Tellus (1998), 50A, 76–94
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TELLUS
ISSN 0280–6495

Initial conditions and ENSO prediction using a coupled ocean-atmosphere model

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(Manuscript received 27 June 1996; in final form 5 September 1997)

ABSTRACT

A coupled ocean-atmosphere initialization scheme using Newtonian relaxation has been developed for the Florida State University coupled ocean-atmosphere global general circulation model. The initialization scheme is used to initialize the coupled model for seasonal forecasting the boreal summers of 1987 and 1988. The atmosphere model is a modified version of the Florida State University global spectral model, resolution T-42. The ocean general circulation model consists of a slightly modified version of the Hamburg's climate group model described in Latif (1987) and Latif et al. (1993). The coupling is synchronous with information exchanged every two model hours. Using ECMWF atmospheric daily analysis and observed monthly mean SSTs, two, 1-year, time-dependent, Newtonian relaxation were performed using the coupled model prior to conducting the seasonal forecasts. The coupled initializations were conducted from 1 June 1986 to 1 June 1987 and from 1 June 1987 to 1 June 1988. Newtonian relaxation was applied to the prognostic atmospheric vorticity, divergence, temperature and dew point depression equations. In the ocean model the relaxation was applied to the surface temperature. Two, 10-member ensemble integrations were conducted to examine the impact of the coupled initialization on the seasonal forecasts. The initial conditions used for the ensembles are the ocean's final state after the initialization and the atmospheric initial conditions are ECMWF analysis. Examination of the SST root mean square error and anomaly correlations between observed and forecasted SSTs in the Niño-3 and Niño-4 regions for the 2 seasonal forecasts, show closer agreement between the initialized forecast than two, 10-member non-initialized ensemble forecasts. The main conclusion here is that a single forecast with the coupled initialization outperforms, in SST anomaly prediction, against each of the control forecasts (members of the ensemble) which do not include such an initialization, indicating possible importance for the inclusion of the atmosphere during the coupled initialization.

1. Introduction

Accurate short-range forecasting is a stringent test for any coupled ocean-atmosphere general circulation model (CGCM). Short-range forecasts depend on knowledge of the ocean's 3-dimensional thermal, momentum and density structure and also require sufficient coupling strength between the variables at the ocean-atmosphere interface. The duality between accurate specifications of the initial conditions and the proper representation of the interactions between the ocean and atmospheric models places large prerequisites on both the models and initialization scheme used to generate seasonal forecasts.

It is known that CGCMs suffer some degree of drift during the course of integration. The drift, if

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large, can significantly lower the skill of the forecast. This problem results from the fact that the mean state of the coupled system is calculated rather than imposed. Drifts manifest themselves for several reasons, most notable are imbalances in the initial conditions and inadequate model physics. Inadequate physics at the air-sea interface results in large differences between observed and coupled model fluxes during an integration which through coupled feedbacks produce the model drift.

Traditionally, ocean models are initialized using climatological forcings and/or atmospheric model derived forcings. These ocean model initializations are usually done "off line", that is they are not coupled to an atmospheric model and therefore lack the coupled interactions needed to reduce the initial condition imbalances. Ghil and Malanotte-Rizzoli (1991) noted that initializing the individual components of the coupled system will not initialize the coupled system since the variability of the coupled system is much different from any simple function of its parts. A way of reducing the drift, due to imbalances in the initial conditions, is to assimilate observed surface and subsurface information directly into the ocean model during the course of a coupled integration. Coupled ocean-atmosphere data assimilation systems (COADAS) are now being developed for the purpose of reducing the initial drift which will allow for higher forecast skill of the coupled system by obtaining accurate and balanced initial conditions.

Several groups are investigating using various forms of coupled data assimilation for their coupled models. National Center for Environmental Prediction (NCEP) uses an end-to-end data assimilation scheme for their couple GCM (Ji et al., 1994). This scheme is based on the method of variational objective analysis (Derber and Rosati, 1989) and is used to continuously assimilate surface and subsurface temperature data into the ocean model. This continuous data insertion of temperature corrections enables the ocean model circulation to slowly adjust to changes in the thermal field. A high level of dynamical consistency was maintained between the analysis and forecast components so that the analyzed ocean initial state does not undergo a sudden drift at the beginning of the forecast. Considerable effort was made to ensure that the atmospheric physical parameterizations produced sufficient wind stress in order to maintain the ocean subsurface structure during the coupled forecast.

The same ocean data assimilation scheme employed by NCEP, was used by the Geophysical Fluid Dynamics Laboratory for the purpose of examining the ability of the data assimilation scheme to capture the temporal variations within the tropical Pacific Ocean (Rosati et al., 1997). This integration conducted over the decade of 1979 to 1988 demonstrated the ability of this data assimilation scheme to capture the large scale oceanic features. While the vertical temperature structure was realistically simulated including the El Niño-Southern Oscillation (ENSO) signature the ocean current structure did not agree well with observations.

Rosati et al. (1997) also examined the impact of using Newtonian relaxation as a data assimilation scheme. The impact of nudging the SST field and a combination of the SST and surface wind fields was investigated. The surface wind field analysis was obtained from the twice daily analysis of NCEP while the SST was from the NCEP monthly analysis. The study concluded that the wind field alone was not enough to define the subsurface temperature structure and that the temperature/wind nudging was more realistic than the SST only assimilation. This result might be model dependent.

Chen et al. (1995) demonstrated that the skill of the SST forecast using a version of the Zebiak and Cane coupled model (Zebiak and Cane, 1987) could be significantly increased if the model was initialized with a combination of observed wind stresses and model derived wind stress. They noted an increase in skill at the beginning of the forecast was carried into intermediate and long lead times.

In this paper we propose using a four dimensional data assimilation/initialization technique based on Newtonian relaxation (hereafter labeled as NR) to initialize our global coupled model. NR has a long history as a initialization technique in numerical weather forecast models (Hoke and Anthens, 1976; Ramamurthy and Carr, 1987; Krishnamurti et al. 1991, Yap, 1995). This technique is being extended and applied to our coupled model for a one year period during which time the relaxation coefficients in both the atmospheric and ocean models are varied in time but not in space. A 3-month seasonal forecast is then conducted using the final relaxed state as initial

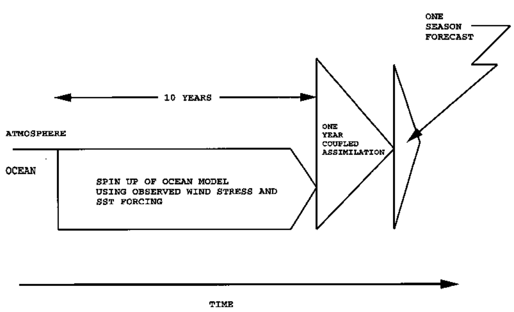

Fig. 1. Schematic of the coupled model initialization.

conditions for the coupled model. It is hypothesized that the one year NR will act as a dynamic initialization by allowing the coupled system to evolve towards a balanced initial state without serious model drift. Fig. 1 shows schematically the coupled NR procedure.

This paper is organized as follows. Section 2 discusses the atmospheric and ocean models used in this study. The NR technique employed and results during the initialization are discussed in Section 3. The results from the ensemble integrations are shown in Section 4. Discussion and conclusions are given in Section 5.

2. Models

2.1. *Atmosphere model*

The atmospheric component of the coupled system is a version of the Florida State University global spectral model. The model is truncated at horizontal resolution of 42 waves (T-42), which corresponds to approximately 2.8° by 2.8° latitude longitude in the tropics. The vertical structure is represented by 12 unevenly spaced σ-levels. The physical parameterizations include fourth order horizontal diffusion (Kanamitsu et al., 1983), modified Kuo-type cumulus scheme (Krishnamurti et al., 1983), shallow convection (Tiedke, 1984), large scale condensation (Kanamitsu, 1975). The long and shortwave radiative fluxes are after Chang (1978), surface energy balance is coupled to similarity theory (Krishnamurti et al., 1991) with the surface fluxes calculated via similarity theory (Businger et al., 1971). The parameterization of the low, middle and high clouds are based on threshold relative humidity values. Vertical turbulent transport for heat, momentum and moisture within the atmosphere are parameterized based on exchange coefficients that are functions of the Richardson number (Louis, 1981). Details of the model's physical parameterizations can be found in Krishnamurti et al. (1990).

Since snow is not yet forecasted in the atmospheric model, the European Centre for Medium Range Weather Forecasts (ECMWF) analyses of monthly snow depth were used to determine the snow coverage. Snow depths greater than 20 cm were assumed to completely cover the ground with an increased ground albedo. Linear interpolation between months were used to modify the snow cover and the resulting albedo field. The snow depth fields were used during the initialization and free forecast phase.

2.2. *Ocean model*

The model is an extended version of the ocean model developed by Ernst Maier-Reimer and used by Latif (1987) to study the ENSO phenomenon in the tropical Pacific. It is a primitive equation model using the Boussinesq and hydrostatic approximation on a horizontally staggered Arakawa and Lamb (1977) E-grid. It includes all three major oceans. The meridional resolution is variable, within 10° of the equator the resolution is 0.5°, it decreases to 5.0° near the northern and southern boundaries located at 70°N and 70°S respectively. In the zonal direction the resolution is constant at 5.0°. The eastern and western coast lines are realistic while the northern boundaries are closed. The model contains 17 irregularly space vertical levels with 10 levels located within the upper most 300 meters. Realistic bottom topography is included. No rigid lid assumption is made; therefore, the sea level is a prognostic variable. The freshwater flux is predicted in this model although since the coupled model is dynamically active between $\pm 55°$, the salinity is relaxed using Levitus (1982) near the poles with a Newtonian formulation with a relaxation constant of 30 days. The ocean model does not include a sea ice model.

The ocean model has two main processes that affect the mixing of heat, salinity and momentum. A Richardson number dependent mixing scheme based on Pacanowski and Philander (1981) and a mixed layer parameterization following Latif et al. (1994b). The mixed layer is diagnosed according to the difference between the surface temperature and subsequent layers below. If this temperature difference is less than a preset ΔT then the layer is considered to be located within the mixed layer which results in the addition of a fixed value to the vertical eddy diffusivity/viscosity deduced from the Richardson number dependent mixing. For this study we have chosen $\Delta T = 0.5^{\circ}$ and the extra mixing term is assigned the value of $2 \cdot 10^{-3}\,\mathrm{m}^2\,\mathrm{s}^{-1}$ for both the viscosity and diffusivity. These values are based on the work of Latif et al. (1994b).

The coupling between the models occurs every ocean model time step (2 hours). Fluxes of heat, momentum and fresh water are passed to the ocean model, and the SST is returned to the atmosphere. Bi-linear interpolation is used to interpolate between the ocean and atmospheric grids.

3. Data initialization

3.1. *Ocean only initialization*

Observational and modeling studies show the importance of the slow modulations within the equatorial Pacific upper ocean heat content in determining the variability of ENSO events. For this reason we have blended the Florida State University pseudo wind stress analysis (Stricherz et al. 1992) for the tropical Pacific with the global Hellerman and Rosenstein (1983) climatological wind stresses. The blended wind stresses were linearly interpolated between months with the maximum weighting given to the 15th of each month. The spin-up of the ocean model was performed using the blended wind stresses for the period January 1976 to May 1986. Starting in January 1980, the ocean temperature equation for the upper ocean layer was modified during the integration by applying a Newtonian relaxation term to the top most temperature equation, i.e.,

$$\left.\frac{\partial T}{\partial t}\right|_{\mathrm{level}=1} + \cdots = -\lambda\,(T|_{\mathrm{level}=1} - T_{\mathrm{obs}}), \tag{3.1}$$

where λ is the Newtonian relaxation coefficient whose value was chosen to be 4 day^{-1}, T_{obs} is the monthly mean observed SSTs from Reynolds (1988). This allows the ocean model's mixing parameterization to determine the subsurface temperatures based on the surface temperature relaxation.

3.2. *Coupled initialization*

A 1-year time-dependent coupled Newtonian relaxation was conducted using the coupled model. Initial conditions for the ocean model were those obtained at the end of a 10-year spin-up and for the atmospheric model, the initial condition was the 12 UTC ECMWF analysis for 1 June 1986. One year was chosen for the coupled initialization since it represents the minimum transit time for the second baroclinic mode (Kelvin plus Rossby wave) to cross the Pacific basin. In addition to establishing the upper ocean adjustment, the one year coupled NR also initializes the atmospheric model's low-frequency components and acts as a dynamical filter on the higher frequencies associated with imbalances between the two model states.

NR of the atmospheric model spectral equations take the form:

$$\frac{\partial A_l^m}{\partial t} = F_l^m(A,\,t) + N(A,\,t)\,(A_l^{om} - A_l^m), \tag{3.2}$$

where $N(A,t)$ is the time dependent relaxation coefficient, A_l^m represents the future value to which the NR is going towards, $F_l^m\,(A,\,t)$ is the forcing term for the equation for variable A_l^m. The NR was done in two steps, first the time tendencies for the model forcing were determined, next the relaxation term is added to the time tendencies. The future values which the relaxation goes towards are the spectral coefficients from the daily 12 UTC ECMWF analyses.

The choice of which variables to relax and the strength of the relaxation is model dependent. In the atmospheric model, NR was conducted on the time tendencies for the vorticity equation at all model levels, the divergence tendencies at the lowest three model levels, temperature tendencies at the upper three model levels and dew point depression tendencies in the polar regions. We chose to carry out the NR at all levels of the vorticity and only the lowest three models for the divergence for several reasons. In the extra-tropics the rotational part of the wind is generally considered to be more important than the divergent part due to the fact that it is the rotational component which represents the slow manifold of quasi-geostrophic motion (Daley and Puri, 1980). Therefore by adjusting the vorticity field the large-scale motions in the extra-tropics are adjusted. Relaxation of the divergent wind field in the lowest model levels allows the model to simulate the wind field associated with tropical convective systems and helps to resolve better the tropical boundary layer structure.

Due to the coarse vertical structure of the atmospheric model (12 vertical level extending to approximately 100 hPa) the temperature in the upper levels have a tendency to cool over both polar regions an average 5°C month^{-1}, therefore, the upper three temperature levels were relaxed in order to maintain proper thermal wind structure. However, only the top model temperature level was relaxed fully, while the magnitude of the relaxation coefficient for the remaining two levels from the top were decreased 25% and 50%, respectively. In the first three model levels near the surface, the dew point depression was relaxed in both polar regions in order to maintain proper radiation balance in the lower atmosphere. Similar to the temperature equation, the dew point depression was fully relaxed at the level closest to the surface.

The task of finding an optimal value for $N(A,t)$ forms the main difficulty in using the NR technique (Stauffer and Seaman, 1990). If the NR coefficients are too large then the relaxation term will dominate over the model forcing and no balance will be achieved within the coupled system. Conversely, if the relaxation coefficients are chosen too small then the analysis will have little effect on the model solution. Additionally, underestimation of the NR coefficients includes the possibility that the coupled system will significantly drift during the one year NR, while overestimation of the NR coefficients carries with it the possibility of large model drifts during the seasonal forecast.

As in atmospheric only modeling studies, the NR coefficients for this study were determined experimentally. Table 1 shows the initial relaxation coefficients used. These values are generally an order of magnitude smaller than the atmospheric studies mentioned above; amounting to approximately only a 3% weighting given to the ECMWF analysis. The small relaxation coefficients were chosen for two reasons. First, the NR coefficients were designed for long term integration; therefore, smaller values allow for the coupled system to evolve towards a final state with little influence from the analysis. Second, initial experiments with higher coefficients, such as those used in atmospheric only studies, resulted in excessive rain rates over large portions in the tropical Pacific during the boreal spring. Ultimately, such large rain rates produced negative feedbacks on the ocean model simulation and was found to be unsatisfactory (LaRow, 1997). In order to provide a smooth transition from the initialization phase to the free forecast, the relaxation coefficients were decreased linearly 70% during the one year initialization.

Table 1. *Model variables and values of the initial relaxation coefficients chosen for the coupled model*

Model variable	Initial relaxation coefficient
vorticity	$1.0 \cdot 10^{-5}\ s^{-1}$
divergence	$0.5 \cdot 10^{-5}\ s^{-1}$
temperature	$0.5 \cdot 10^{-5}\ s^{-1}$
dew point depression	$5.5 \cdot 10^{-6}\ s^{-1}$
SST	$4.6 \cdot 10^{-5}\ s^{-1}$

3.3. Annual means

Due to the relatively weak relaxation coefficients chosen, the possibility exists for model drift to occur during the initializations which could dampen or reduce the model's ENSO responses. The ability of the coupled model to simulate variability during the initialization phase was therefore assessed by examining annual mean fluxes of freshwater, net heat, and wind stress from the two, one-year coupled initializations over the tropical Pacific (Fig. 2). The annual means were calculated from 1 June 1986 (87) to 1 June 1987 (88). The two initializations produced differences in these fields, some are explained by variations associated with the swing in the ENSO event, while other differences are attributed to weaknesses in model parameterizations.

Climatological patterns are simulated in the freshwater flux fields (Figs. 2a, b) during both initialization periods with the subtropical high pressure centers exhibiting greater evaporation while the ITCZ region is characterized by larger

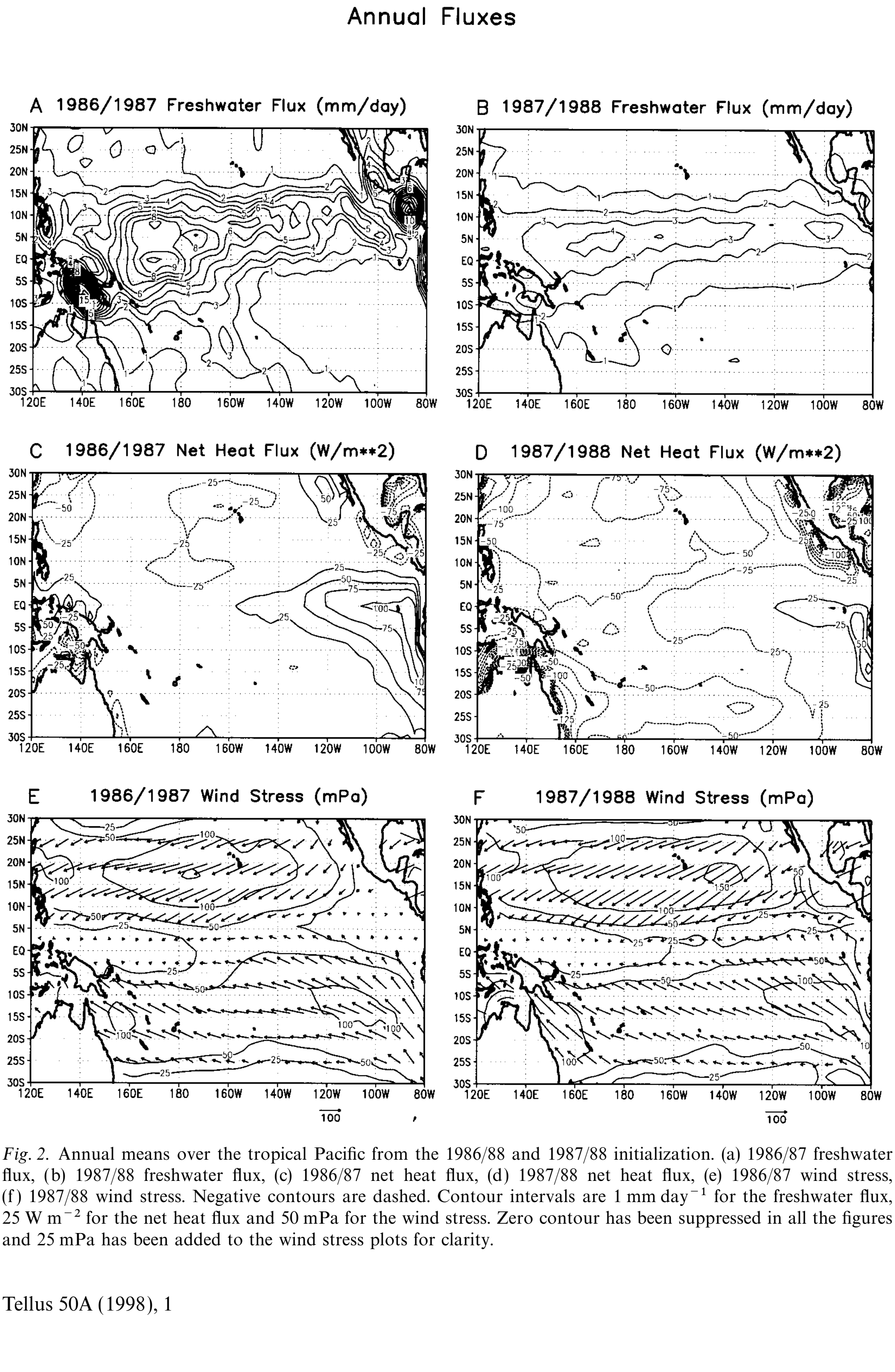

Fig. 2. Annual means over the tropical Pacific from the 1986/88 and 1987/88 initialization. (a) 1986/87 freshwater flux, (b) 1987/88 freshwater flux, (c) 1986/87 net heat flux, (d) 1987/88 net heat flux, (e) 1986/87 wind stress, (f) 1987/88 wind stress. Negative contours are dashed. Contour intervals are 1 mm day^{-1} for the freshwater flux, 25 W m^{-2} for the net heat flux and 50 mPa for the wind stress. Zero contour has been suppressed in all the figures and 25 mPa has been added to the wind stress plots for clarity.

precipitation amounts. During the 1987/1988 initialization, weaker gradients in the freshwater flux field are noted (Fig. 2b). This occurs in conjunction with the stronger surface easterly flow (Fig. 2f) and enhanced latent heating in the equatorial region. This perhaps indicates that the atmospheric model fails to conserve moisture or that the initialization is working towards producing a more stable atmosphere by enhancing the vertical mixing.

The initialization does simulate the basic climatological (Oberhuber, 1988) features of the net heat flux during both years (Figs. 2c, d). The 1986/87 initialization overestimated the net heat flux by approximately 25 W m^{-2} along the coast of South America, compared to the Oberhuber (1988) climatology. This is related to the coarse zonal resolution of the ocean model and an overestimation in the amount of shortwave radiation reaching the surface along the coast (LaRow, 1997). The overestimation in the shortwave radiation is linked to an underestimation of the stratus cloud cover, which is a known weakness in many coupled models (Latif et al., 1994a; Stockdale et al., 1994). During the 1987/88 initialization, the model generated negative heat flux covering most of the tropical Pacific, with the only exception in the eastern equatorial Pacific. Large negative net heat fluxes approaching -50 W m^{-2} are located in the far western equatorial Pacific during the 1987/88 initialization, while small positive values are found here during the 1986/87 initialization. Barnett et al. (1991) indicate that climatological annual mean values are closer to the latter value.

The differences in the net heat flux between the 1986/87 and 1987/88 initialization can be explained by the larger latent and sensible heating associated with the stronger low level circulation during 1987/88. The latent and sensible heating were an average 20 W m^{-2} and 10 W m^{-2}, respectively larger during the 1987/88 initialization. The remaining contribution to the differences arose from differences in the surface shortwave and longwave radiation terms.

The coupled initialization was able to simulate the proper response in the wind stress associated with the warm and cold ENSO phases (Figs. 2e, f). As is generally observed with evolving cold events, the 1987/88 initialization produced stronger easterly stresses covering a larger portion of the Northern Hemisphere Pacific region compared to 1986/87. The maximum wind stresses (150 mPa) are displaced east of Hawaii, near 17°N and 140°W during the 1987/88 initialization, while during the 1986/87 initialization the maximum is found further to the west, near 17°N and the date line. Furthermore, during the evolving cold event, the model generated stronger wind stresses in the eastern equatorial Southern Hemisphere. These increased wind stresses enhanced the coastal upwelling and reduced the model's positive net heat flux along the coast of South America (Fig. 2d). In general the two initializations were able to reproduce the general characteristics associated with the changing ENSO conditions. Both initializations showed pronounced variability between the two years.

4. Ensemble results

From a coupled model perspective, seasonal predictions are assumed to be almost entirely dependent upon the initial conditions (prediction of the first kind). Debate exists within the climate community as to the merits of atmospheric initialization when concerned with coupled modeled forecasts. Current "thinking" says that accurate specification of the initial state of the ocean is more important than the atmosphere, due to the larger inertia and slower response time of the ocean compared to the atmosphere. While this is true, studies such as Chen et al. (1995) and Anderson and Gudgel (1996) have shown that coupled model SST skill can be increased if the atmospheric model's forcings are allowed to participate in the ocean initialization. In this section we compare seasonal forecasts for the boreal summers of 1987 and 1988 using initial conditions provided by the initialization scheme discussed in section 3 against ten different atmospheric initial states for each season.

4.1. *Experimental setup*

The impact of the coupled initialization on the free forecast was examined by performing two 10-member ensemble integration, one for each season. Free forecast here denotes integration of the coupled model without any relaxation being applied to the model. Each ensemble consists of ten 90-day integrations; one ensemble for each

June-August period. The atmospheric initial conditions for the ensemble are the 12 UTC ECMWF daily analysis separated by two days centered on 1 June of the respective years (e.g., 21, 23, 25, 27, 29 May and 1, 3, 5, 7, 9 June). These atmospheric initial conditions did not participate in the coupled initializations. The ocean's initial conditions for all the integrations are the final ocean analysis at the end of the appropriate one year coupled initialization, i.e., 1 June 1987 and 1 June 1988.

The selection of the atmosphere to form the basis of the perturbation for the ensembles rather than the ocean fields was done for several reasons. First, it is known that the memory of the coupled system lies in the ocean. Therefore, by perturbing the ocean the memory of the coupled system is altered. Also, perturbing the ocean's initial state essentially discards the results of the coupled initialization. Additionally, the atmosphere's response time is much faster than the ocean's; therefore, by selecting initial conditions five days on either side of the initial time, it is hoped that the atmospheric model's phase space is adequately sampled.

4.2. *Initial conditions*

To highlight the differences in the initial conditions between the ECMWF analyses used to form the ensembles and the final state reached after Newtonian relaxation, meridional averages between 5°S to 5°N across the Pacific were calculated for the mean sea level pressure, surface temperature, 1000 hPa zonal winds and the 850 hPa temperature. The mean of each ensemble is plotted against its respective initialized counterpart for June 1. In this section, Reynolds (1988) weekly SST analysis are plotted for the surface temperature of the ensemble.

As seen in Figs. 3a–d, the initialized 1 June 1987 and the ensemble average analyses all display similar east-west trends across the Pacific except in the eastern Pacific where the mean sea level pressure and 850 hPa temperatures diverge from the analysis (Figs. 3a, d). The large decrease in the surface temperature seen with the initialized model in the western Pacific is a reflection of the New Guinea land mass which is included as part of the surface temperature. The initialization produced generally weaker easterly 1000 hPa winds with larger variability in the central and western Pacific compared to the mean of the ensemble (Fig. 3c).

Initial differences between the initialized 1988 and average of the ensemble for 1988 are shown in Fig. 4a–d. In contrast to the 1987 initial conditions, the 1987/88 initialization drifted less with respect to the ECMWF analyses. In contrast to the 1 June 1987 850 hPa temperature (Fig. 3d), the 1987/88 initialization underestimated the 850 hPa temperatures (Fig. 4d) by an average of almost 2 K across the Pacific basin. The initialization did produce a slightly weaker east-west pressure gradient in the eastern Pacific basin (Fig. 4a), however the total east-west gradient across the basin is very similar. The model's 1000 hPa zonal wind reflects the weaker pressure gradients in the eastern Pacific (Fig. 4c). Additionally the surface temperatures (Fig. 4b) were reproduced by the initialization in the central Pacific, with only the far western and eastern Pacific having temperatures greater than 1°. The cause of the large cooling seen in the western Pacific is due to the large negative net heat flux ($-50\ \mathrm{W\ m^{-2}}$) the model simulated during the 1987/88 initialization (Fig. 2d). These large temperature differences in the western Pacific were not noted during the 1986/87 initialization. In fact the model was able to simulate the western Pacific temperatures to within 0.5° (Fig. 3b). This is partly due to the correct magnitude and sign of the net heat flux which was simulated during the 1986/87 initialization. The large variation in the intensity of the 1000 hPa zonal winds seen during both initialization was due to the fact that these are instantaneous values.

Zonal mean cross sections of the specific humidity obtained at the end of each initialization were compared to the ensemble's initial conditions in Figs. 5a, b. The zonal means only cover the Pacific (120°E–80°W) basin. The 1986/87 initialization produced a 1 June boundary layer structure which has higher specific humidities than the corresponding ensemble means. These differences over the boundary layer were not as large for the 1987/88 initialization.

5. Seasonal forecast results

As shown in the Subsection 4.2, exact matches to the ECMWF analysis were not obtained at the

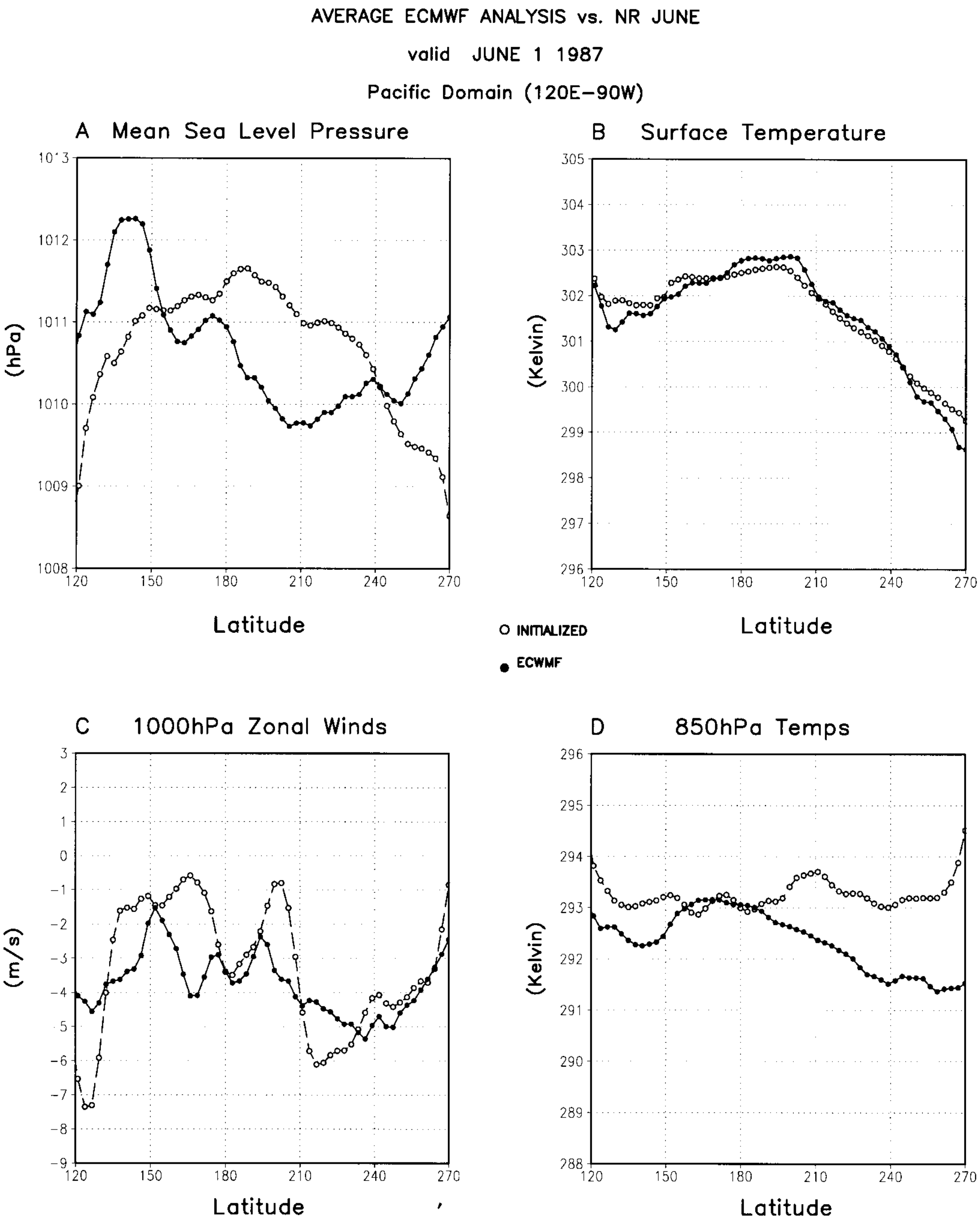

Fig. 3. 1 June 1987 meridional average between 5°S to 5°N across the Pacific Ocean from the initialized model and ensemble. (a) Mean sea level pressure. (b) Surface temperature. (c) 1000 hPa zonal wind. (d) 850 hPa temperatures. Solid line with closed circles are the average from the ensemble and the dashed line with the open circles are the initialized fields.

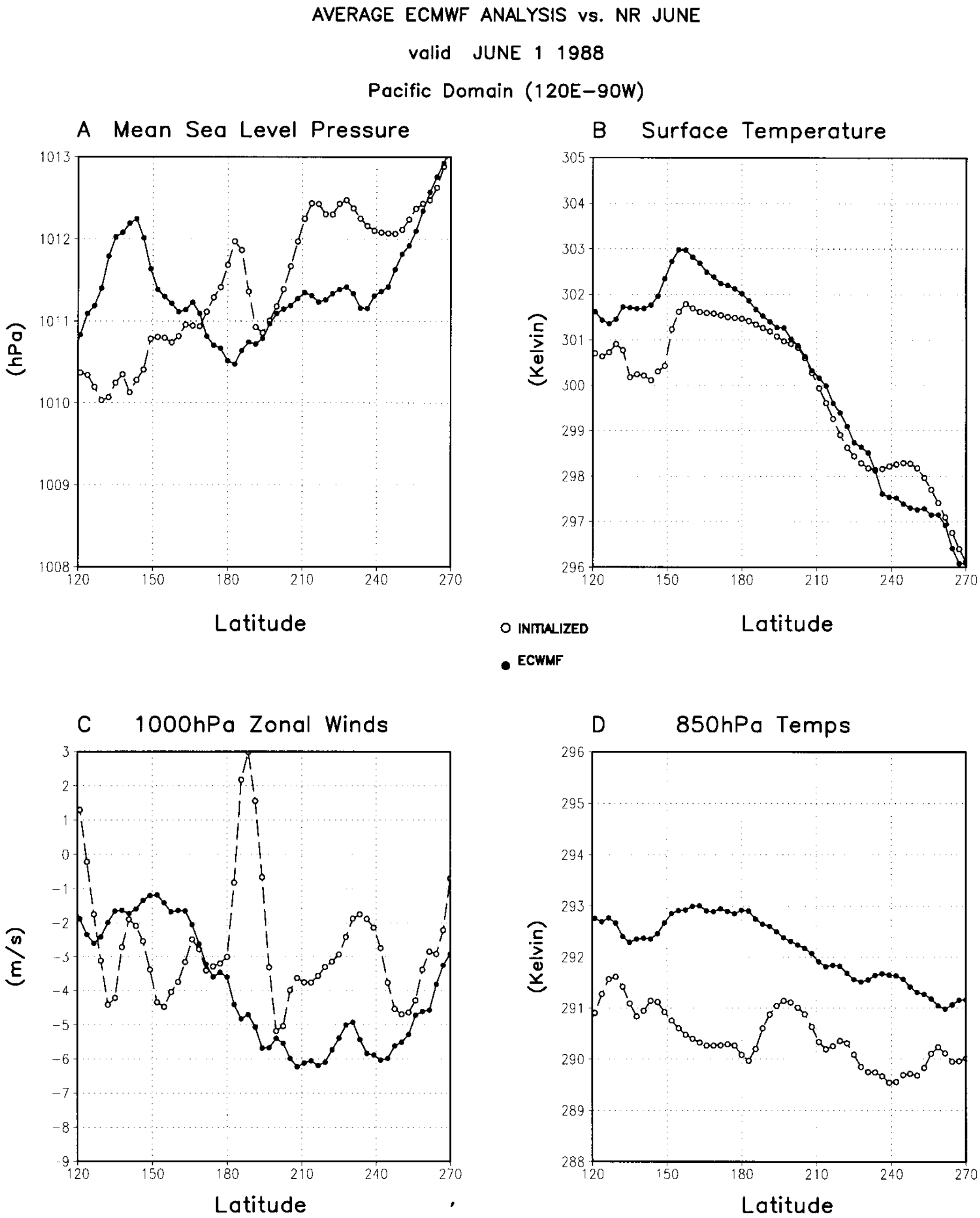

Fig. 4. 1 June 1988 meridional average between 5°S to 5°N across the Pacific Ocean from the initialized model and ensemble. (a) Mean sea level pressure. (b) Surface temperature. (c) 1000 hPa zonal wind. (d) 850 hPa temperatures. Solid line with closed circles are the average from the ensemble and the dashed line with the open circles are the initialized fields.

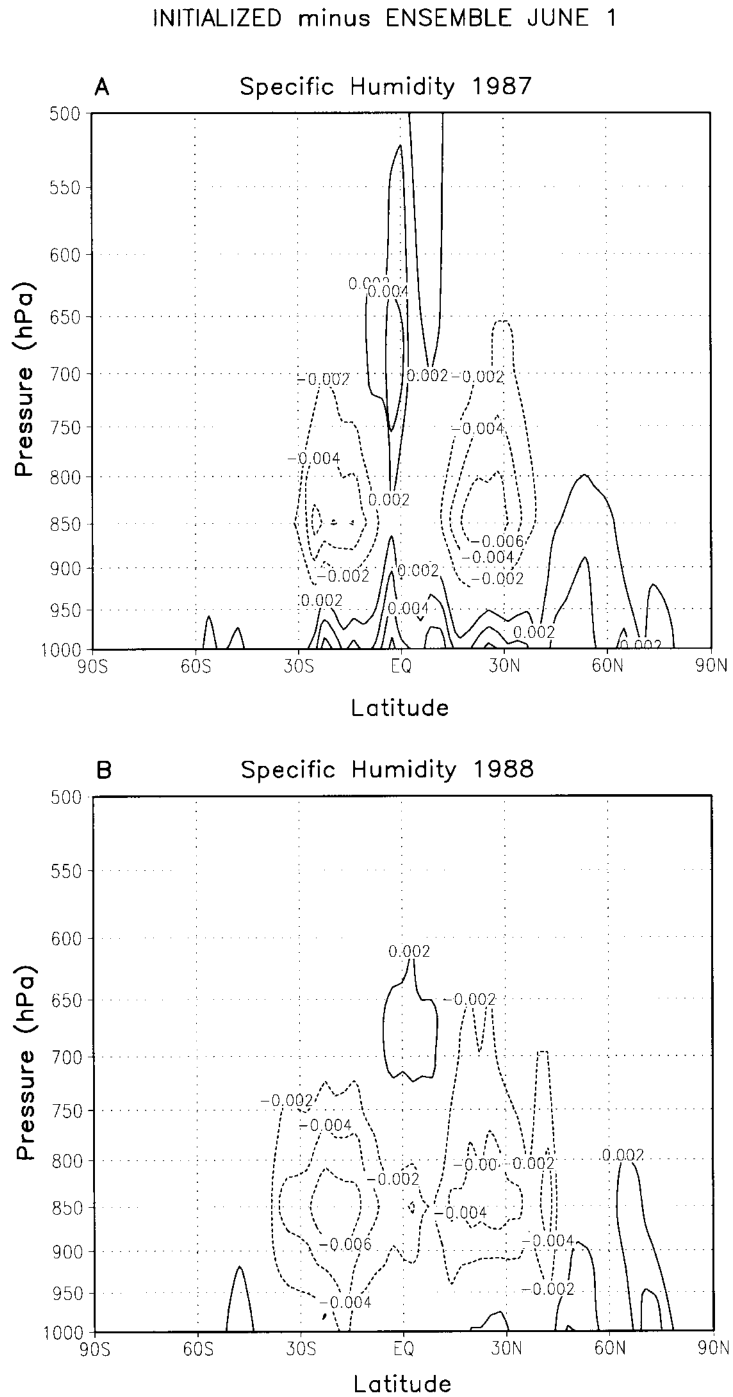

Fig. 5. Zonal average of the specific humidity differences between the initialized 1 June and ensemble mean across the Pacific Ocean (a) 1987, (b) 1988. Contour interval is 0.002 kg kg^{-1}.

end of the initialization. This results from the weak relaxation coefficients chosen and dynamical adjustments occurring between the two models during the initializations. Additionally, the SSTs obtained at the end of the initialization did not match the observed (Reynolds 1988) SST. However, despite the drifts which occurred, the coupled initialization did improve the model's seasonal (June–August) SST forecasts for both 1987 and 1988. The metric used to measure the skill of the forecast SSTs are root mean square (rms) error and anomaly correlations in the Niño 3 (N3) (5°N–5°S, 150°–90°W) and Niño 4 (N4) (5°S–5°N, 160°E–150°W) regions. The rms error and anomaly correlations were calculated using Reynolds (1988) monthly mean observed SSTs.

The 1987 and 1988 monthly mean SST rms error in the N3 region, as a function of forecast month for the initialized forecast, all members of the ensemble, and persistence are shown in Figs. 6a, b. The solid line is the initialized model, dotted lines are the individual members of the ensemble, and the dot-dashed line is persistence. The initial rms error in the N3 region for 1987 and 1988 are 0.61 K and 1.01 K, respectively. These initial SST errors result from the combination of the ocean model's known cold bias, a dynamical adjustment occurring in the ocean to the changing wind stresses, and deficiencies in the model's parameterizations. Most notable of the deficiencies are the weak wind stresses in the equatorial regions. Ji et al. (1994) speculated that the cause of the weak equatorial wind stresses in the NCEP coupled model are a result of the diffuse ITCZ, which prevents concentrated heat sources from generating stronger low level equatorial flow. This could be the case since the coupled models used at NCEP and FSU both use a Kuo cumulus parameterization scheme.

During the 1987 seasonal forecast, the initialized model experienced a slow and steady increase in the rms error to a final August value of 1.34 K; an increase of only 0.73 K during the 90-day forecast. In contrast, all members of the ensemble undergo a rapid increase in the rms error during the 1st month of the forecast. The ensemble's mean error growth during the first month is 1.3 K, and obtains a final mean rms error of 2.1 K; almost twice the error of the initialized model. The rapid error growth observed indicates substantial imbalances in the ensemble's initial conditions.

Meridional averages between 5°N and 5°S across the Pacific for the net heat flux, zonal wind stress, precipitation and wind stress averaged over the first month of the 1987 free forecast are shown in Fig. 7. The ensemble shows higher amounts of precipitation in the eastern Pacific (Fig. 7c) and stronger easterly wind stresses in the east and central Pacific (Fig. 7b). Although not shown, the near equatorial region (near 15°N and 15°S) wind stresses were also stronger in the ensemble by an average 25 mPa near the coast of South America and 75 mPa in the central Pacific near 15°N. Contributing to the colder SSTs in the ensemble during June were the negative net heat flux (Fig. 7a). The initialized model showed positive net heat fluxes across the Pacific basin, with small positive values of 10 W m^{-2} in the western Pacific, increasing to 75 W m^{-2} in the eastern Pacific. Both the initialized and ensemble exhibit similar values in the meridional wind stresses (Fig. 7d), and are close to the observed values given by Stricherz et al. (1992).

The slow linear increase observed with the initialized forecast during 1987 indicates that the one year coupled initialization produced a more consistent and balanced initial state. Additionally, the lack of any large increase in the SST rms error associated with the initialized model also indicates that the SST errors that do develop are more a result of model deficiencies than initial condition imbalances at the ocean-atmosphere interface. The errors associated with the ensemble are almost twice as large as the initialized forecast throughout the period. In the ensemble integration, 87% of the total error growth occurred during the first month of the forecast.

In the N3 region, persistence is seen to have a lower SST rms error than the initialized model at lead times less than one month (Fig. 6a). However, at lead times greater than 2 months, the initialized model's rms error was smaller than persistence, such that by August the initialized model was 0.4 K lower than persistence. In contrast, all members of the ensemble fail to out perform persistence during the entire forecast period. This occurs in spite of the fact that the same ocean initial state was used as in the initialized case.

Similar findings regarding the Niño 3 SST rms error were also observed during the 1988 free

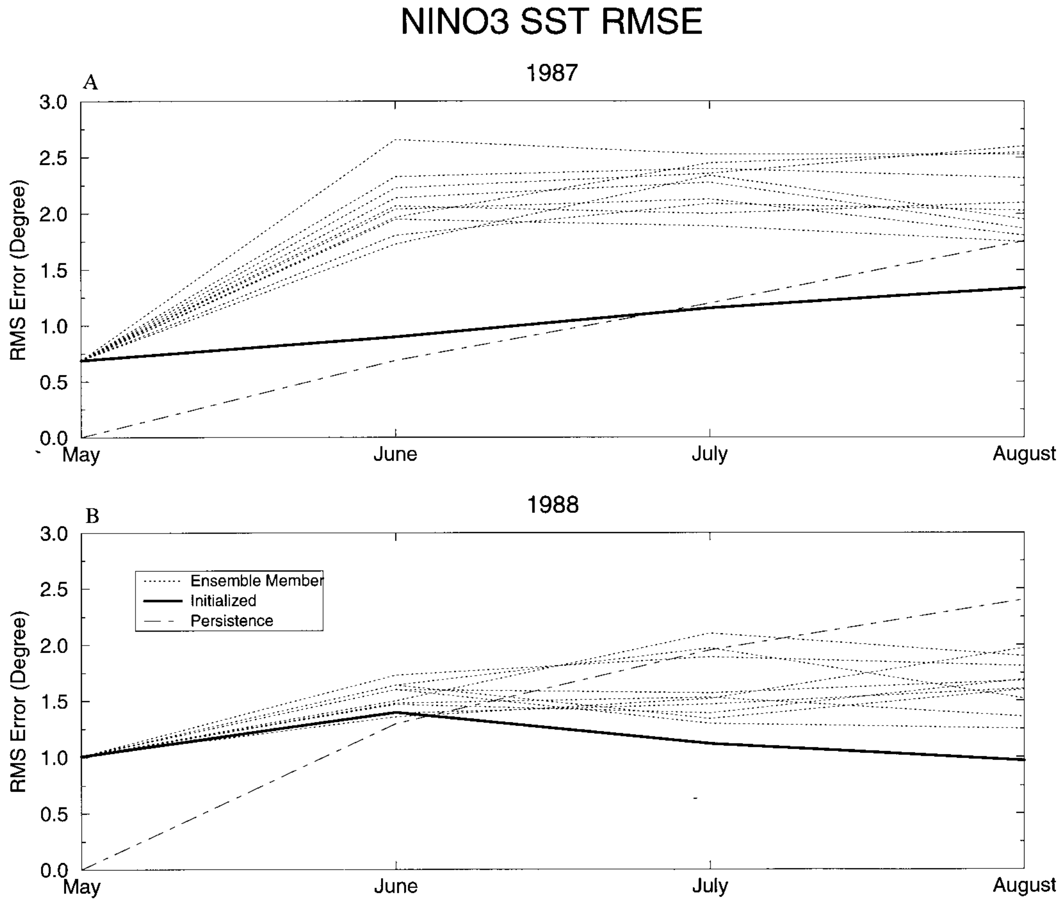

Fig. 6. Sea surface temperature root mean square error in the Niño 3 region as a function of forecast month. (a) 1987, (b) 1988. Solid line is the initialized model, dotted lines are individual model integrations from the ensemble and the dashed-dotted line is persistence.

forecast (Fig. 6b) with the initialized forecast outperforming all members of the ensemble at lead times greater than one month. The spread between the initialized and ensemble members during 1988 was not as large compared to the 1987 forecast (Fig. 6a). Nevertheless the impact of the atmospheric initial conditions are clearly evident in the determination of the Niño 3 SST rms error during both seasonal integrations.

The initialized model also had a smaller SST rms error compared to the ensemble in the western Pacific (Niño 4 region) during the 2–3 month forecasts (Fig. 8). During the 1987 forecast (Fig. 8a), the ensemble's August SST rms error was 1.78°, whereas the initialized forecast had only a 1.06° error. Smaller differences between the ensemble members and the initialized model were found during 1988. In fact it is only during third month that the initialized forecast displayed a smaller rms error. The August difference between the ensemble mean and the initialized forecast was 0.3° (Fig. 8b). Similar to the eastern Pacific, the impact of the atmospheric initial conditions were again very noticeable in the western Pacific during the warm event, with all members of the ensemble being substantially larger than the single realization from the initialized forecast. The impact of the initial conditions is however less obvious during the cold event.

As shown in Figs. 6b and 8b the initialized

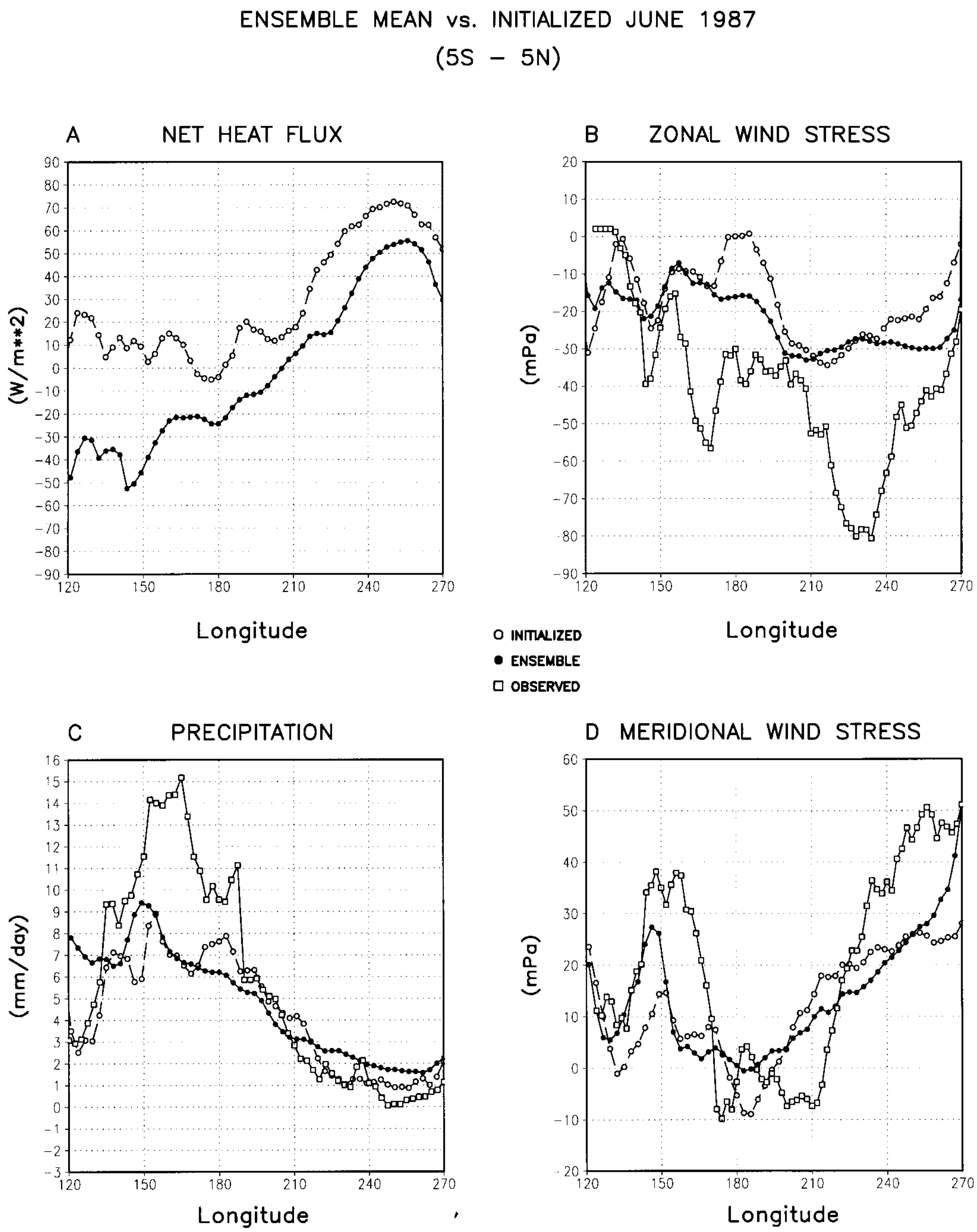

Fig. 7. June 1987 monthly mean meridional averages between 5°S–5°N across the Pacific during the free forecast. (a) Net heat flux. (b) Zonal wind stress. (c) Precipitation. (d) Meridional wind stress. Dashed line with the open circles is the initialized forecast, solid line with the closed circles are the mean of the ensemble and the solid line with the open squares are the observations. In (b) and (d), the observed wind stress is the FSU wind stress analysis, while in (c) the observed precipitation is from the GPC data.

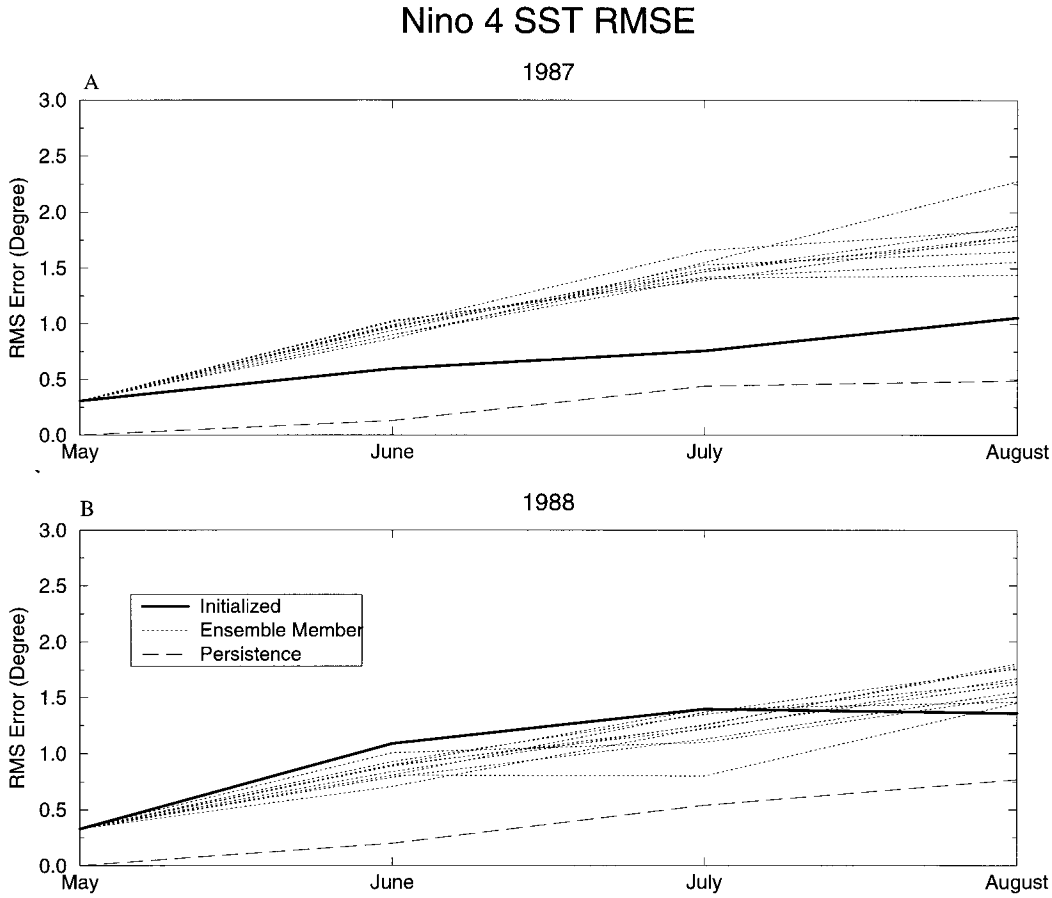

Fig. 8. Sea surface temperature root mean square error in the Niño 4 region as a function of forecast month. (a) 1987 (b) 1988. Solid line is the initialized model, dotted lines are individual model integrations from the ensemble and the dashed-dotted line is persistence.

Niño3/Niño4 SST rms errors were smaller than the ensembles in the equatorial Pacific during the second/third months of the forecast for 1988. This indicates that perhaps the atmospheric initial conditions do not play a significant role in determination of the cold SST anomalies and what might be more important is the ocean's subsurface temperature structure. The ocean's upper heat content was not significantly altered during the 1988 free forecast by the ensemble integrations. In fact, one month prior to the free forecast, the coupled initialization was already beginning to develop the cold event in the eastern Pacific and this information was carried by the coupled model into the free forecast. Additionally, the ocean model's known cold bias coupled with the evolving cold event acted positively on the system to prevent large SST rms errors (with respect to the ensemble and initialized forecast) from occurring.

Fig. 9 shows the meridional averages between 5°N and 5°S across the Pacific for the net heat flux, zonal wind stress, precipitation and meridional wind stress averaged over the first month of the 1988 free forecast. Similar to the 1987 June forecast, the ensemble zonal wind stresses shows almost no variation in the intensity in the eastern Pacific, whereas the initialized model's zonal wind stresses shows increasing easterlies off the coast of South America (Fig. 9a). Although it is not as strong as the FSU zonal wind stress analysis, the

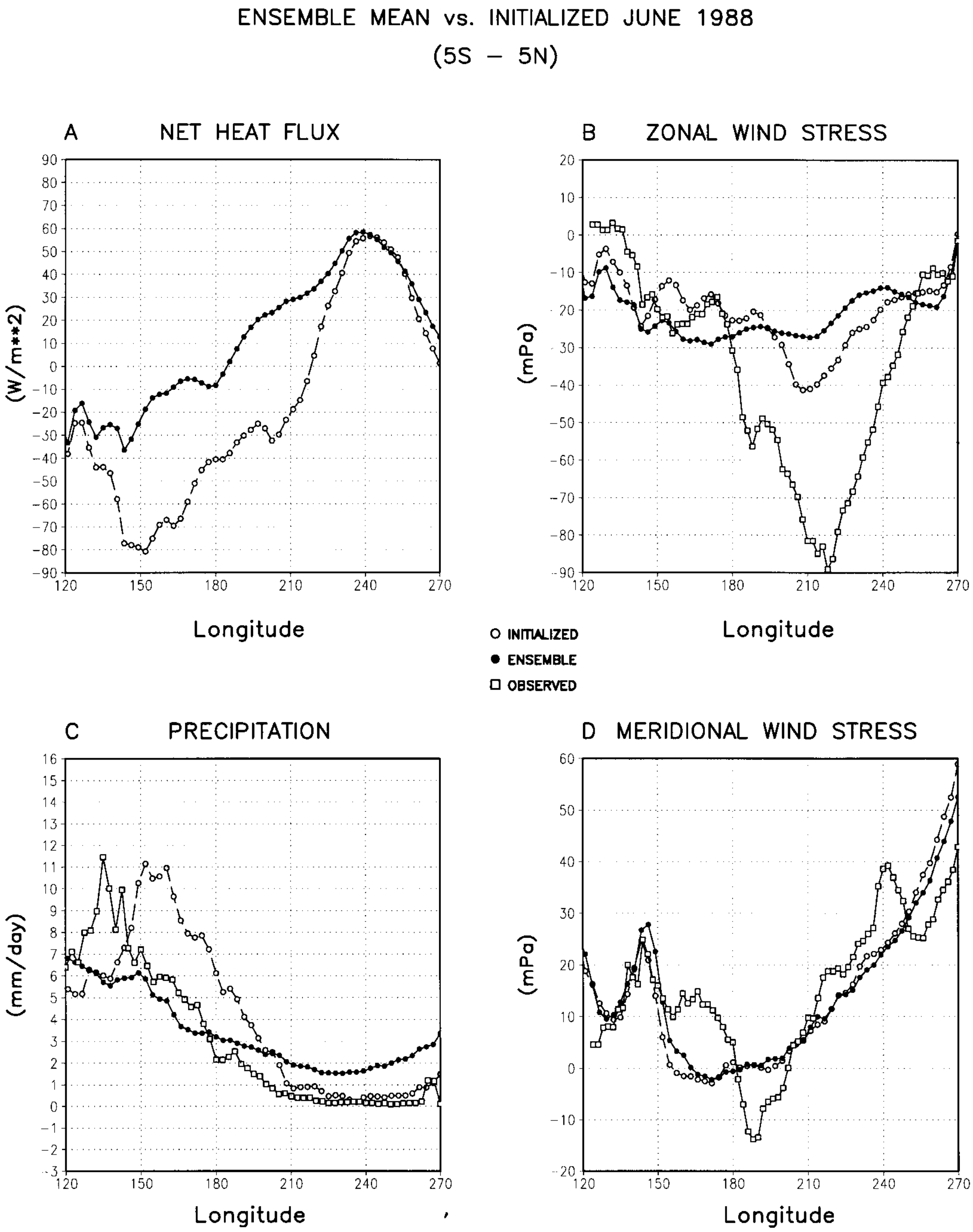

Fig. 9. June 1988 monthly mean meridional averages between 5°S–5°N across the Pacific during the free forecast. (a) Net heat flux. (b) Zonal wind stress. (c) Precipitation. (d) Meridional wind stress. Dashed line with the open circles is the initialized forecast, solid line with the closed circles are the mean of the ensemble and the solid line with the open squares are the observations. In (b) and (d), the observed wind stress is the FSU wind stress analysis, while in (c) the observed precipitation is from the GPCP data.

initialized model's maximum was slightly shifted west of the observed maximum. While the model zonal wind stresses showed large differences with the observed, the initialized and ensemble meridional wind stresses both agreed well with the FSU analysis (Fig. 9d). Similarly, the precipitation from the initialized forecast shows very similar patterns compared to the GPCC (Huffman et al., 1995) rainfall (Fig. 9c). Both have maximum amounts of 11 mm day^{-1} in the western Pacific with sharp decreases towards the east. The ensemble shows no sharp increase towards the west. Coinciding with the increase in the precipitation in the western Pacific is the large negative net heat flux shown by the initialized forecast (Fig. 9a). The initialized model does however exhibit stronger cooling over the Pacific region. This is possibly a continuation from the initialization phase.

Tables 2, 3 show the initialized and ensemble's SST anomaly correlations in the N3 and N4 regions for the two seasonal forecasts. During both seasonal forecasts the initialized model has consistently higher anomaly correlations compared to the ensembles. The poor 1987 ensemble forecast is again displayed by the negative anomaly correlations. Taken together, the SST rms error and the anomaly correlations demonstrate the positive impact the coupled initialization had on the seasonal forecast.

6. Summary

In this paper, we have introduced the notion of a Newtonian relaxation within a coupled ocean-atmosphere initialization scheme. This scheme is similar to that used by Chen et al. (1995); however, it is more inclusive using relaxation on both the atmosphere and the ocean system simultaneously. We found that by initializing the coupled system we could reduce the drifts which can occur with any coupled model at the beginning of the forecast. The idea is to use the scheme to bring both general circulation models into a balanced state (with respect to the individual model's parameterizations) while preventing large coupled model drifts. For this reason, we have refrained from using Newtonian relaxation on any of the parameterizations. This permits the coupled model's parameterization to determine ocean's subsurface temperature, wind stresses and net heat flux. This then translates into increase skill of the SST forecast in the tropical regions.

We selected the values of the relaxation coefficients based on model experimentation. These coefficients are generally an order of magnitude smaller then the values quoted in atmosphere only studies. We found that higher values of the relaxation coefficients, such as those used in atmospheric only studies, produced negative feedbacks onto

Table 2. *1987 and 1988 SST anomaly correlations for the Niño 3 region for the initialized and ensemble forecasts*

	SST anomaly correlations N3 region					
	1987			1988		
	June	July	August	June	July	August
initialized	0.74	0.68	0.78	0.86	0.84	0.70
ensemble	−0.42	−0.55	−0.29	0.84	0.79	0.67

Table 3. *1987 and 1988 SST anomaly correlations for the Niño 4 region for the initialized and ensemble forecasts*

	SST anomaly correlations N4 region					
	1987			1988		
	June	July	August	June	July	August
initialized	0.77	0.74	0.40	0.70	0.79	0.92
ensemble	−0.08	−0.27	−0.13	0.61	0.75	0.89

the ocean model. One possible reason for this is that atmosphere only studies are generally concerned with short range forecasting and therefore they relax the model strongly towards observational values over a very short time period (usually 24 hours). These atmosphere only studies experience only limited shock when the relaxation is turned off, since coupled feedbacks do not take place. For this study, we chose to relax as few variables as possible and as weak as possible thereby giving the coupled system the ability to come into an equilibrium without much drift. Additionally, the relaxation coefficients were reduced linearly 70% during the year long initialization in order to further reduce the shock during the free forecasts.

The higher skill in the Niño 3 and Niño 4 regions, as measured by smaller SST rms error and better spatial structure, in the initialized model is due to the unique coupled initialization procedure. The coupled initialization allowed coupled interactions to occur between the two models during the initialization thus reducing the initial data imbalances at the start of the free forecast by making the model's initial conditions more self-consistent. It was speculated that important to the coupled initialization was the generation of a balanced (with respect to the atmospheric model) moisture field, this was especially important in the tropics and most noticeable during the warm event year. The balance was achieved during the coupled initialization process. It is therefore possible that some of the initial drifts in the SSTs which many coupled models experience, result from a spin-up or spin-down in the tropical moisture field. Perhaps the most striking result of this study is the demonstration that a single realization from the coupled assimilation/initialization outperforms all member of a control ensemble forecast. This skill is largely seen in the SST's over the Niño-3 and Niño-4 regions during the 1987 warm event. The model initialization also showed skill in the Pacific during the cold event of 1988 at lead times greater than two months.

7. Acknowledgments

We would like to thank Dr. Mojib Latif for his valuable suggestions and comments during the early stages of this work and to Dr. Andreas Sterl for providing the basic ocean model used. This work was supported under the following grants: Navy research fellowship grant no. N00014–93–1-G912, NOAA Global Change Studies grant no. NA56GP0013, and NASA grant no. NAG5–1595. Since February 1997 TEL has been supported by COAPS, The Center for Ocean Atmosphere Prediction Studies, with principal support from the Sec. of Navy Grant by the Office of Naval Research.

REFERENCES

Anderson, J. and Gudgel, R. 1997. Impact of atmospheric initial conditions on seasonal predictions with a coupled ocean-atmosphere model. *Proceedings of the 21st Annual Climate diagnostic Workshop*. Department of Commerce/NOAA, 61–62.

Arakawa, A. and Lamb, V. R. 1977. Computational design of the basic dynamical processes for the UCAL general circulation model. *J. Comput. Phys.* **16**, 173–263.

Barnett, T. P., Latif, M., Kirk, E. and Roeckner, E. 1991. On ENSO Physics. *J. Climate* **4**, 487–515.

Businger, J. A., Wyngaard, J. C., Izumi, Y. and Bradly, E. F. 1971. Flux profile relationship in the atmospheric surface layer. *J. Atmos. Sci.* **28**, 181–189.

Chang, L. W. 1978. Determination of surface flux of sensible heat, latent heat and momentum utilizing the bulk Richardson number. *Papers in Meteor. Res.* **1**, 16–24.

Chen, D., Zebiak, S. E., Busalacchi, A. J. and Cane, M. A. 1995. An improved procedure for El Niño forecasting. Implications for predictability. *Science* **269**, 1699–1702.

Daley, R. and Puri, K. 1980. Four-dimensional data assimilation and the slow manifold. *Mon. Wea. Rev.* **108**, 85–99.

Derber, J. D. and Rosati, A. 1989. A global oceanic data assimilation system. *J. Phys. Oceanogr.* **19**, 1333–1347.

Ghil, M. and Malanotte-Rizzoli, P. 1991. Data assimilation in meteorology and oceanography. *Advances in Geophysics* **33**, 141–266.

Hellerman, S. and Rosenstein, M. 1983. Normal monthly wind stress over the world ocean with error estimates. *J. Phys. Oceanogr.* **13**, 1093–1104.

Hoke, J. E. and Anthens, A. 1976. The initialization of numerical models by a dynamic-initialization technique. *Mon. Wea. Rev.* **105**, 1551–1556.

Huffman, G. J., Adler, R. F., Rudolf, B., Schneider, U. and Keehn, P. R. 1995. Global precipitation estimates based on a technique for combining satellite-based estimates, rain-gauge analysis, and NWP model precipitation information. *J. Climate* **8**, 1284–1295.

Ji, M., Kumar, A. and Leetmaa, A. 1994. A multi-season climate forecast system at the National Meteorological Center. *Bull. Amer. Meteorol. Soc.* **75**, 569–577.

Kanamitsu, N. 1975. *On numerical prediction over a global tropical belt*. Report No. 75–1. Available from the Department of Meteorology. Florida State University. Tallahassee, Fl. 32306, 1–282.

Kanamitsu, N., Tada, K., Kudo, K., Sato, T. and Isa, N. 1983. Description of the JMA operational spectral model. *J. Meteor. Soc. Japan* **61**, 812–828.

Krishnamurti, T. N., Low-Nam, S. and Pasch, R. 1983. Cumulus parameterization and rainfall rates (II). *Mon. Wea. Rev.* **111**, 815–828.

Krishnamurti, T. N., Kumar, A., Yap, K. S., Dastoor, A., Davidson, N. and Sheng, J. 1990. Performance of a high resolution mesoscale tropical prediction model. *Advances in Geophysics* **32**, 133–286.

Krishnamurti, T. N., Jishan Xue, Bedi, H. S., Ingles, K. and Oosterhof, D. 1991. Physical initialization for numerical weather prediction over the tropics. *Tellus* **43A/B**, 53–81.

LaRow, T. E. 1997. Seasonal Simulation using a coupled ocean-atmosphere model. PhD Thesis. Florida State University, pp. 191.

Latif, M. 1987. Tropical Ocean circulation experiments. *J. Phys. Oceanogr.* **17**, 246–263.

Latif, M., Sterl, A., Maier-Reimer, E. and Junge, M. M. 1993. Climate variability in a coupled GCM. Part I: The tropical Pacific. *J. Climate* **6**, 5–21.

Latif, M., Barnett, T. P., Cane, M. A., Flugel, M., Graham, N. E., Von Storch, H., Xu, J. S. and Zebiak, S. E. 1994. A review of ENSO prediction studies. *Climate Dynamics* **9**, 167–179.

Latif, M., Stockdale, T., Wolff, J., Burgers, G., Maier-Reimer, E., Junge, M., Arpe, K. and Bengtsson, L. 1994b. Climatology and variability in the ECHO coupled GCM. *Tellus* **46A**, 351–366.

Levitus, S. 1982. *Climatological atlas of the world ocean.* NOAA Professional paper no. 13, 173 pp., 17 microfiche.

Louis, J.-F. 1981. A parametric model of vertical eddy fluxes in the atmosphere. *Boundary Layer Meteorology* **17**, 187–202.

Oberhuber, J. M. 1988. *An atlas based on the COADS data set: the budgets of heat, buoyancy and turbulent kinetic energy at the surface of the global ocean.* Max-Planck-Institute for Meteorology/Hamburg Report No. 15.

Pacanowski, R. and Philander, G. 1981. Parameterization of vertical mixing in numerical models of tropical oceans. *J. Phys. Oceanogr.* **23**, 1443–1451.

Reynolds, R. W. 1988. A real-time global sea surface temperature analysis. *J. Climate* **1**, 75–86.

Ramamurthy, M. K. and Carr, F. H. 1987. Four-dimensional data assimilation in the monsoon region. Part I: Experiments with wind data. *Mon. Wea. Rev.* **115**, 1678–1706.

Rosati, A., Miyakoda, K. and Gudgel, R. 1997. The impact of ocean initial conditions on ENSO forecasting with a coupled model. *Mon. Wea. Rev.* **125**, 754–772.

Stauffer, D. R. and Seaman, N. L. 1990. Use of Four-Dimensional Data Assimilation in a limited area mesoscale model part II. Effects of data assimilation within the planetary boundary layer. *Mon. Wea. Rev.* **119**, 734–754.

Stockdale, T., Burgers, G., Latif, M. and Wolff, J. 1994. Some sensitivities of a coupled ocean-atmosphere GCM. *Tellus* **46A**, 367–380.

Stricherz, J. N., O'Brien, J. J. and Legler, D. M. 1992. *Atlas of Florida State University winds for TOGA* 1966–1985, 256 pp.

Tiedke, M. 1984. The sensitivity of the time-mean large-scale flow to cumulus convection in the ECMWF model. *Workshop on Convection in large-scale numerical model* ECMWF, 297–317.

Yap, K. S. 1995. Impact of a Newtonian assimilation and physical initialization on the initialization and prediction by a tropical mesoscale model. *Mon. Wea. Rev.* **123**, 833–861.

Zebiak, S. E. and Cane, M. A. 1987. A model of El Niño-Southern Oscillation. *Mon. Wea. Rev.* **115**, 2262–2278.