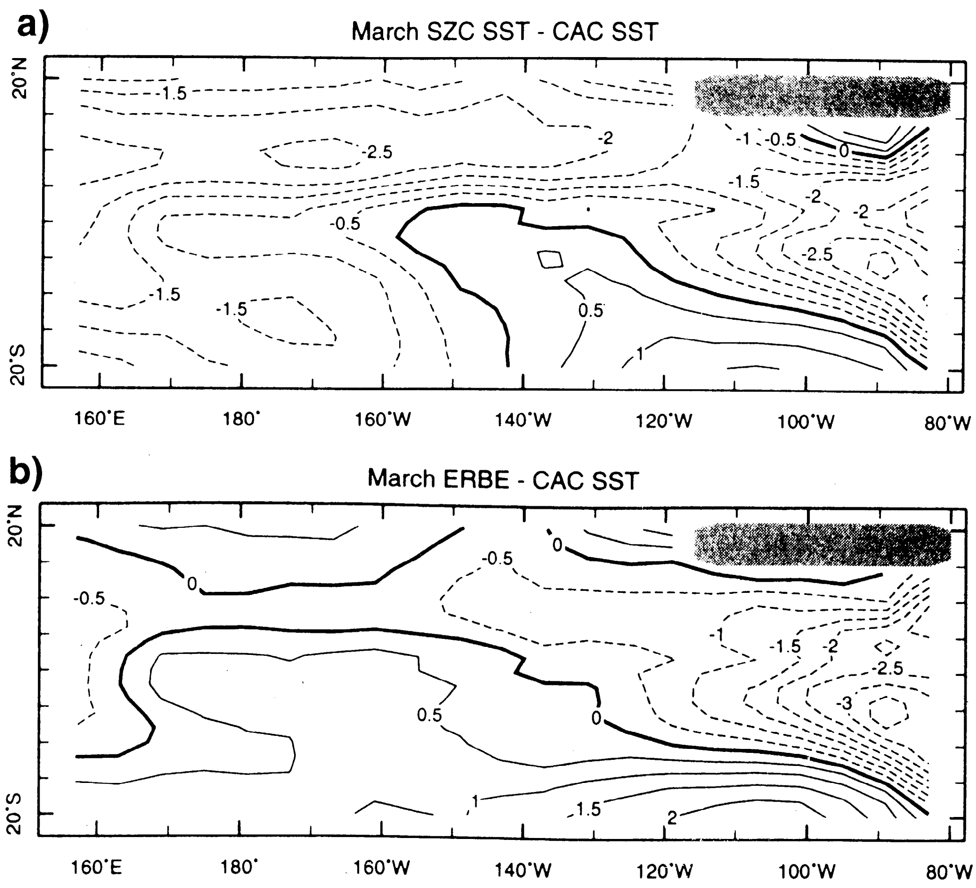


Figure 1. SST error for March from two runs of the *Seager* [1989] intermediate model. (a) surface heat flux calculated using a surface cloud climatology and (b) surface heat flux using Earth Radiation Budget Experiment (ERBE) solar radiation data. In the latter case the forcing data is considered accurate enough to conclude that the cold eastern Pacific is due to model error. From *Seager and Blumenthal* [1994]. Copyright American Meteorological Society.

They showed that these two advances led to an improved SST simulation (Figure 1). Further, using the same optimization technique as *Blumenthal and Cane* [1989], they were able to unambiguously identify a major model failing: inadequate treatment of entrainment and mixed layer processes in the east Pacific that caused errors in the seasonal cycle.

The latter conclusion was also suggested by *Chang* [1994]. He used two intermediate models: one a version of the SZC model and the other using *Kraus and Turner* [1967] mixed layer physics. While he included a term that relaxed the SST to its observed value, which prevents a complete assessment of how well these models are performing, he was able to show that the SZC model has a bias toward dynamically induced SST variability in the east Pacific. In contrast, the model with more realistic mixed layer physics responds more to variations in surface fluxes, mixed layer depth, and entrainment.

Intermediate models to simulate the Atlantic seasonal cycle have also been used by *Blumenthal and Cane* [1989] and by *Sennechael et al.* [1994]. Both studies used essentially the same model (an adaptation of the SZC model), and in both cases the simulated SST was greatly in error. The errors were sufficiently large that no amount of error in the forcing data could account for them. The model was demonstrated to be at fault, with the main problem being excessively high SSTs off the coast of Africa. Both studies thought the local equilibrium SZC heat flux formulation could be at fault. However, the Atlantic is a much smaller ocean than the Pacific, which enhances the relative role of coastal currents and upwelling, both of which are processes handled poorly by the simple linear dynamical model.

Sophisticated ocean GCMs have not necessarily performed better. The OGCM most commonly used is the GFDL tropical ocean model [*Philander and Pacanowski*, 1986]. *Philander et al.* [1987] provide a detailed description of the heat and mass budgets of the tropical Pacific version of this model. The model reproduces many aspects of the seasonal cycle of the ocean circulation including a strong South Equatorial Current and Countercurrent in northern summer and fall but an Equatorial Undercurrent that has maximum strength in spring. They identified a surge of warm water eastward in spring that occured as the southeast trades relaxed. Although they use idealized surface heat fluxes and do not show the model SST, this appears to be related to the mechanism for spring warming suggested by SZC.

A version of the GFDL model to simulate Pacific SSTs was used by *Gordon and Corry* [1991]. They used imposed surface fluxes, either from data or from an atmosphere model, together with a term relaxing the SST to observed values. They were able to identify what has become the most persistent and troublesome problem with tropical GCMs, a tendency to produce an excessively cold equatorial cold tongue. *Harrison* [1991], using the GFDL ocean model and a similar experimental setup to that by *Philander et al.* [1987] found the same problem. In Harrison's experiments, very large values of downward surface heat flux were needed to counter the cooling by vertical motion.

The issue of the equatorial cold tongue was examined in a controlled OGCM comparison instigated by the TOGA numerical experimentation group (TOGA NEG). Identical forcing and heat flux boundary conditions were specified for each model, and the ability to simulate the mean seasonal cycle of the equatorial Pacific was examined by the use of identical plots and diagnostics (Figure 2). All models showed a cold bias in the equatorial upwelling region, but the bias varied considerably between models, ranging from 0.5°C to more than 2°C in the annual mean. The smallest bias (from the LODYC model) was probably within the errors expected from the less than perfect forcing data, but most models seemed to have a definite problem. The wind stress used was *Hellerman and Rosenstein*'s [1983] monthly values multiplied by 0.75, so it is unlikely that the stresses were too strong. The report of the comparison [*Stockdale et al.*, 1993] attempted to explore some of the reasons for the differences in the model performances. One key area appeared to be resolution and the consequent specification of horizontal mixing processes. High-resolution models with low friction and a well-developed Equatorial Undercurrent gave only small cold biases, while those with lower resolution, higher viscosity, and more sluggish currents suffered most from an equatorial cold bias. Sensitivity experiments with one of the models indeed showed that horizontal viscosity was a factor in the cold bias (although recent work by *Maes et al.* [1997] modifies this conclusion; see section 3.4). Other sensitivity experiments showed that changes in vertical mixing could also have a significant impact on the cold bias problem. The overall conclusion was that a high-resolution, low-viscosity equatorial model is necessary for a good representation of the equatorial currents and that this in turn is important for the heat budget and hence SST.

One other item of interest arose from the study. This was the inclusion of SST fields from an intermediate model forced in the same way as the OGCMs. The model (derived from that of *Anderson and McCreary* [1985]) had a very different error pattern to those of the OGCMs. There was no equatorial cold bias, but there was a significant underestimate of the amplitude of the annual cycle in the eastern Pacific. This again underlies the very different behavior of OGCMs compared to simpler models and can be compared with work by *Miller et al.*[1993], where the interannual variability of different types of model was found to be different: see the discussion in the next section.

Recently, *Chen et al.* [1994a] have used a reduced-gravity, primitive equation GCM to simulate the Pacific seasonal cycle. Their model includes a quite complex representation of surface mixed layer physics. The surface heat flux scheme is that of SZC, which allows a comparison to the results of the intermediate model. The important conclusion reached by Chen et al. is that the seasonal cycle in the eastern equatorial Pacific is primarily controlled by the seasonal variation in the surface heat flux and the vertical mixing. In particular, the strong seasonal cycle of SST requires a seasonally varying mixed layer depth. For example, during northern spring the solar forcing is large and the winds are weak, both of which allow the mixed layer to shallow. In the presence of net surface heating in spring the shallow mixed layer allows the seasonal warming. A model with a fixed mixed layer depth (e.g. SZC) cannot capture this process and is liable to underestimate the strength of the seasonal cycle.

OGCMs have also been used to simulate the tropical Atlantic. *Blanke and Delecluse* [1993] present simulations of the 1982-1984 period. They chose to specify the air temperature, which disallows a direct comparison with the intermediate model simulations mentioned above. Hardly surprising their SST simulation was improved over the intermediate models, although it also possessed large errors off the African coast that were unexplained. To date, the best SST simulation of the

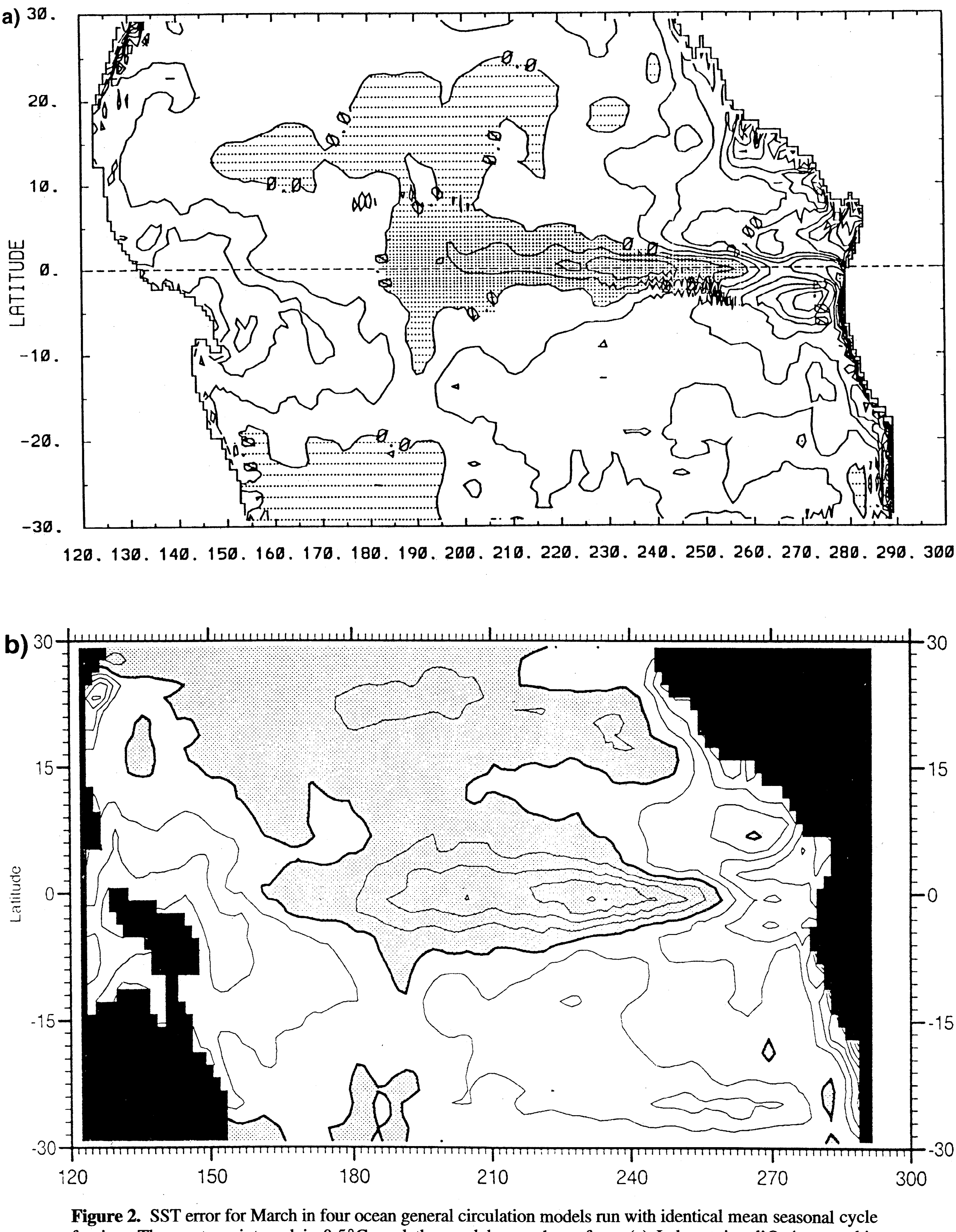
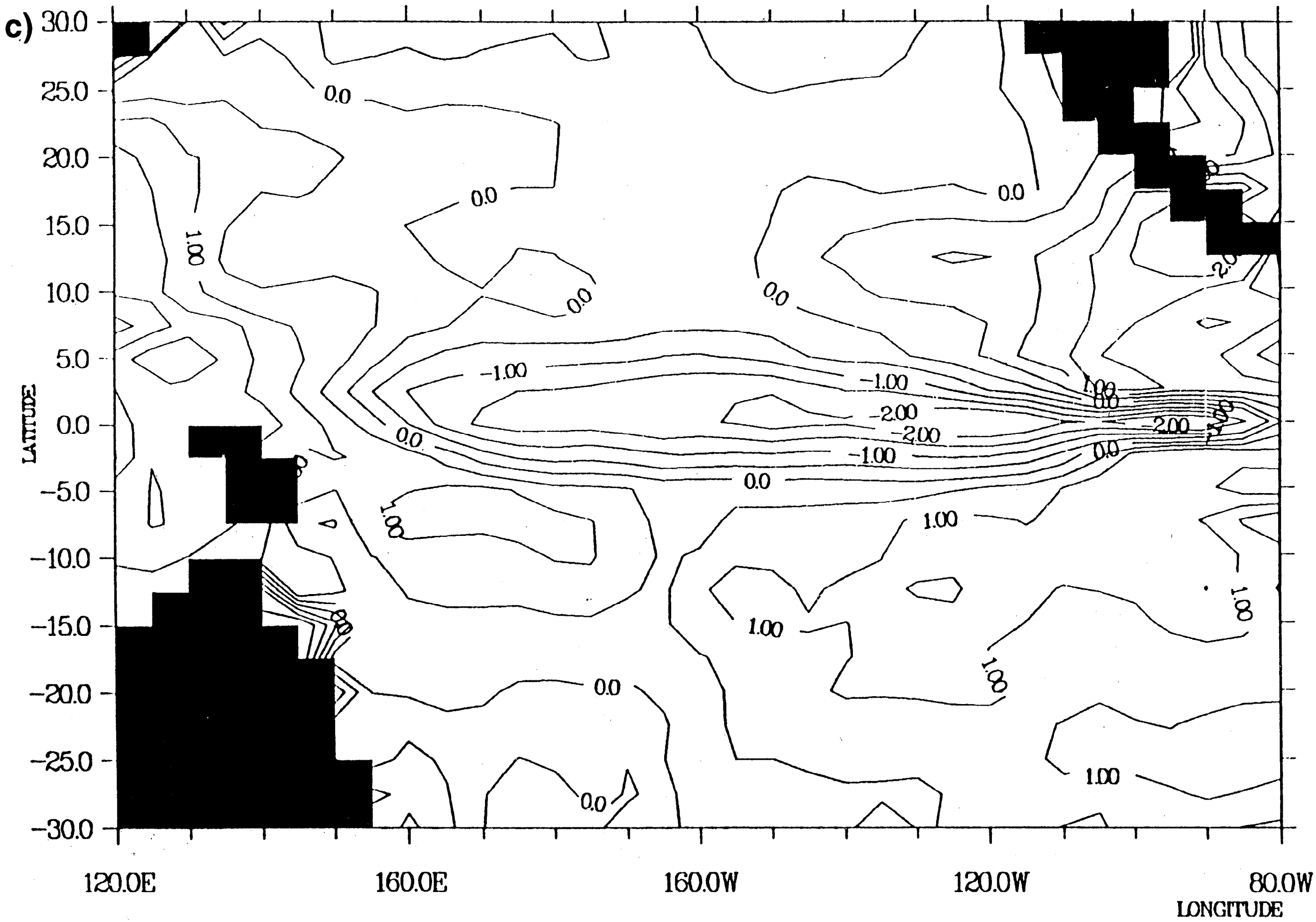


Figure 2. SST error for March in four ocean general circulation models run with identical mean seasonal cycle forcing. The contour interval is 0.5°C, and the models are those from (a) Laboratoire d'Océanographie Dynamique et de Climatologie (LODYC), (b) Oxford (*Cox* [1984] code), (c) Max-Planck-Institut fur Meteorologie and (d) U. K. Meteorological Office; from the TOGA numerical experimentation group intercomparison of OGCMs [*Stockdale et al.,* 1993]. All models have a cold bias in the central equatorial Pacific, although its magnitude varies. Copyright World Meteorological Organization.



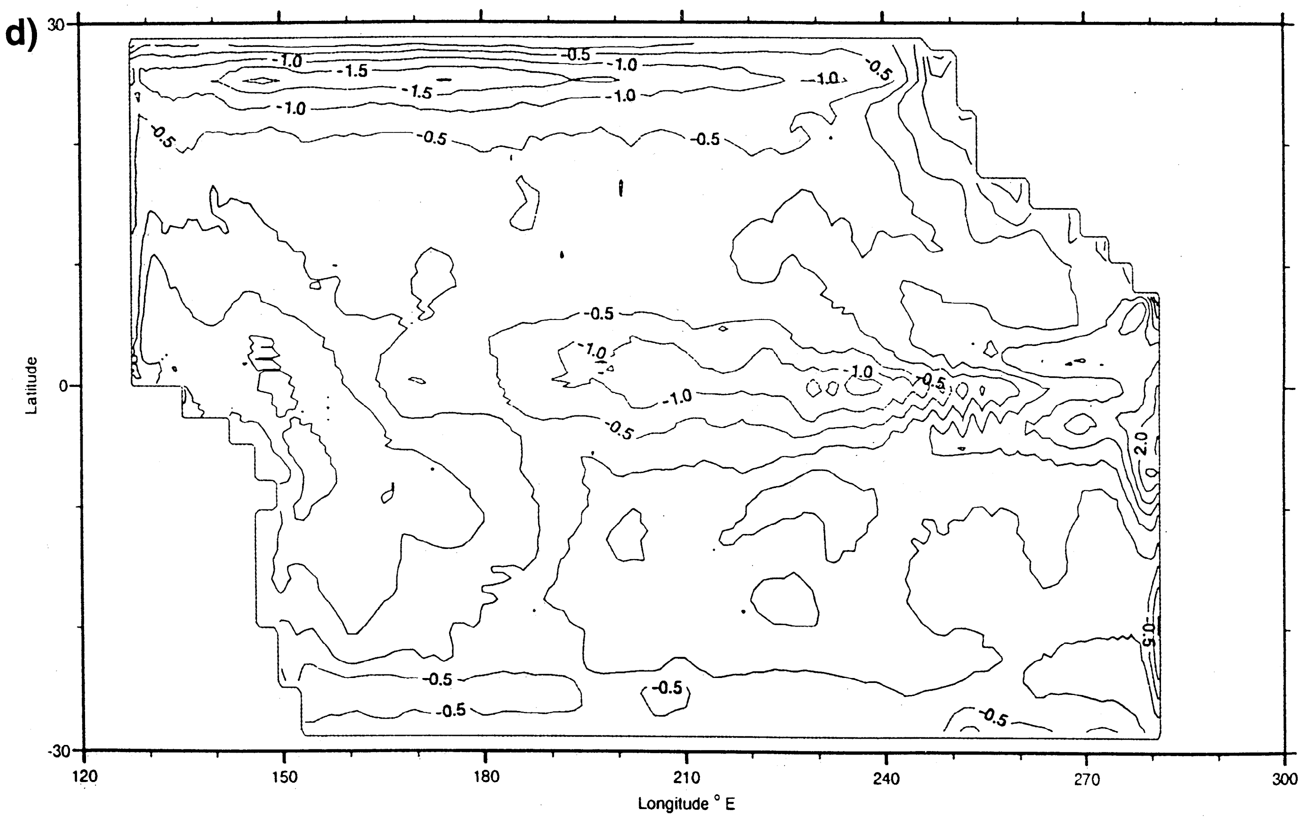


Figure 2. (Continued)

Atlantic is that by *Murtugudde et al.* [1996]. They used a version of the *Gent and Cane* [1989] OGCM coupled to the model of the advective marine boundary layer introduced by *Seager et al.* [1995a]. With this setup they were able to simulate Atlantic SST with the same degree of skill as in the Pacific. Further, when they used a local heat flux scheme instead, their SST simulation was as poor as those derived using the intermediate models. This conclusively demonstrated the role moisture advection plays in determining Atlantic SST and, in particular, the effects of advection of dry air off Africa. In contrast, the inclusion of more complex dynamics and mixing parameterizations leads to little improvement over the intermediate model simulations.

The similarity in the spatial patterns of the SST and surface winds of the tropical Atlantic and the eastern third of the Pacific is striking. This similarity does not, however, extend to the seasonal cycles of thermocline depth and the strength of the undercurrent. This has been explained by *Philander and Chao* [1991]. The difference arises because, in the Pacific, there is a seasonal cycle of zonal winds in the western basin that is out of phase with that in the east. Hence, during northern spring, weak easterlies combine with westerlies in the west to accelerate the undercurrent eastward. In contrast, in the Atlantic, there are no westerlies in the western part of the basin, and the seasonal cycle of the undercurrent strength is weak. Philander and Chao also implicate the proximity of the western boundary in the Atlantic to explain the much stronger seasonal cycle of thermocline depth in the Atlantic. They do not fully explain how this works. It is interesting that despite these differences in the dynamical response of the two oceans, the SST seasonal cycle is so similar.

As these results indicate, the TOGA decade saw considerable progress in the modeling and understanding of the tropical Pacific seasonal cycle. The emerging consensus is that the seasonal cycle of the cold tongue region is governed not so much by ocean dynamics as by vertical mixing and the surface heat flux. The seasonal variation of SST is hence determined by processes that are distinctly different to those affecting interannual variability. *Xie* [1994] has presented a simple analytical model forced by the semiannual cycle of solar radiation and an annual cycle in southerly winds. The latter is a response to the seasonal migration of the ITCZ. Without reference to ocean dynamics this model simulates an annual cycle of SST and its westward propagation. The variation of the surface flux and the turbulent mixing are responsible. Ocean dynamics, via the annual mean tilt of the thermocline, only affect the zonal variation of the phase and amplitude of the seasonal cycle of SST.

These results do leave a number of questions unanswered. *Chang*'s [1994] appeal to a role for wind stress variability in producing the eastern Pacific seasonal cycle does not identify how this mechanism operates. Does the wind stress variability primarily influence the vertical mixing as suggested by *Xie* [1994] or does it affect the SST via its influence on thermocline depth and upwelling? Is it really true that in the SZC model the seasonal cycles of thermocline depth and upwelling determine that of SST? Chang identified this kind of model as being biased toward the ocean dynamics means of altering SST, but would it perform better if mixed layer physics were represented in a more appropriate manner?

As a final comment it is worth noting that despite the widespread use of the GFDL tropical model, to date, it has not been run in tandem with the kind of heat flux boundary conditions more commonly used with the intermediate models. Hence no direct comparison of the SST simulations of the two types of models is possible. A subjective assessment is that so far the intermediate models are hard to beat. But, because the intermediate models do not need to simulate the vertical thermodynamic structure of the ocean, they are incomplete. This makes simulation of the seasonal cycle with intermediate models an easier task. It is apparently not a hindrance in ENSO studies, presumably because the timescale for maintenance of the thermocline structure is long compared to that of ENSO variability. For longer-term studies the specified vertical structure of the intermediate models severely compromises their applicability.

4.2. Interannual Variability

Ocean model studies of interannual variability during TOGA were predominantly focused on the ENSO phenomenon in the tropical Pacific Ocean basin. Although interannual simulations of variables such as sea level existed prior to the TOGA program, realistic simulations of year to year changes in SST, the most critical ocean variable for ocean-atmosphere coupling, were a major accomplishment early in TOGA. The successful simulation of interannual SST fluctuations in stand-alone ocean models was a significant prerequisite for the subsequent coupling of ocean-atmosphere models. In this respect the extreme magnitude of the 1982-1983 El Niño event was a robust test of the simulation skill for tropical ocean models. (Conversely, one could argue that this is not a very stringent test for a model and that more attention should be given to simulating more subtle El Niño events.) The qualitative and quantitative success that was demonstrated in simulating this particular El Niño encouraged the initiation of a number of multiyear hindcast experiments encompassing several ENSO events. Even though most of the focus was on ENSO-related variability in the tropical Pacific, model-based studies of the interannual variability in the tropical Atlantic and Indian Oceans were also performed. Similar to the situation with 1982-1983 El Niño in the Pacific, the 1983-1984 anomalous warming in the Atlantic and the in situ observations taken at that time during the Seasonal Response of the Equatorial Atlantic (SEQUAL)/Programme Français Océan-Climat en Atlantique Equatorial (FOCAL) experiment led to a number of modeling investigations of Atlantic Ocean variability.

One of the early attempts at modeling the 1982-1983 El Niño was the work of *Philander and Seigel* [1985]. In this Pacific Ocean hindcast study the GFDL OGCM was forced by a monthly mean 1000-mbar wind field analysis from the National Meteorological Center (NMC). This simulation was successful at reproducing several important aspects of the El Niño event such as the temporal evolution in the vertical structure of the temperature and zonal velocity fields along the equator. In response to a relaxation of the trade winds the model solution depicted anomalous eastward surface currents, a related west to east redistribution of heat, elevated SST in the eastern equatorial Pacific, and the cessation of the Equatorial Undercurrent. Analyses by *Philander and Hurlin* [1988] of the meridional transport of heat indicated the transport was northward, similar to climatological conditions, but with a noted intensification near 10°N attributed to exceptionally strong trade winds and Ekman drift. Shortcomings in the solution included a thermocline that was too diffuse and an amplitude for the SST warming that was too high in the east and never really recovered after July 1983. The quality of the wind data used in this study precluded a quantitative validation of the solution. Hence the identification of discrepancies due to deficiencies in model physics versus those in the forcing was not pursued.

A more rigorous assessment of this solution was performed in a series of papers by Harrison and collaborators. *Harrison et al.* [1989] addressed the problem of uncertainty in the wind forcing by repeating the experiment of *Philander and Seigel* [1985] with five different wind products and assessed the solutions using XBT observations along three major ship transects. These comparisons indicated that the quality of the solutions was highest within the equatorial waveguide, most notably for dynamic height, and to a lesser extent for SST as root-mean-square (rms) errors could be as high as 2-3°C. Away from the equator, the quality of the solutions broke down, especially in the region of the North Equatorial Countercurrent (NECC), and pointed to problems with the wind stress curl forcing. *Harrison et al.* [1990] analyzed the equatorial heat budgets of these solutions. Anomalous eastward advection was identified to be a major source of warming in all the solutions. However, the role of vertical advection, meridional advection, and local versus remote forcing effects varied from one forcing function to the other. Against this backdrop of problems with available wind data, *Harrison* [1989] conducted an OSSE to demonstrate that most of the simulated 1982-1983 variability in heat content, zonal velocity, and SST along the equator could be recovered when the model was forced with a subset of the full wind forcing restricted to winds only within ± 7° of the equator. These findings had important implications for the latitudinal extent of the full deployment of the TOGA TAO array. Heat flux sensitivity studies by *Harrison* [1991] illustrated the need for very accurate heat flux information in regions of light winds and high SST such as in the western Pacific. Using a similar model, *Rosati and Miyakoda* [1988] highlighted the improvements that could be obtained in SST by using 12 - hour forcing rather than monthly mean forcing.

The success in simulating the 1982-1983 El Niño with an OGCM led to hindcast experiments of interannual SST variability in several different models. *Latif* [1987] forced a level OGCM with more than two decades of Florida State University (FSU) winds and a Newtonian damping condition for the heat flux. *Seager* [1989)] forced the ocean component of the Lamont coupled model [*Zebiak and Cane*, 1987], i.e., one with linear dynamics in a single layer and fully nonlinear thermodynamics, with the FSU winds for 1970 to 1987 and parameterized the heat fluxes in a manner that did not require air temperature or humidity to be specified [*Seager et al.*, 1988]. Figure 3 shows the result, which in general terms typifies the performance of most interannual SST simulations. *Miller et al.* [1992] examined the simulation skill of an isopycnal model (OPYC) with an embedded bulk mixed layer model forced by the FSU winds for 1970-1985 and bulk formula heat fluxes modified with a Newtonian damping. *Brady* [1994] considered the interannual variability in a sigma coordinate model with a constant mixed layer depth and forced by the FSU winds for 1971 to 1990 with heat fluxes parameterized with the *Seager et al.* [1988] scheme. All of the models demonstrated reasonable performance at simulating the low-frequency equatorial variability in variables such as SST, zonal current, and thermocline depth fluctuations. Reasons for significant differences in the quality between solutions, *e.g.*, problems with SST off the equator, a too diffusive thermocline,

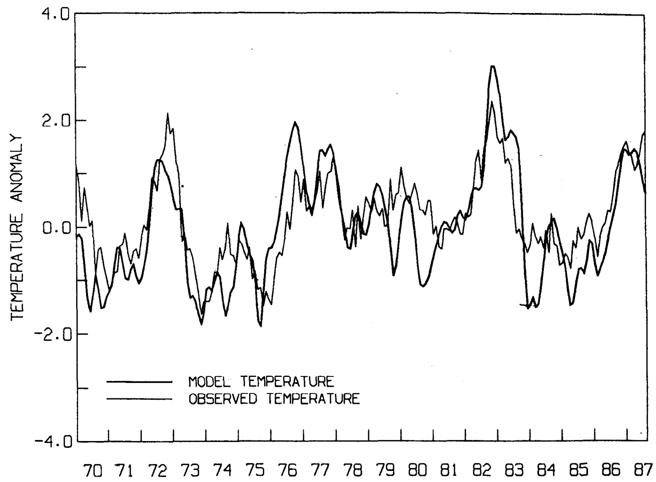


Figure 3. SST anomalies in the central equatorial Pacific modeled by *Seager* [1989]. Thin line is observed anomaly; thicker line is model simulation. Copyright American Meteorological Society.

or the recovery from El Niño events, are hard to identify because of differences in the model parameterizations and forcing functions. In an attempt at reducing some of these differences, *Miller et al.* [1993] performed a controlled set of experiments by forcing the GFDL model, the Max-Planck-Institute z-coordinate model (MPIZ) used by *Latif* [1987], the Lamont model, and the OPYC model with the same FSU winds for 1970-1985 and Newtonian damped heat fluxes. The three GCMs (GFDL, MPIZ, OPYC) exhibited maximum simulation skill for anomalous SST variability in the central equatorial Pacific, whereas the Lamont model performed best near the equatorial eastern boundary. The physical mechanisms responsible for SST changes were different from model to model. In the Lamont model, vertical temperature advection was the predominant factor influencing large-scale SST. For the MPIZ model, meridional advection was most important for governing changes in SST. In the OPYC model, zonal, meridional, and vertical advection were all important. These intercomparisons illustrated that although the low-frequency behavior of SST was being simulated, the physics inducing this variability varied from model to model. Since all of these models were being tested for their suitability for coupling with an atmospheric model, potentially different coupled behavior might be expected as a result of the differences in the dominant ocean physics.

Just as the large SST deviations for the 1982-1983 El Niño motivated a number of primitive equation model experiments, the large changes in basin-wide dynamic topography also gave rise to a number of reduced-gravity model studies. *Pazan et al.* [1986] used a reduced-gravity simulation of the 1982-1983 El Niño to supplement dynamic height observations in order to monitor the propagation of off-equatorial Rossby waves. This off-equatorial wave activity was noted to induce positive anomalous heat content in the tropical western Pacific that preceded the equatorial sea level changes by 1 to 2 years. In a follow-on study by *Inoue et al.* [1987] the validation of this reduced-gravity simulation of the 1982-1983 El Niño was extended beyond point comparisons with tide gauge data to include dynamic height fields from XBT measurements. A high degree of fidelity between model and observations was noted for the basin-scale and low-frequency mass changes associated with El Niño. The pattern correlation was as high as 0.9. However, on smaller spatial and temporal scales, the comparisons with the observations were not so good. *Kubota and O'Brien* [1988)] calculated the spatial and temporal decorrelation scales of interannual variability in this type of model using a hindcast for 1965-1984. The temporal decorrelation for the model height field was between 9 and 12 months. The zonal decorrelation scales were found to be a function of latitude and frequency, being largest at low frequencies and low latitudes, of the order of 20° - 30° of longitude. Meridional decorrelation scales were ~ 4° in latitude.

Many of the other applications of reduced-gravity models extended beyond 1982-1983 case studies and concentrated on the year to year changes in oceanic heat content and meridional heat transport over several decades. *Pares-Sierra et al.* [1985] used a solution from 1962-1979 to study the interannual heat transport within the tropical Pacific basin. The cross-equatorial heat transport during El Niño events was depicted as an enhancement of the annual cycle of Ekman transport which did not contribute to changes in equatorial heat storage. Similar results were subsequently found in the OGCM analyses of *Philander and Hurlin* [1988]. *Zebiak* [1989] examined the equatorial heat storage for a composite of the 1972, 1976, 1982, and 1986 El Niño events. Equatorial Rossby wave processes, in contrast to the implication of off-equatorial Rossby waves in work by *Pazan et al.* [1986], were found to make significant contributions to the equatorially convergent heat transport. Consistent with the hypothesis of *Wyrtki* [1985], heat content was found to increase near the equator in the year prior to the composite El Niño year. However, *Springer et al.* [1990], in their model-based analyses of the 1982-1983 heat content, did not find any clear evidence of a buildup prior to this particular event and suggested this was just another one of the unusual aspects of this El Niño in addition to the timing and intensity.

Simple models have also been used to isolate the role of equatorial and coastal wave processes on ENSO timescales. *Wakata and Sarachik* [1991] used a reduced-gravity model forced by 28 years of FSU winds to examine the role of equatorially trapped waves in ENSO mechanisms such as the delayed action oscillator. The evolution of Kelvin and Rossby wave behavior in the wind-forced model was found to be consistent for both the warm and cold phases of ENSO, with the coupled instability mechanism proposed by *Suarez and Schopf* [1988], *Battisti* [1988], and *Battisti and Hirst* [1989]. *Pares-Sierra and O'Brien* [1989] and *Johnson and O'Brien* [1990] investigated the possibility of an interannual oceanic teleconnection between the equatorial Pacific and coastal changes at higher latitudes along North America. Reduced-gravity model simulations demonstrated that a significant amount of the interannual variability of sea level within the California Current region and north along the coast to 50°N could be explained by the poleward propagation of coastally trapped Kelvin waves excited in the equatorial Pacific.

In the tropical Atlantic Ocean, interannual variability is often considered to be secondary in importance to the influence of the seasonal cycle. Nonetheless, year to year and decadal fluctuations within this ocean basin are not negligible. A prime example is the time period 1983-1984 when the extensive United States and France oceanic field program, SEQUAL/FOCAL, was taking place in the tropical Atlantic [*Weisberg*, 1984]. Although this program was designed to monitor the forcing and response of a full seasonal cycle, there were, in fact, major disruptions to the seasonal cycle during the experiment. During 1984, there were major perturbations of SST, wind, convection, and ocean circulation that had some resemblance to an El Niño event in the Pacific [*Philander*, 1986; *Lamb et al.*, 1986; *Weisberg and Colin*, 1986; *Hisard et al.*, 1986; *Katz et al.*, 1986; *Horel et al.*, 1986]. The data collected during SEQUAL/FOCAL and the anomalous nature of the variability stimulated modeling studies of the interannual variability in the tropical Atlantic. For example, reduced-gravity models were used by *Reverdin and du Penhoat* [1987], *Weisberg and Tang* [1987], and *du Penhoat and Gouriou* [1987] to analyze the forced response of dynamic height in terms of the superposition of equatorial Kelvin and Rossby waves. Comparisons with the in situ observations suggested that the large-scale response of these solutions was consistent with that observed. *Braconnot and Frankignoul* [1993, 1994] and *Frankignoul et al.* [1995] exploited SEQUAL/FOCAL to evaluate several different model simulations of this event, as well as to discriminate between model deficiencies, forcing uncertainties, and data uncertainties; see *Frankignoul et al.* [1989] for the methodology used. One of the conclusions was that despite all the limitations in the forcing, the model simulations of the thermocline depth were a long way from being compatible with the observed data. Nonetheless, it was possible to distinguish between the (in)ability of the various models (ranging from a two-mode linear model to an OGCM) to fit the data.

Decade-long time series of in situ observations such as sea level are not as prevalent in the Atlantic or Indian Oceans as they are in the Pacific. Hence there has been less validation data and motivation for multiyear simulations of interannual variability. However, long time series of SST are available, and recently, an increasing amount of attention has been devoted to modeling the year to year changes in Atlantic SST. *Carton and Huang* [1994], *Servain et al.* [1994], and *Huang et al.* [1995] have used OGCMs to study the interannual fluctuations in SST during the 1980s. Qualitative agreement was found between some of the larger modeled and observed SST anomalies along the equator such as in 1984 and 1988. Relaxation of the trade winds in the western equatorial Atlantic was found to induce a deeper model thermocline in the eastern equatorial Atlantic and thereby reduce the vertical heat flux from the upper layers leading to higher model SSTs during these episodes. Quantitatively, however, discrepancies with the observations could be as large as 1-3°C.

With the advent of nearly continuous satellite altimeter observations, simulation studies of SST are being supplemented with modeling studies of surface topography in the tropical Atlantic and Indian Oceans. Similar to what was encountered for the Pacific, wind-forced simulations of model sea level were found to be highly correlated on low frequencies and large scales with the first several years of Geosat altimeter observations for the tropical Atlantic [*Arnault et al.*, 1992] and Indian Ocean [*Perigaud and Delecluse*, 1992, 1993; *Anderson and Carrington*, 1993]. Yet, on smaller spatial and temporal scales, significant differences were present.

4.3. Model Simulations of Higher-Frequency Variability

There is energetic upper ocean thermal and current variability across a very wide range of timescales in the tropical Pacific. Most of this variability is in direct response to atmospheric variability, but the tropical instability waves are the result of a hydrodynamic instability of the near-equatorial eastern Pacific oceanic circulation. These two types of phenomena deserve separate discussion.

Tropical instability waves (TIWs) are a prominant aspect of the central tropical Pacific during most months of the year; their amplitude typically is weakest between March and May in a "normal" year, and often they are not present during ENSO events when the easterly trade winds have weakened. Our understanding of the nature and effects of TIWs has improved during the TOGA period [see *Proehl*, 1996; *McCreary and Yu*, 1992], although that is not the focus of the discussion here.

Many primitive equation ocean models feature TIWs as intrinsic aspects of the model flows. As an example, *Philander et al.* [1986] present an analysis of the waves as seen in their Pacific and Atlantic simulations. However, model TIW amplitudes are in general very sensitive to the level of model subgrid scale mixing of heat and momentum adopted. Model results indicate that the net meridional transport of heat and momentum from TIWs can be substantial when the mixing is chosen to produce model TIW amplitudes comparable to those observed. While the model TIWs can be made qualitatively consistent with observations, because of the model TIW sensitivity we are unlikely to be able to assess the representativeness of model TIW results until measurements of the TIW variances and covariances in the ocean have been made and model-data comparisons can be carried out. This has not been a priority activity during TOGA but may be an important model validation issue in coming years. Because current and temperature variability in TIWs are greater than their seasonal and interannual variability, observing strategies must be such as to limit aliasing of the TIW energy into the averages of interest.

Atmospheric forcing results in variability from minutes to the monthly and longer scales of most interest to TOGA. Diurnal processes in the mixed layer and upper ocean can be vigorous, especially when the wind stress is light. Further, much of the variance in the large-scale winds lies in periods shorter than 30 days. Unlike the midlatitude ocean, the tropical ocean is capable of basin-scale response to forcing events that last for as little as a few days. Forcing on the few to 10-day scale occurs strongly during tropical cyclones, particularly when these form pairs that straddle the equator, and during what have come to be called "westerly wind events". The physical processes responsible for westerly wind events (WWEs) remain incompletely understood, as is the range of oceanic responses to the different sorts of WWEs. The TOGA Coupled Ocean-Atmosphere Response Experiment (COARE) field program was carried out in part to try to improve our understanding of a few such events and the oceanic response to them. There is weaker forcing on the 30-60-day timescale by atmospheric variability associated with the Madden Julian Oscillation; the manifestations of this forcing are clearer in thermocline and subthermocline variability than they are in surface and near-surface changes. Only in the past few years have useful high-frequency wind fields become available for simulations of the high-frequency tropical Pacific variability; the early studies of response to WWEs made use of idealized wind patterns [e.g. *Harrison and Giese*, 1991; *Giese and Harrison*, 1991]. These revealed that substantial central and eastern equatorial Pacific response can result from those WWEs that have substantial near-equatorial zonal wind variability. They also illustrated the importance of higher vertical mode Kelvin motions in the eastern Pacific response; although the sea level response to WWEs is dominated by the first baroclinic mode, the second and third modes are strongly forced by a typical WWE,and these typically have a larger effect on the model SST and near-surface currents than does the first-mode response. Because the higher vertical modes propagate eastward much more slowly than the first mode, current and temperature changes continue to occur in the east months after a strong WWE. However, the detailed response to any particular event is strongly a function of the structure of the equatorial waveguide east of the WWE because of the scattering effects of the eastward shallowing thermocline as well as wave-mean flow interaction in the complex equatorial circulation.

Under ideal circumstances, as few as three strong WWEs properly timed can account for the patterns and amplitude of SST warming described by *Rasmussen and Carpenter* [1982] in their composite ENSO event. This suggests that WWEs may play a significant role in the initiation and warming phase of ENSO. However, not every equatorial WWE produces the warming expected from idealized studies; clearly, additional work is needed to understand better the various factors that control the SST changes subsequent to WWEs. Such work is now being carried out, using tropical wind fields constrained by the TAO wind data; hopefully, these analyses will permit detailed hindcasts of particular WWEs to be carried out successfully and detailed understanding of the upper ocean current and temperature changes to be gained.

The effects of 30-60-day atmospheric forcing have not been explored in the context of straight ocean model studies. Although forcing energy definitely exists in this band in surface

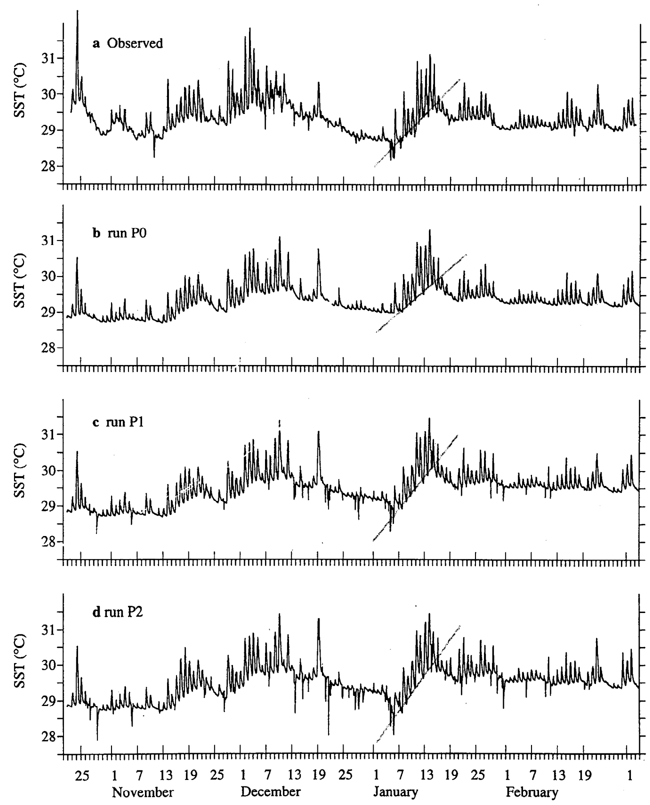


Figure 4. SST at a buoy during TOGA COARE, observed and modeled with a one-dimensional mixed layer model. Experiments P0, P1, and P2 use 0, 1, and 2 times the observed values of precipitation respectively, which is seen to influence the SST evolution. From *Anderson et al.* [1996]. Copyright American Meteorological Society.

wind records and in ocean mooring records, the level of forcing energy is small compared to that in the diurnal and WWE period ranges. Initial value studies with coupled ocean-atmosphere models in this period range have been done but will be discussed in another contribution to this volume.

Diurnal response is very importantly affected by the wind stress (actually by "ustar cubed", the 3/2 power of the magnitude of the wind stress), by the incident short wave radiation, by the sum of evaporation minus precipitation, and by the net air-sea heat flux. Typically, it is not possible to know these various forcing quantities with sufficient accuracy to carry out realistic simulations of diurnal variability. However, one-dimensional studies have been done, and they illustrate the magnitude of the upper ocean thermal changes that can occur through these forcing processes [*Schudlich and Price*, 1992; *Anderson et al.*, 1996]. An illustration of the variability of SST on diurnal timescales and the ability of one dimensional models to simulate it when forced with in-situ data is given in Figure 4. Such results have not yet been demonstrated with more general ocean models. Improved atmospheric planetary boundary layer processes in the weather center surface tropical analyses will, it is hoped, make ocean circulation model studies of processes on these timescales feasible before long. The incorporation of improved oceanic mixed layer representations in ocean models has been explored but has not proven physically useful because of the shortcomings of the available forcing information. Again, the TOGA COARE program results offer many opportunities for model studies in this timescale and will lead to improved forcing information and information about model behavior.

5. Studies Using Ocean Models

5.1. Role of the Ocean in ENSO

So far as we are able to determine at this time, ENSO is truely a coupled ocean-atmosphere phenomenon. A recent review of ideas about coupled mechanisms for ENSO is given by *Battisti and Sarachik* [1995], and this topic is addressed elsewhere in this volume. Here we briefly review some particular aspects of the ocean that pertain to its role in ENSO.

Foremost, the ocean provides a huge reservoir of heat, which is available to alter SST through a variety of processes. To the best of the authors' knowledge and when attention is restricted to the equatorial waveguide, atmospheric net surface heat flux has always been found to respond to changes in large-scale ocean SST rather than to cause the SST changes associated with ENSO events. Thus changes in oceanic processes (which may be driven by wind stress or, much less importantly, net freshwater flux) lead to large-scale SST changes, and the atmosphere typically responds in a way that will produce a net surface heat flux which opposes the SST change. No description of the particulars of the atmospheric heat flux response (short wave, long wave, sensible, and latent heat flux components) in different regions of the tropical Pacific during the different phases of ENSO is available, although coupled models will soon be analyzed for this response.

We do know that the observed sequence of surface wind and SST changes varies considerably from ENSO event to event and within events. In some cases, wind changes in the western and/or central equatorial Pacific precede the appearance of equatorial SST anomalies; in others, wind and SST changes occur simultaneously (to within the time resolution of our historical observations). The simple relationship between SST anomalies and surface wind anomalies offered in the *Gill* [1980] model greatly oversimplifies the situation as we have been able to observe it. Another oversimplification, which has been widely used, is that SST anomalies are inversely proportional to ocean thermocline depth anomalies. In some regions of the ocean and during some phases of some ENSO events this behavior is observed, but it is not generally a correct view of the oceanic relationship between these two variables.

For better or worse, the oceanographer views the oceanic changes during ENSO as the response to atmospheric forcing, and considerable success in modeling the evolution of the ocean during recent ENSO periods has been possible by imposing the winds and very simple parameterizations of the net surface fluxes. From these studies it is clear that SST changes tend to occur in different regions and different phases of ENSO for different reasons. Changes in the wind stress in the western and central Pacific can cause substantial SST warming farther east along the equatorial waveguide; this type of behavior is most common in the initial warming phase of ENSO. Once the local zonal wind stress begins to reduce in the central equatorial Pacific during the main phase of ENSO, warm SST can be maintained through a combination of reduced upwelling of cold subsurface water, of upwelling of anomalously warm subsurface water, of reduced meridional advection of heat (because of weakened Ekman transport), and of enhanced zonal advection. The detailed balances that pertain depend on the region of interest, the details of the wind forcing, and the ENSO event of interest, according to recent model studies of oceanic heat budgets. The termination of ENSO has occurred because of the return toward normal conditions of the prevailing easterly wind stress in every model hindcast known to the authors. Thus, given the time and space evolution of the wind stress, we can reproduce the major aspects of SST change using just an ocean model. Thus ENSO for the ocean is basically in the winds; determining the winds is beyond the ability of ocean physics alone.

The ocean provides means for the atmosphere-ocean system to develop long-period coupled modes of behavior, because oceanic processes typically change SST along the equator on timescales of months. Even when the ocean is forced by high-frequency atmospheric events, the ensuing SST changes take months for completion. However, it seems very unlikely that we shall understand the fundamental mechanisms of ENSO by viewing the system from the perspective of either fluid; ENSO appears to be a truly coupled ocean-atmosphere process.

5.2. Surface Fluxes

Solar radiation and wind stress are the only sources of energy for the ocean. Over most of the ocean the SST is determined through a balance of the surface fluxes of heat and vertical mixing in the ocean mixed layer. It is only in limited regions of the ocean that dynamical processes (horizontal and vertical advection and eddy fluxes) substantially affect the SST. Those areas, however, include the equatorial cold tongues, which are of great importance in determining the tropical climate and its variability. Even within these regions the processes governing the seasonal variability and the interannual variability are emerging to be quite different. For the seasonal cycle it is the surface fluxes and ocean mixing processes that primarily govern the SST while, for interannual variability, ocean dynamical processes involving thermocline adjustments and advection are dominant. The annual mean state itself is, however, established through a coupled interaction involving both dynamical processes and surface fluxes. In terms of the surface fluxes it follows that in studies of interannual variability, attention must focus on the fluxes of momentum while, for the seasonal cycle, the surface fluxes of heat are of equal or greater importance.

Interannual variability of the tropical Pacific Ocean is largely driven by variability of the zonal wind stress along the equator. Longitudinal shifts in the location of atmospheric deep convection force anomalous easterly and westerly wind stresses, which force ocean motions. The ocean response can alter the distribution of SST, which, in turn, influences the atmospheric circulation. At the start of the TOGA decade, even the zonal stress was poorly known. Ship-based observations were the primary data source. *Harrison et al.* [1989] showed that different compilations of wind stress data, when used to force an ocean GCM, could lead to simulations of ocean variability that were distinctly different. As pointed out by *Miller and Cane* [1996], prior to 1990, the wind stress data were of such poor quality that it was extremely difficult to identify model failings or to demonstrate that one model was better than another. Such an unsatisfactory situation, which made progress in ocean modeling almost impossible, led to the deployment of the TOGA TAO array of moored buoys that now straddle the full length of the equatorial waveguide [*Hayes et al.*, 1991]. These observations have significantly improved the quality of the wind stress data. However, the array covers only a fraction of the tropical Pacific. Ultimately, wind stress data sets of satisfactory accuracy and coverage will only come from satellite-based estimates. *Busalacchi et al.* [1993] have analyzed the data derived from the special sensor microwave imager (SSM/I) [*Atlas et al.*, 1991] and compared it with surface and model-produced data sets. Their conclusion is that at some time in the future, reliable retrieval of wind stresses from space will be a reality. If this is so, then it will constitute a major advance, one that will considerably reduce the uncertainties associated with simulations of the tropical oceans and one that should lead to more rapid progress in identifying and remedying model errors.

Given the paucity of our knowledge of surface heat flux variations over the oceans, it is fortunate that ENSO SST variability does not depend critically on the surface heat fluxes. The Lamont forecast model assumes that the anomalous surface heat flux acts to damp the SST on a fixed timescale, a simple approach that is evidently not a major source of error, given the relative success of the simulations. Further, this modeling of heat flux anomalies as a linear damping appears to be a reasonable approximation according to *Barnett et al.* [1991], who analyzed results from an atmospheric GCM, and according to *Seager et al.* [1995b], who used a simple model of the marine boundary layer.

The full story of the heat flux response to SST anomalies is complex. Bulk formula estimates indicate that during a warm event in the central Pacific the latent heat flux is reduced because of the weak winds, and this will act to reinforce the SST warming. *Seager* [1989] found this to be of some importance in hindcasts of the 1982-1983 and 1986-1987 ENSO events. On the other hand, solar radiation is also reduced in such a situation because of the increase in convective clouds. *Waliser et al.* [1994] examined the role of the solar forcing in SST simulations while holding the wind speeds fixed. Their ambiguous results suggest that it is quite unwise to allow only one component of these large and cancelling fluxes to vary. The net effect of both changes may be small. Reliable, quantitative studies of the combined effects of heat flux variability on tropical SST have not been made, however, and we do not know (for instance) how the different character of different El Niño events might be related to the surface heat flux.

One aspect of ENSO variability in which surface heat fluxes may be implicated involves the off-equatorial variations of SST. Ocean models typically produce SST anomalies that are too tightly confined to the equator, whereas the observed anomalies extend poleward some 20° of latitude. Ocean dynamics or horizontal mixing appears incapable of creating this spreading. Furthermore, away from the coasts, ocean dynamics contribute little to the SST variability in these regions. This suggests that the surface heat budget, perhaps through changes in low cloud cover [e.g. *Klein and Hartmann*, 1993] or evaporation, may be responsible. This observed aspect is still awaiting an explanation and is yet to be reproduced by a model.

For the case of the seasonal cycle, errors in the wind forcing assume less significance while those associated with the surface heat fluxes become dominant. In every region of the tropical oceans the surface heat fluxes are of first-order importance in establishing the seasonal cycle. This is true even in those regions where ocean dynamics also play a role (i.e., the cold tongues and areas of coastal upwelling). The actual heat flux at the ocean surface is not known and, in fact, is not directly measurable. The estimates that do exist are evaluated with bulk formulae that are calibrated against some turbulence measurements that were taken in some distant location. *Blanc* [1987] has performed an analysis of the errors involved in bulk formula estimation of surface fluxes. These stem from the accuracy of the turbulence measurements used in the calibration, the choice of exchange coefficient in the bulk formula, the accuracy of the measurements of air properties used in the formula, and the distortion of those properties by the ship. Total errors of 50% or so are typical. While in some areas of the ocean (e.g., the TOGA COARE region) we are likely to shortly have more accurate assessments of the surface fluxes, over the majority of the oceans the fluxes will remain essentially unknown.

A good rule of thumb is that 50 W m^{-2} of heat will warm 50m of water by one degree in 50 days. Fifty W m^{-2} is, however, not an unusual level of error in surface flux data sets. Obviously, imposing such erroneous data on an ocean model will lead to enormous errors in SST. Normally, the fluxes are not imposed absolutely but with a negative feedback term. Alternatively, the fluxes can be modeled using bulk formula estimates. In either case the uncertainties introduced are so large that it is often not possible to identify model failings. While it is possible to arm wave about this problem [e.g., *Seager et al.*, 1988], it is also possible to deal with it quantitatively. Since the fluxes and their evaluation are uncertain, it is allowable to vary the uncertain parameters in the heat flux calculation within an accepted range. For example, exchange coefficients that have been used to evaluate the latent heat flux vary by 30% or more in the ocean modeling literature. Using a statistical optimization technique, *Blumenthal and Cane* [1989] found the set of uncertain parameters that gave the best SST simulation possible within the constraints of a combined ocean-surface heat flux model. If this best possible SST simulation still differed from the observed SST by more than could be accounted for by errors in the forcing data (e.g., winds and solar radiation), then the model is unambiguously indicated to be in error. If the errors are small enough that they can be accounted for by errors in the forcing data, then it is not possible to prove that the model is wrong. This technique is quite powerful and has been used to assess simulations of tropical SSTs with intermediate models by *Blumenthal and Cane* [1989], *Seager and Blumenthal* [199]), and *Sennechael et al.* [1994]. This work has shown that in the Pacific the model treats entrainment and mixing processes inadequately and that in the Atlantic a local heat flux boundary condition [*Seager et al.*, 1988] leads to enormous errors in SST. So far this technique has not been applied to a GCM, although it could be.

These considerations highlight how important it is, in formulating heat flux boundary conditions, to model the feedback between the fluxes and the SST. If this is done correctly, then the model at least stands a chance of producing an acceptable SST while determining its own surface flux. Both the SST and the flux can be used as verification while remembering that the latter is a very weak constraint because its true value is so poorly known. The flux-SST feedback can be faked by including a term in the boundary condition that relaxes SST to values derived from observations [*Haney*, 1971] accounted for implicitly by imposing atmospheric quantities (which assumes that the atmosphere has an infinite heat and moisture capacity, e.g., *McCreary and Kundu* [1989], *Miller et al.* [1992], and *Giese and Cayan* 1993] or through the use of a model of the marine boundary layer that allows the atmosphere to adjust to the underlying SST [e.g., *Seager et al.*, 1995a].

An appropriate treatment of surface heat flux is also important in the testing of any ocean model prior to its coupling to an atmosphere GCM. Coupled models of the tropical Pacific are plagued by problems of double ITCZs, overdeveloped cold tongues, and warm SSTs in the southeast Pacific [*Mechoso et al.*, 1995]. These problems of climate drift emerge through coupled interactions that, in part, involve the surface heat fluxes. If the uncoupled ocean model is too tightly constrained to observations (either through a strong feedback term or through the use of imposed atmospheric conditions in bulk formula calculations), then errors which will become significant in coupled mode may easily be overlooked.

When choosing a heat flux boundary condition for an ocean modeling study, thought must be given as to the purpose and interpretation of the experiment. Perhaps the safest procedure is to calculate the surface flux from appropriate bulk formulae, using, as input, data sets such as wind and cloudiness (and/or surface solar radiation). This should provide a realistic flux-SST feedback, allowing any errors or anomalies in the simulation to develop appropriately. On the other hand, it can be a lot of work to construct a good surface flux model, and simplifications can cause problems in the mean flux in particular. An alternative approach is to specify the heat flux appropriate to the observed SST and provide a model of how the flux varies with temperature. In the case of such an "anomaly" model of heat flux a simple linear feedback will generally be adequate, and indeed, the use of a constant value of about 10-15 $Wm^{-2}/°C$ throughout the tropics is a reasonable approximation to the observed variation of heat flux with temperature. A stronger feedback can be applied if a more realistic simulation is desired and errors in the original specified heat flux are thought to be significant, as will often be the case. This heat flux model can be applied either using the mean seasonal cycle as a base, in which case any interannual SST variability will be damped to climatology and it is important that an appropriately weak coefficient is used, since it is used to simulate real heat flux variability, or it can be applied using specified interannually varying heat flux as a base, in which case the relaxation should be to the interannually varying SST and the strength of relaxation can be chosen freely since it is compensating only for errors in the simulation. The use of this linear feedback technique is simple to implement and allows use to be made of the best available heat flux estimates. It will generally give less insight into surface flux processes and a correspondingly greater risk of misapplication than the full modeling approach.

5.3. Data Assimilation and Forecast Initialization

Ocean models can be used to examine how the ocean works. Good models can also be coupled and used both for looking at how the coupled system behaves and for making specific predictions of future behavior. But another important aspect of ocean modeling is work which seeks to combine models with data, a process known as data assimilation. This is important for testing models and forcing using models to extract maximum information from expensive data, and initializing coupled forecast models. Ocean data assimilation is an extensive subject, which could easily fill a paper of its own. Here we simply give a selective view of some of the most relevant aspects, the work done on methods of data assimilation and the use of data assimilation in model initialization and model testing and development. A more thorough review of tropical ocean data assimilation can be found in work by *Busalacchi* [1996], with related ocean assimilation topics covered in companion papers in the same volume. Another recent review of oceanographic data assimilation is by *Anderson et al.* [1996], and a good general discussion of assimilation can be found in work by *Ghil and Malanotte-Rizzoli* [1991]. A large collection of papers on the use of satellite altimetry, not much discussed here, can be found in a recent Journal of Geophysical Research special section [*Cheney*, 1995].

5.3.1. Data assimilation methods. Tropical ocean data assimilation is a field which has been pioneered during the TOGA period. Many studies have sought to clarify the adjustment of models to various types or combinations of data and to clarify the ability of assimilation schemes to constrain a model towards observations. Early papers used relatively simple assimilation techniques together with idealized experiments. In these an initial model simulation defines, "truth" from which observations of the desired type, accuracy, and coverage can be sampled. Further simulations, with errors of various kinds introduced, are used as test beds into which observations can be assimilated by various means.

Studies of this type were performed by *Moore et al.* [1987], who used an OGCM and a reduced-gravity model to look at the potential for data assimilation in the Indian Ocean. Assimilating temperature data at every grid point once every 10 days is sufficient to cause rapid convergence of the model to the "true" solution; velocity data alone are insufficient. If temperature data are restricted to possible XBT routes, the assimilation technique used does not produce convergence, essentially because the scale of features (for example, in the Somali gyre) is too small to be resolved by the data. When the same restricted temperature data are assimilated into a linear reduced-gravity model (whose solution does not have such small-scale features), convergence is much more successful.

The relative role of temperature and velocity data was further examined by *Anderson and Moore* [1989]. They demonstrated that although specifying the thermal field is in general more important than the velocity field, for equatorial waves (especially Kelvin waves) the zonal velocity is also important. This is basically due to the energetics of oceanic adjustment: in most of the ocean, potential energy dominates, but Kelvin waves have an equipartition between potential and kinetic energy, and so the thermal field alone is insufficient to define a signal of the right amplitude. Of course, repeated assimilation of thermal data during a model integration will allow the Kelvin wave amplitude to approach its correct value. The possible role of salinity data was considered by *Cooper* [1988], who showed that in the Indian Ocean, initialization of an OGCM with temperature data while leaving the salinity field constant resulted in larger errors than not using any data at all. In realistic assimilation systems, such as those run today, the issue is how important salinity anomalies are to the density field and whether assimilating temperature data without corresponding information on salinity might worsen the model density field. Somewhat disturbingly, no studies seem to have been made of this.

In situ ocean data will always be sparse because of the vast size of the world ocean and the expense of making measurements, especially away from commercial ship tracks. Making the very best use of observational data is therefore important to an ocean analysis system. Satellite data offer the possiblity of global coverage but are unfortunately restricted to measurements of the surface of the ocean. Present-day instruments measure temperature and sea level. The latter is very important because it is related to the subsurface structure of the ocean, but the inverse problem of retrieving the subsurface density (or temperature) structure from sea level is difficult because the problem is underdetermined. Relatively straightforward techniques can make some use of the data [*Fischer and Latif*, 1995; *Fischer et al.*, 1997], although how effective they are at combining sea level and subsurface data in a real system remains unclear. The limits on the availability and type of ocean data have stimulated much interest in the use of advanced data assimilation techniques. These promise to make better use of limited data, but because of their cost, studies so far have been restricted to relatively simple models.

Optimum interpolation (OI) formally creates the best possible instantaneous analysis, given a certain set of observations, a background or "first-guess" field, and the appropriate statistical details of the errors in the inputs, as well as the statistics of the length scales and correlations of these errors. The key idea of a model-based analysis system is that the background field can be provided by the model running forward from the previous analysis; in this way both previous data and the surface forcing are able to contribute to any given analysis. Optimum interpolation, or approximations to it, is widely used both in meteorology and oceanography and can be considered as a baseline against which more advanced methods are to be measured. Two such methods are Kalman filtering and the use of four-dimensional variational techniques.

The basic idea of Kalman filtering is to describe not just the model state (temperature, velocity, etc.) but also the errors expected in the present value of the state. When the model is run forward, the evolution of the errors is also calculated, including an allowance for the imperfect model. At the next analysis time the expected errors in the background are then fed into the analysis procedure, and they help produce a better analysis than would be possible from simply using climatological values for the background error field. As the integration moves forward, the model converges to an optimum solution which makes full use not only of our knowledge of the data but also our knowledge of the evolving error statistics of the system. Unfortunately, to specify the error statistics properly requires not a vector of size $\mathbf{N}$, where $\mathbf{N}$ is the number of model degrees of freedom, but a matrix of size $\mathbf{N}^2$. Needless to say, this causes problems for complex models such as OGCMs.

Miller and Cane [1989] pioneered the use of the Kalman filter in tropical oceanography, applying it to a meridional mode model of the equatorial Pacific. Observed sea level data from six islands were assimilated. Error estimates suggested a reduction in sea level rms error of 1 cm or more across the waveguide; rms errors from the wind forced run varied spatially but had typical values of 5-6 cm. Independent data from four other islands supported the reduction in error. Clearly, sea level data at six points are insufficient to fully determine the system, but it was an encouraging start. *Miller* [1990] used the same model to look at the impact of island sea level and limited TAO mooring data on model sea level and showed that the mooring data gave a substantial improvement in the estimation of sea level in the eastern equatorial Pacific. *Miller et al.* [1995] used a low-resolution grid point model (2° by 5°) and included dynamic height data from XBT ship tracks as well as island sea level data. The high cost of the Kalman filter necessitated the coarse model resolution, and it was to circumvent this that *Cane et al.* [1996] introduced a procedure to approximate the calculation by using truncated empirical orthogonal function (EOF) expansions. This effectively reduces the number of degrees of freedom, which helps keep down small scale-noise as well as making the problem tractable for reasonable models. They used the method with a 0.5° by 2° model and were able to obtain results at least as good as *Miller et al.* [1995]. Kalman filter techniques have also been used to assimilate Geosat altimeter data into simple models of the Pacific [*Fu et al.*, 1991, 1993] and Atlantic [*Gourdeau et al.*, 1992]. Studies with the more accurate altimeter data now available from TOPEX/ POSEIDON are underway, and initial results are encouraging. For example, *Fukumori* [1995] used an approximate Kalman filter method to assimilate 1 year of data into a reduced-gravity model of the Pacific. The resulting solution shows significantly better fit to both island tide gauge data and zonal velocity mooring measurements when compared to a purely wind forced simulation (Figure 5).

The basic idea of the variational technique is to define the problem in terms of a cost function, which is to be minimized. The cost function is a function of the trial analysis and measures the difference between that trial analysis and the observations and background field in a way that accounts for the error statistics. Iterative methods can be used to find the trial analysis that minimizes the value of the cost funtion; this analysis is then the desired solution. In three dimensions the method is equivalent to an OI analysis and has a similar cost. The problem can be made four-dimensional by looking not just for an instantaneous analysis but a model trajectory through time. In this case the solution should minimize the differences between all the observations and the model trajectory. To carry out the minimization efficiently, it is necessary to create the adjoint of the model code, which roughly speaking runs the model backward. A number of iterations (forward and backward) over the assimilation period are needed, and so the method is significantly more expensive than the three-dimensional calculation. In principle the forcing fields as well as the initial conditions can be adjusted to create the optimum

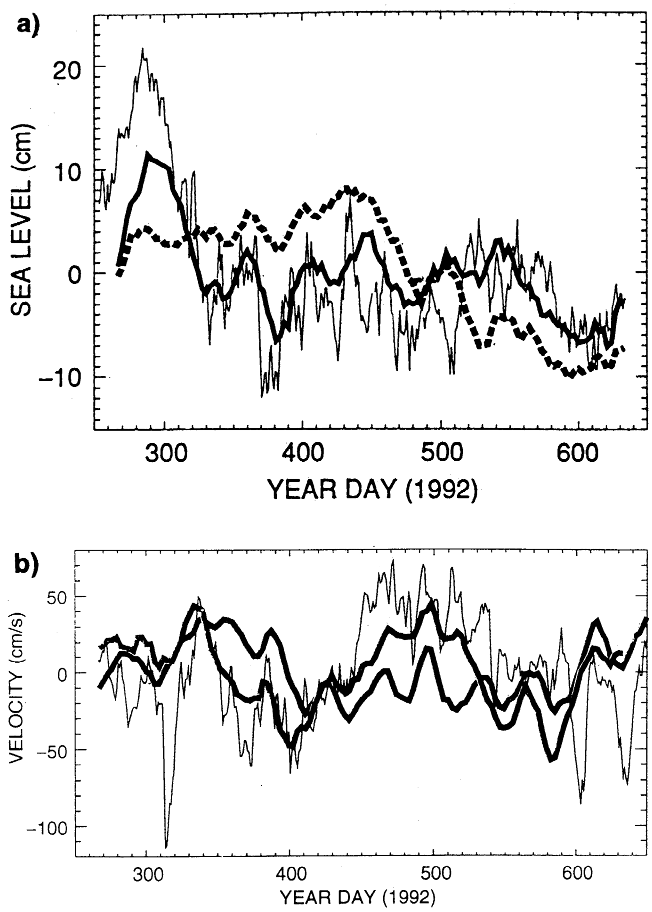


Figure 5. Comparison of an approximate Kalman filter assimilation of TOPEX sea level data with independent observations. (a) Sea level variability at Majuro (171°E, 7°N) from tide gauge (thin line), wind forced model (thick dashed line) and TOPEX assimilation (thick solid line). (b) Zonal velocity anomalies at 140°W from Tropical Atmosphere-Ocean (TAO) mooring at 25 m (thin line), wind-forced model (thick dark line), and TOPEX assimilation (thick light line). In both cases the assimilation is considerably better than the wind forced only simulation. From *Fukumori* [1995].

trajectory, although care must always be taken that the problem remains well determined. The four-dimensional variational method (4D-Var) makes explicit use of past (and future, if available) data to define an analysis. By taking a global view it can create a better analysis than simple OI, which carries information forward in time only in the model fields. A drawback is that 4D-Var does not take account of model error. Four dimensional variational techniques are being actively developed for use in global numerical weather prediction, for example at ECMWF.

The application of 4D-Var to tropical oceans was first studied in an idealized setting by *Long and Thacker* [1989a, b]. They found that a combination of complete sea level data and sparse in situ temperature data was sufficient to reasonably constrain a multilayer equatorial ocean model. Further work by *Sheinbaum and Anderson* [1990a] assimilated actual XBT observations of the depth of the 16°C isotherm into a linear reduced-gravity model. Some problems were found in the east, where the model simulation seemed incompatible with the observations. By repeating the assimilation with several different initial trajectories it was deduced that the data were constraining the solution moderately well across much of the equatorial region. *Kamachi and O'Brien* [1995] assimilated synthetic drifting buoy data into a single vertical model reduced-gravity model, and were able to recover upper layer thickness near to the equator, although not in some other regions. *Greiner and Perigaud* [1994] used Geosat altimeter data to adjust the thermocline depth in a nonlinear reduced-gravity model of the Indian Ocean. *Kleeman et al.* [1995] use a four-dimensional variational technique to assimilate subsurface thermal anomaly data into a simple ocean model, which is then used to make coupled model ENSO forecasts. The skill of the forecasts is significantly enhanced compared to the "wind-forced only" initialization, particularly at longer lead times of 6-15 months. This is the only application of an advanced assimilation technique that is in operational use at the moment. The relative merits of the Kalman filter and variational approaches to practical oceanographic problems are still contested, and it is not known in what form(s) advanced assimilation techniques will eventually be used in more complex models. Much work remains to be done in this area.

5.3.2. Data assimilation uses. Data assimilation can be used for preparing ocean analyses, for initializing ocean models prior to coupled forecasts, and for tuning and testing ocean models. Although ocean analyses can be prepared directly from the data [e.g., *Smith*, 1995], it is more common for an ocean model to be used. This may not always result in a more accurate analysis of the temperature field, but the use of a model is convenient for producing a complete description of the ocean state and is also a good starting point for other studies. Most effort has gone into producing comprehensive ocean analyses, which require an OGCM or something like it to represent the detailed temperature and velocity structure. The use of a complex model implies that the assimilation method must be relatively straightforward if it is to affordable.

Important work was done by *Derber and Rosati* [1989], who developed a full-scale global ocean data assimilation system. This used a relatively high resolution OGCM (1° by 1°, with 1/3° meridionally near to the equator), extensive quality control procedures, and an iterative approximation to a three-dimensional variational analysis. An analysis is made each model time step using data from a 30-day window, and the model thermal fields are then corrected. The continuous insertion of the thermal analysis into the model helps the velocity field adjust with the minimum of disturbance. This work was built upon by *Leetmaa and Ji* [1989], who set up an operational assimilation system for the tropical Pacific. This system has continued to run at NCEP ever since, albeit with enhancements, which have been made from time to time [*Ji et al.*, 1995]. The ocean analyses are used for monitoring purposes as well as initial conditions for coupled forecasts.

The Japan Meteorological Agency has used an ocean assimilation system for ENSO monitoring for some years; results are regularly published in their *Monthly Ocean Report*. OGCM-based assimilation systems for ENSO prediction are also being worked on by several other groups worldwide, but little published material is yet available. Another ocean analysis system is that developed by the Forecasting Ocean Atmosphere Model (FOAM) group at the UKMO. This is a global system, based on a 1° by 1° model, and uses a continuous insertion technique similar to that used in the UKMO atmospheric analyses. The U.S. Navy also has a global ocean analysis system [*Clancy et al.*, 1990]. Neither of these systems pays special attention to the equatorial oceans, although a tropical version of the FOAM system is in use.

The issue of initializing the ocean component of a coupled system prior to making a forecast gives rise to considerations other than the accuracy of the initial state. Two concerns are the "initialization" of the ocean model per se and how the ocean state affects the evolution of the coupled system. In numerical weather prediction it is necessary (at least with most analysis methods) to remove energy from the gravity waves which are otherwise falsely excited by inevitable errors in the analysis. Failure to do so causes problems such as unrealistic oscillations in surface pressure in the early stages of a forecast. In the case of the ocean both high-frequency inertia-gravity waves and inappropriate lower frequency equatorial signals have the potential to contaminate ocean model forecasts. The higher-frequency gravity waves are most likely to cause problems in the assimilation cycle itself because of their relatively short timescale. Most working assimilation systems deal with the problem by using a near-continuous updating of the ocean model fields, which tends to smooth out spurious high-frequency disturbances. Inertia-gravity waves are in any case fairly well damped in most models and will not persist much beyond a month.

Possible contamination of forecasts by low-frequency changes due to spurious imbalances in the initial conditions is more difficult to deal with. *Latif et al.* [1993] used a coupled GCM for ENSO forecasting, and rather than initialize the ocean model to be as close to reality as possible, they tried to initialize the ocean model to have the climatology of the equilibrium coupled system plus the anomaly of the real ocean. The idea was to minimize climate drift, which otherwise threatened to swamp the anomaly signal in the forecast. The opposite approach was used by *Stockdale* [1997], who initialized the ocean as accurately as possible and then removed the estimated mean drift (which was large) from the forecasts as an a posteriori correction. *Ji et al.* [1994a, b] instead adjust the coupling of the atmosphere and ocean models to be partly in anomaly mode when the forecasts are run. This reduces the climate drift of the system, even when the full ocean analysis fields are used to initialize the forecasts.

D. Chen et al. [1995] take a rather different approach to model initialization. They do not use any data other than the usual wind and SST forcing, but rather than force the ocean model to prepare an initial state, they assimilate the data into the coupled system. The essence of the method is to relax the wind field in the coupled model toward observed values (FSU anomalies), with a relatively weak relaxation near the equator but a stronger relaxation poleward of 5°N/S. The most noticeable effect of this is to produce a much smoother initial state: the ocean model forced by the FSU winds has a fair level of month to month variability, whereas the model assimilated winds keeps only the low-frequency ENSO signal. The impact on forecast skill is fairly dramatic, with significant improvement, at least in the period 1982-1992. The correct interpretation of these results is not yet certain. It may be that the coupled spin-up eliminates certain fast-decaying modes, thus allowing the "correct" low-frequency mode to dominate the initial conditions. It may be that the *Zebiak and Cane* [1987] model is upset by the higher frequency variability, and so removing it gives better forecasts of a possibly dominant slow mode of interannual variability. Or it may be that the procedure does actually produce a more realistic initial state, with the loss of any "signal" in the winds being more than compensated by the loss of noise. Until these issues have been better explored, it is difficult to draw any general conclusions about initialization techniques. In particular, coupled GCMs strive to reproduce a much wider range of processes and timescales, and they may have very different initialization requirements.

The final area in which assimilation has been used is that of model tuning and testing. *Yu and O'Brien* [1991] used a variational technique to estimate a vertical viscosity profile in the upper ocean from 10 days of wind and velocity data at a mooring in the Sargasso Sea. The drag coefficient was also estimated by the calculation, and the resulting values of both drag coefficient and viscosity profile were judged reasonable. In another study, *Smedstad and O'Brien* [1991] took a simple reduced gravity model of the Pacific and used the time record of sea level at three island stations to estimate the most appropriate value of c^2 for the model. The variable c^2 was estimated as a function of longitude with a 1^0 resolution, although a small amount of smoothing was needed to prevent noisy retrievals. The function retrieved was again quite plausible in terms of its spatial structure and the differences in it found between El Niño and non-El Niño conditions.

Data assimilation should ideally give a clear idea of how well the model is able to fit the data and as such should be a powerful tool for model testing and evaluation. The big problem with this is the uncertainty in the surface forcing. Indeed, errors in the surface-forcing can be a big problem in trying to create good ocean analyses. The paradigm we would like to have would be that our knowledge of surface forcing variability is incomplete, and by including extra data from the ocean subsurface we can create a better analysis. In practice, what we typically find is a conflict between the mean state of the forced model and the mean state of the data. The results of *Ji and Smith* [1995] are one example of this (Figure 6). Deciding whether the fault lies with the surface forcing or the model is, as ever, difficult, although the nature of the discrepancies can sometimes be suggestive one way or the other. Despite the difficulties of partitioning error, assimilation certainly gives a sense of how well a model system is performing. Given the size of increments typically required to keep an OGCM close to the data from the TAO moorings, much progress is still required. The discrepancies between the model/forcing and the data can cause problems for an assimilation system, and these discrepancies are the reason that stand-alone analysis systems (i.e., that do not include a numerical model) are not necessarily a bad idea. In the worst case the contention between model and data can cause spurious variations in the analysis. It is also a finding of *Ji and Smith* [1995] that the temperature data available for the equatorial Pacific are not sufficient to constrain the assimilation if the surface forcing is changed. For both these reasons, careful attention to surface forcing and/or model improvement will be needed to produce high-quality analyses.

6. Summary

The TOGA decade has seen a substantial development of the use of ocean models related to El Niño and the tropical oceans, although the foundations had been solidly laid in the late 1970s and early 1980s. As we have shown in this paper, much work has been done in modeling the mean seasonal cycle and interannual variability of the tropical Pacific Ocean. Particular attention has been given to modeling SST because of its importance for coupling to the atmosphere. Refinements in the

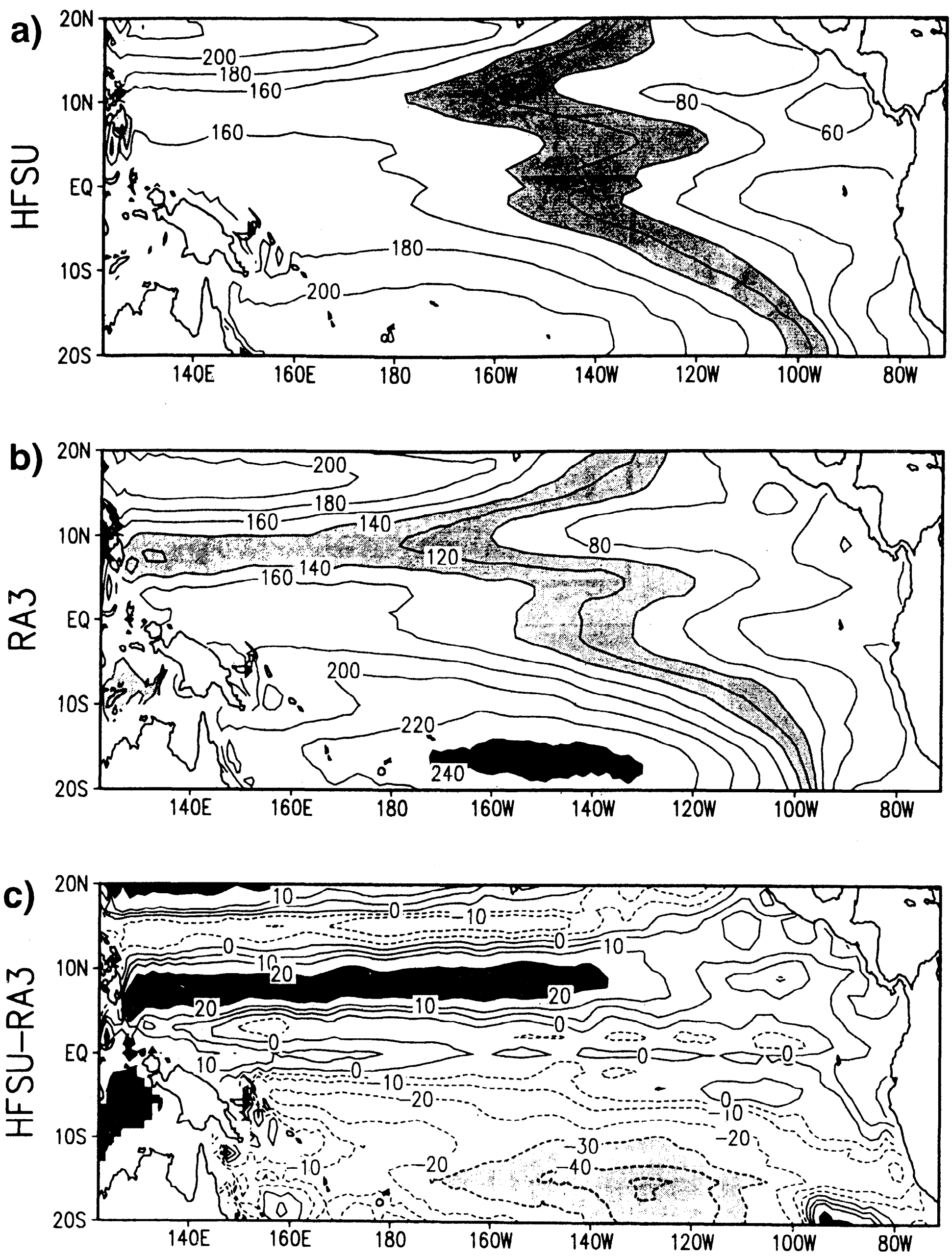


Figure 6. The 11 year annual mean depth (in meters) of the 20°C isotherm in the tropical Pacific. (a) OGCM simulation forced by a combination of *Hellerman and Rosenstein* [1983] and Florida State University (FSU) wind stresses. (b) The same model forced with the same wind, but including assimilation of subsurface thermal data. (c) The difference field. From *Ji and Smith* [1995]. Copyright American Meteorological Society.

models (primarily extra physics in the simpler models and extra resolution in the OGCMs) and careful attention to the forcing fields have led to apparently quite realistic simulations of the ocean mean state and its variability. Quantifying the success of the models has remained difficult, however, because of the uncertainties in the forcing fields.

Further indications of the abilities of our ocean models come from their performance in coupled systems. This is an area which has not been covered in this paper (see *Delecluse et al.* [this issue], for a full discussion), but the fact that coupled models, simple or complex, are able to produce realistic looking seasonal cycles and/or variability suggests that the ocean models are properly representing important aspects of reality. The fact that initial conditions from wind-forced ocean model runs are often able to produce skillful forecasts when coupled is a further indication of a real ability to model the variability of the system.

The large number of papers related to oceanic ENSO modeling might also be taken as indicative of the "success" of our ocean models, as might the increasingly common use of data assimilation in tropical ocean models. But, in fact, real and substantial problems remain, and a superficial impression that tropical ocean modeling is largely a "solved problem" is seriously misleading.

There are five major issues which arise from our review of TOGA ocean modeling. The first two of these can already be commented on in a sensible way, although future developments will doubtless change our perspective to some extent. The three remaining issues, however, remain as serious challenges for ocean modelers. They are important questions, and our present ability to answer them is very limited.

6.1. The Relative Performance of OGCMs and Intermediate Models

For anyone contemplating the use of an ocean model the choice as to which type of model to use is a fundamental one. There is no absolute answer as to which is preferable, however. It seems that an intermediate model must be of sufficient complexity (e.g., a variable mixed layer depth) to successfully simulate SST, although simple models can represent some aspects of interannual variability. It is also apparent that the lower resolution OGCMs tend to have significant problems with the equatorial cold tongue. The better high-resolution GCMs seem good but are, of course, quite expensive to run. The better intermediate models are also quite good, but there is always the risk that processes (e.g., those responsible for decadal variability) are missing. The choice of model to be used depends on the resources available and the problem in hand. Both types of model will doubtless continue to be developed and refined, testing their ability to handle various situations. There is certainly scope for proper intercomparisons (e.g., of the mean seasonal cycle and, perhaps more crucially, interannual SST variability) to help assess the relative strengths of these different classes of model.

6.2 Methods of Forcing Ocean Models

A number of different methods are used for forcing tropical OGCMs. The wind stress is generally specified as a given external forcing, but how to handle the heat fluxes is more difficult. The heat fluxes are important in any attempt to quantify the skill of SST simulations, but the feedbacks between heat flux and SST, coupled with our relatively poor knowledge of the fields, makes direct specification problematic. The two common approaches are 1: to specify a "best estimate" Q, plus a relaxation to observed SST, or 2: to provide an explicit model of the heat flux components and their dependence on model SST. The latter approach generally gives better insight into processes, since it forces attention to be paid to the terms of the heat budget. The former is more convenient, allows maximum exploitation of others' work on heat fluxes, and can be used for assessing the accuracy of model SST processes.

6.3. Knowledge of the Forcing Fields

Any experiment with an ocean model, coupled or uncoupled, requires the specification of forcing fields at the surface of the ocean, and it is these fields that drive and largely determine the behavior of the ocean model. We know that if we take our best estimates of these fields and use them to force an ocean model, we are able to qualitatively reproduce the features and variability of the tropical oceans, which confirms that we have some knowledge of these forcing fields; how much knowledge is unknown. Some data sets come with formal error estimates, usually derived from sampling limitations, but sometimes allowing for algorithmic uncertainty. The error bars on the climatologies are quite large. When it comes to interannual variability of the fluxes, the situation is yet more difficult. If we are only interested in relative variations rather than absolute values, then the problem is a little easier, but good studies of expected errors in forcing fields have not been published. Our knowledge of heat flux variability is poorer than that of wind stress; even the qualitative changes are uncertain. It is generally thought that heat flux variations are less important for the ENSO problem than wind. From an ocean-modeling perspective this is apparently true (wind-forced models reproduce El Niño), but this may only be because of how the heat flux in fact varies; at least one study shows that heat flux variability has the potential to drive substantial SST changes even in the Nino 3 region [*Schneider and Barnett*, 1995].

The impact of newer observing systems is also little known. The TAO array was designed to produce reasonable sampling of the surface wind in the near equatorial region. The data are most often used when merged with ship observations (the FSU product) or incorporated into numerical weather prediction (NWP) analysis products, and one would hope that ocean model simulations for the 1990s (with TAO data) are more accurate than those from earlier periods. Studies to confirm this have not yet been published, at least with regard to the interannual variability. Indeed, superficial experience with intermediate models forced by FSU winds suggests that there is no particular improvement in SST simulation (e.g., the El Niño in late 1994 was not well reproduced by many models). Whether this is because the ocean models are sufficiently poor and do not benefit much from an improved wind field, because there are problems in the surface heat fluxes, or because the last few years have just been atypical is not clear.

There is also an increasing use of satellite data to derive or constrain surface-forcing fields. Proper assessment of the quality of satellite-derived wind and surface heat flux fields is difficult because of the paucity of in situ data. However, the satellite data are probably already at a stage where they are useful, and the quality is likely to improve in the future. Satellite data are also important for operational NWP analysis systems. In principle, these systems can synthesize all available data in a consistent way and produce surface forcing fields. The various reanalysis projects (NCEP, NASA Goddard Data Assimilation Office (DAO), ECMWF) are designed to give consistent products over the 1980s and 1990s or longer and should be useful for ocean modeling. The overall quality of the surface fluxes is not guaranteed, however, since the input data are relatively sparse and vary from year to year and since the fluxes are highly dependent on the atmospheric model physics.

There is still much work to be done on the issue of forcing fields. One of the legacies of TOGA and other work in the last decade is that there has been real progress in observational data relevant to surface fluxes over the tropical ocean. This has not yet produced definitive, high quality surface forcing fields for the ocean modelers. It will be difficult to produce such datasets independently from ocean models, basically because fluxes are hard to measure. Progress is likely to require careful assessment of many simulations with different models and different forcing fields.

6.4. Performance of Our Ocean Models

Ocean models are evidently able to reproduce the general structure and circulation of the tropical oceans, as well as interannual variability in the equatorial waveguide. But how good they are in quantitative terms is less clear. "How good" an ocean model is is a slippery concept. Two criteria we will develop here are the following: is model error small enough that it is swamped by uncertainties in the forcing fields, and are the model errors large enough to have an impact on the phenomena being studied (primarily ENSO in the discussion here)?

We have already pointed out that errors in surface forcing are poorly known but probably serious, particularly when interannual variability is being simulated. Nonetheless, consensus opinion is that ocean model errors are playing a significant role too. The few studies that have attempted to look at this are based on the approaches by *Blumenthal and Cane* [1989] or *Frankignoul et al.* [1989] and were discussed in sctions 5.2 and 4.2, respectively. They typically demonstrate that errors in a simulation are larger than can be accounted for by forcing error. The ability to separate model error from forcing error depends on the fields considered; errors in the simulation of interannual variability of equatorial SST have not been shown to be due to model errors, for example. Even where formal demonstrations cannot be made, modelers often suspect that their model is in error. A better handle on this problem would be useful, however. It may well be that the use of TAO data (wind and subsurface temperature), perhaps together with data assimilation techniques, will enable a clearer picture of ocean model error, at least beneath the surface. Obtaining heat and momentum fluxes accurate enough to constrain surface mixing processes in the models may be more difficult.

The second criterion for ocean model error is whether it is large enough to be important. If the aim of the model is to able to simulate SST changes accurately enough to predict the development of El Niño, then this criterion is very tight. Although large El Niño events have an amplitude of several degrees Celsius, SST variability is usually much smaller than this. Even relatively small SST changes apparently can affect the atmosphere, and the coupled feedbacks that occur (or not, as the case may be) mean that SST errors can amplify. Experience with atmosphere GCMs suggests that it would be desirable to keep large-scale SST errors at or below a level of a few tenths of a degree Celsius if the errors are not to influence the evolution of the coupled system. Present-day model simulations are not this good. How well today's models would perform with perfect forcing is not known. We do not know if our ocean models are good enough for reliable ENSO forecasting.

Although the impact of model error on interannual variability is hard to determine, there are plenty of other indications that ocean models are not perfect. Subsurface temperature structure and currents differ appreciably from those observed; OGCMs, for example, tend to have too diffuse a thermocline along the equator. Data assimilation studies show that large and continuous increments are needed to keep the model close to the data. Ocean modelers should certainly not be complacent, even if the relevance of these subsurface problems for ENSO processes is unknown.

6.5. How Ocean Models Might Be Improved

Our present day models of the tropical oceans do a reasonable job, but they are doubtless some way from being perfect. As improvements in the specification and understanding of surface forcing take place, some of the shortcomings of ocean models will become more clearly exposed. This is good, but it still leaves open the question of how our models might be improved.

There are two perspectives on this question. First, which oceanic processes are important for ENSO, and how accurately do they need representing? And second, how can we improve the physics of our models, and how much effort will this require? The first perspective is most easily seen from the intermediate models. What are the major causes of errors in the simulation of SST variability? Are some nonnegligible processes being omitted, or is the problem with inadequate paramaterization of entrainment, mixed layer processes, or something else? Answers to these questions would be very useful for any modeler, although it may well be that more progress is made empirically by testing out enhancements and improvements rather than by attempts to carefully analyze model error.

The second perspective concerns our ability to make the models look more like reality. A GCM is supposed to represent everything required to determine the large-scale circulation, and in this context the emphasis may not be on asking which processes are more relevant but seeking to put right anything that is wrong. Here, too, there are unanswered questions. It is believed that errors in the subsurface thermal structure are related to problems with the parameterization of vertical mixing. Many ideas exist for improving this, but it is not clear whether any of them are completely adequate. We do not know how much effort may be required to produce suitable schemes, or even whether it is a problem that can be solved at all, in a form that is both accurate and computationally reasonable. The impact of imperfect forcing and verifying data on model development and improvement is also not known.

Despite these five areas of questioning, there are some grounds for optimism. Improvements in computer technology mean that more people are able to run more experiments with more sophisticated models. In the coming years it will be feasible for most scientists to use models of choice rather than necessity. There has also been significant development in the data available for ocean modeling. The TAO array is the most significant new data source for the Pacific, but satellite data are also increasingly important, and many other observing systems play their part. It may sometimes seem that ocean modelers are slow to exploit these new data, but in no way should this be taken as a lack of interest. Sometimes it just takes a long time to catch up with the observational data, and it is likely that the maximum benefit of many of the TOGA measurements will be extracted over the next decade or so. Maximum benefit will also require continuity of the observing systems over many years to come.

It is not just in the availability of data that progress elsewhere will help ocean modelers. Improvements in atmospheric modeling are being driven by the NWP and climate change communities, both of which are well resourced. Better atmosphere models may help to expose problems in the ocean models, both in coupled experiments and in creating improved analyses of the surface-forcing fields. A further impetus for the ocean-modeling community comes from the continued interest in and resources for climate variability and seasonal forecasting. Although TOGA is now at an end, its work continues, both through Climate Variability and Predictibility (CLIVAR) and, increasingly, the involvement of operational agencies. It is important that proper support for ocean modeling is provided, particularly in terms of human resources; the tropical ocean modeling community is quite small. We trust that they are nonetheless able to exploit the many new opportunities available and make substantial progress in tropical ocean modeling.

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Appendix: Additional References

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