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Sea level assimilation experiments in the tropical pacific

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ABSTRACT

Idealised twin experiments with the HOPE ocean model have been used to study the ability of sea level data assimilation to correct for errors in a model simulation of the tropical Pacific, using the *Cooper and Haines* (1996) method to project the surface height increments below the surface. This work should be seen in the context of the development of the comprehensive real-time ocean analysis system used at ECMWF for seasonal forecasting which currently assimilates only thermal data.

Errors in the model simulation from two sources are studied: those present in the initial state, and those generated by errors in the surface forcing during the simulation. In the former, the assimilation of sea level data improves the convergence of the model towards its twin. Without assimilation convergence occurs more slowly on the equator, compared to an experiment using only correct surface forcing. With forcing errors present the sea level assimilation still significantly reduces the errors almost everywhere. An exception was in the central equatorial Pacific where assimilation of sea level did not correct the errors. This is mainly due to this region responding rapidly to errors in wind stress forcing and also to relatively large fresh water flux errors imposed here. These lead to errors in the mixed layer salinity, which the Cooper and Haines scheme is not designed to correct. It is argued that surface salinity analyses would strongly complement sea level assimilation here.

1. INTRODUCTION

The importance of ENSO (El Nino Southern Oscillation) for variability of the earth's climate on seasonal to inter-annual time scales has lead to the development of many systems to forecast SST anomalies in the tropical Pacific. These vary widely in complexity, ranging from purely statistical methods to fully coupled dynamical models of the ocean/atmosphere system (*Latif et al.* 1998). *Trenberth* (1998) reviews operational and near operational forecasts of the 1997/98 El Nino and concludes that the systems that performed best were those based on coupled dynamical models of the ocean and atmosphere.

One of the main aims of the ECMWF operational seasonal forecasting system is the forecasting of ENSO up to six months lead time (*Stockdale et al.* 1998), based on a fully coupled ocean atmosphere model and an ocean data assimilation system. The ocean assimilation is carried out by solving the OI equations, currently using only in situ temperature profiles. Details are given in *Alves et al.* 1999.

Another potential source of ocean data is from satellite altimeters. Since 1992 satellite sea surface height data have been available from the TOPEX/POSEIDON instruments (hereafter T/P), albeit with some delay. These data are now available in near-real time (*Le Traon et al.* 1998), making it feasible to use such data in the ECMWF real-time ocean analysis system. The purpose of this paper is to investigate methods of assimilating such data and to begin to study the potential benefits.

Over the last decade or so there has been some work investigating methods of assimilating sea level data into ocean models, (see review by *Anderson et al.* 1996). *De Mey and Robinson* (1987) projected information below the surface by developing a statistical relationship between sea level changes and changes in sub-surface quantities. They used the POLYMODE data set for the Gulf Stream region to determine correlation between surface and sub-surface pressure. *Hurlburt* (1986) and *Berry and Marshall* (1989) used twin experiments to investigate the ability of an ocean model to propagate surface current information downwards. *Holland and*

Malanotte-Rizzoli (1989) used a nudging method, again in a twin model environment, to assimilate sea level by updating the surface potential vorticity, which had some success in projecting the information to depth. *Mellor and Ezer* (1991) and *Ezer and Mellor* (1994) directly correlated sea surface height changes with changes in the sub-surface temperature and salinity, for the Gulf Stream region. *Haines* (1991) projected altimeter information in the vertical by conserving sub-surface potential vorticity, so that surface current changes were only brought about by changes in the surface potential vorticity. Later *Cooper and Haines* (1996) and *Drakopoulos et al.* (1997) adapted this scheme, for use in primitive equation models, to preserve water mass properties (temperature and salinity) on isopycnal surfaces. This essentially resulted in a lowering or lifting of the water column, the amount of lowering or lifting being set by the requirement that the bottom pressure remained unchanged. *Oschlies and Willebrand* (1996), on the other hand, correlated sea surface height changes with changes in the sub-surface currents. Changes to the density field were then calculated from these using the thermal wind relation, while also preserving the T/S relation. *Woodgate and Killworth* (1996) investigated whether the decomposition of the vertical structure of oceanic variables into normal modes would be useful for altimeter assimilation. They found extreme sensitivity to the barotropic mode making the decomposition unusable for assimilating altimeter data. All of these schemes were tested in mid-latitude model environments.

Weaver and Anderson (1997) used a four-dimensional method to examine the extent to which a time sequence of altimeter measurements can determine the sub-surface flow in a linear multi-layer model of the tropical Pacific Ocean. Although this model used a 'next-generation' assimilation system, the model thermodynamics was very simple and no account was taken of salinity effects on density. The complications of salinity were also ignored by *Verron et al.* (1999), who used an advanced assimilation system based on the Kalman Filter to assimilate altimeter data. Both these studies used models designed for the upper ocean. *Carton et al.* (1996) used a more conventional GCM to look at the relative contribution of each part of the ocean observing network; TOGA-TAO array, XBTs and TOPEX/POSEIDON altimeter, to their sub-surface ocean analyses. However, they too projected all of the altimeter signal onto the thermal field.

One recent study which did address the more realistic situation of both salinity and thermal contributions to the density field was that of *Vossepol and Behringer* (1999). They used a statistical approach to projecting altimeter data locally onto the thermal and salinity fields. The results were generally positive, though the scheme did not correct salinity below 100m. Our paper reports on a different approach to using altimeter data to correct both temperature and salinity. Neither approach is a complete solution to the problem but both are substantial advances over what is currently done. In a later paper we will report on seasonal forecasts made from analyses into which real data from T/P have been assimilated using the scheme reported on here. These results confirm that the positive impacts of the scheme here described are carried over into seasonal prediction.

Some work has also been done to investigate the potential benefit of sea level data for simulating and predicting ENSO. *Fischer et al.* (1995) used a statistical technique to assimilate pseudo sea level data from an NCEP (National Center for Environmental Prediction) analyses (*Ji and Leetmaa,* 1995, *Ji et al.*, 1995). They produced ENSO forecasts from these analyses and compared them to those produced from actually assimilating the sub-surface temperature data from the NMC analyses. They found that the forecasts from the initial conditions generated with sea level assimilation yielded similar results compared to the forecasts with initial conditions generated with sub-surface temperature assimilation. *Ji et al.* (1999) used T/P data in their analysis cycle to prepare initial conditions for their ENSO forecasts. They used a statistical projection scheme translating all the sea-level difference between model and observation into a thermal correction.

In this paper the *Cooper and Haines* (1996) (hereafter referred to as CH) scheme is used in a twin model set up, based on the ECMWF seasonal forecast system, to investigate the potential benefit of sea level assimilation, in particular in the tropical Pacific. Earlier work by CH addressed mid-latitudes. Section 2 describes the sea level assimilation scheme, the ocean model used and the experimental set up. The first set of experiments and results to investigate the ability of sea level assimilation to correct for errors in the initial state are presented in section 3. In section 4 a further set of experiments deals with the ability of sea level assimilation to correct for errors due to incorrect surface forcing. This is an important application as inaccurately known winds are a serious source of error in the tropical oceans. We will show that assimilation of altimeter data provides a very useful improvement in ocean analyses, largely correcting errors due to wind forcing. Finally section 5 summarizes the main results of this paper.

2. DESCRIPTION OF MODEL AND ASSIMILATION SCHEME

2.1 Ocean Model

The ocean model used is based on HOPE (Hamburg Ocean Primitive Equation model) version 2 (*Latif et al.*, 1994, *Wolff et al.*, 1997). The main differences to this basic version are: the horizontal pressure gradients are calculated at the middle, rather than the bottom, of each model level. This allows sea level gradients to be more consistent with the sub-surface pressure gradient field, and has been found to be important for sea level assimilation. A pseudo ice model is used to constrain the model solution over the polar regions, along with a slightly different topography.

The model is forced at the surface with specified daily fluxes of heat, momentum and fresh water. The solar radiation penetrates below the surface layer with an exponential decrease. Additional relaxation for temperature and salt constrain these fields, mainly at the surface, as described later. The model is global with horizontal discretization on an Arakawa E grid with a variable grid spacing: the zonal resolution is 2.8° while the meridional resolution varies from 0.5° in the equatorial wave region to 2.8° in the mid latitudes. There are 20 vertical levels, 8 of which are in the top 200 m.

2.2 Method of Altimeter assimilation

The method used for assimilating sea surface height is based on the CH scheme. In this method there is a local vertical adjustment of the water column. If the model sea level is too high the model water columns are displaced upwards and some light surface waters are lost and replaced by some denser bottom waters. Similarly if the sea level is too low the water column is lowered to decrease the weight of the water column. The amount of vertical displacement is set uniquely by specifying that the pressure at the ocean floor should not change. The hydrostatic adjustment

$$\rho_0 g \Delta\eta + \int_{-H}^{0} g \Delta\rho \, dz = \Delta p_b$$

may be written; where $\Delta\eta$ is the sea level change and $\Delta\rho$ is the density change profile. The above equation is general with only $\Delta\eta$ assumed to be known. To close the problem $\Delta p_b=0$, and $\Delta\rho(z)$ are defined by a single quantity, Δh, the vertical displacement of the water column. The profiles to be lowered or raised (called the background or first guess) are taken from the model. The increments to the model temperature and salinity profiles at a given level are defined by the amount of lowering or raising of the water column, Δh. A cubic spline is used to extrapolate between model vertical levels during displacement to allow the change in temperature and salinity due to raising or lowering the water column by a few meters to be accurately calculated. The above equation must be integrated using the same finite difference algorithm as the model.

The properties of the above method are: (i) It is local and derived purely dynamically, i.e. is model independent, and it is easy to apply; (ii) It preserves the T/S relationship of the water column; (iii) It preserves the volume of each water mass and hence the stratification on potential density surfaces, except at the top and bottom and cannot therefore initiate convection. And (iv) by not changing the bottom pressure it avoids changing bottom torques and hence reduces interactions with steep topography.

CH applied the full increment to the temperature and salinity fields immediately after assimilation, followed by geostrophic adjustment to the currents, away from the equator. The approach tested here is to add the T/S increments slowly and to allow the model to adjust the velocity field to the slowly varying mass field, which is more compatible with the method we use for assimilating in-situ sub-surface data in our 'real-time' ocean analysis system. The increments are split by dividing them by the number of time steps between successive assimilations, and then a partial increment is added each time step, so that by the next assimilation the full increment has been added. This removes the need to make geostrophic corrections to the currents, which cannot in any case be done near the equator. The advantages of this approach, compared to discrete application of the increments, are discussed in section 3.

In this idealised twin model study we concentrate on the vertical projection of the sea level data onto the temperature and salinity fields. Thus sea level observations from a twin experiment on the model grid are used. The problem of spreading information in the horizontal (for example, when assimilating along track altimeter data) is not considered. Furthermore, the sea level observations are assumed to be perfect and given full weight. Taking account of observation and model background errors and their horizontal covariance patterns is also left to a future study, although these are essential before the CH scheme can be used to produce analyses for seasonal forecasts (see *Segschneider et al.* 2000).

3. CORRECTION FOR INITIAL STATE ERRORS

3.1 Integrations

The first set of experiments is aimed at determining the extent to which assimilation of sea level can correct for errors in the model simulation resulting from errors in the initial state. The model was initially integrated from 1985 to the end of 1993, taking Levitus temperature and salinity as the initial conditions. The surface forcing fields for this period (heat and fresh water fluxes and wind stresses) were taken from the ECMWF re-analyses (ERA) as daily mean fields. In addition to the surface heat fluxes, the surface temperature was strongly relaxed to Reynolds (1988) SST data on a three day time scale and to Levitus surface salinity data on a one month time scale. This effectively constrained the model surface layer temperature to lie close to observed values, and generally within 0.2°C. Since SST analyses (for example, *Reynolds*, 1988) are generally available, the relatively strong SST relaxation effectively represents an SST assimilation, and is used in our real-time ocean analysis system. On the other hand, surface salinity observations are very sparse. It is therefore more appropriate to only weakly constrain the model surface salinity field, again as in our real-time analysis system.

The two years 1992, 1993 were used for testing the assimilation procedure. This first experiment, called the control, provides the observations for the assimilation experiments. The control integration from the beginning of 1992 to the end of 1993 is repeated, but starting with different initial conditions, to simulate errors in the initial state. Erroneous initial conditions for 1 January 1992 were produced by taking the January 1990 conditions from the control and integrating from 1990 to 1992 using different surface forcing, namely heat and fresh water fluxes, and wind stress from the ECMWF operational archives.

A full list of these experiments is given in table 1. Each integration was performed for 2 years starting on 1 January 1992 and was forced at the surface as for the control. Experiment names follow a convention as described in the table.

Even without any sea level assimilation, errors relative to the control decrease with time in the tropics if the correct surface forcing is used. The experiment IM demonstrates this, with convergence of the tropical circulation and water properties over a period of months to a year. This provides a baseline with which to compare experiments which also include sea level data assimilation, which should converge faster towards the control.

Experiments which assimilate the sea surface height taken from the control are called twin experiments. Three were performed: IA1, IA2 and IA3; the differences being in the frequency of assimilation and the method of adding the increments. IA1 only had sea level assimilated once at the very beginning of the integration. The full increments to the temperature and salinity fields were added to the initial state. In this idealised set up full weight was given to the observed sea level (observed sea level is assumed to be perfect). Thus IA1 enabled a study of how much a single assimilation, which fully corrected the sea level, also corrected the model state and its subsequent evolution.

Although the first assimilation in IA1 corrected the sea level, the sea level error relative to the control subsequently grows again, since the assimilation does not fully correct the temperature, salinity or current fields. The benefit of additional assimilations at regular intervals was studied with experiments IA2 and IA3. IA2 had the sea level from the control assimilated every 10 days throughout the integration. The increments were also added discretely, i.e. in one go, at each assimilation time. Experiment IA3 was like IA2 except that the increments were added smoothly during the 10 days following each assimilation time. The total assimilation increment was simply divided by the number of time steps (120) during which the changes were to be added. This smooth addition should help to avoid exciting gravity waves as the model adjusts more slowly to the changes in the density field. Although the model has a free surface the sea level was not explicitly changed because it responds rapidly through the barotropic adjustment to the density gradients. The barotropic mode in HOPE is solved implicitly with the same time step as the baroclinic mode, viz 2 hours. Although the HOPE model is global and the sea level assimilation was applied globally, we will concentrate on the tropical Pacific, as this is the main area of interest for ENSO forecasting.

3.2 Impact of a single assimilation

The initial sea level error between IM and the control is shown in fig 1. Along the equator, errors in sea level were generally less than 2 cm, but just off the equator, at around 5°N and 5°S, they reached 8 cm and were generally positive (with higher sea levels in IM). Between 10°N-20°N and 10°S-20°S the errors were negative with peaks in excess of 10 cm. While there was little error in the sea level gradient along the equator there was a clear error in the meridional gradient of sea level at most longitudes.

Errors in sea level often arise largely from errors in the thermal structure of the upper ocean and hence heat content errors are to a large extent correlated with the sea level errors. One would, therefore, expect that sea level assimilation would have its greatest impact in correcting errors in the integrated heat content. It is more difficult for the assimilation to correct for errors in quantities at specific depths or errors in the depth of the 20°C isotherm, which is an indicator of the depth of the thermocline. Figure 2a shows the errors in the 20°C isotherm depth at the start of experiment IM, i.e. on 1st January 1992. Within 10° of the equator the 20°C isotherm is too deep, by up to 15 m in places, and it is too shallow further away from the equator, consistent with the sea level errors described above. At around 15°S, in the western half of the Pacific, it is over 25 m too shallow.

Figure 2b shows the 20°C isotherm depth errors immediately after assimilation at the beginning of experiment IA1. It can be seen that errors have decreased almost everywhere. For example, within 10° of the equator, errors are generally less than 5 m, except in the far eastern Pacific, compared to values up to 15 m in IM. At around 15°S, in the western Pacific, errors reaching over 30 m in IM are generally reduced to below 15 m with the first assimilation. In the central Pacific at 10°S, however, assimilation has had only a small impact, with errors of around 20 m still remaining. In the sub-tropics, errors in the Northern Hemisphere have been reduced much more than in the Southern Hemisphere. The residual errors arise for more than one reason. Figure 1 shows that at 140°W there are only small sea level errors and therefore no prospect of correcting T or S with altimeter data by this or any other scheme. Further west it is very different with sea level errors ~15 cm.

We now consider the vertical structure of the errors in the west pacific. Since the surface temperature is always strongly relaxed to the observed SST the errors in the mixed layer temperature are very small. Between 10° and 20° latitude in both hemispheres temperatures are too cold in IM in the region of the thermocline (see fig. 3a for a cross section along 165°E). Sea level assimilation (experiment IA1, fig 3b) has had a beneficial impact on most of these errors, although it has not eliminated them entirely.

The impact of assimilation on the salinity can be seen by comparing figs 3c,d.

Since the surface salinity is only weakly relaxed (one month time scale) to the Levitus climatology, errors in the salinity field in IM around 0.2 ppt occur both in the thermocline and the mixed layer. The most striking feature, however, is the salinity error at 100 m at 14°S where the salinity is too high in experiment IM and below, at 150 m, it is too low. Figures 3 e and f show the full temperature and salinity along 165°E. One can see that there is a strong salinity maximum in this region and the errors are associated with it being wrongly located both vertically and horizontally, in experiment IM. After assimilation the salinity error is much reduced but not entirely eliminated. Remaining regions are where T and S errors compensate in the density field. For example, from 8°S to 12°N corrections to both T and S have been modest. Examination of fig. 1 shows that this is a region of low sea level error due to errors in T and S compensating.

3.3 Temporal evolution of errors

Sea level errors along the equator for the first three months of IM are shown in fig. 4a. Errors in the initial state are not dynamically in balance with the ERA wind stress forcing and hence can propagate eastwards along the equator as free Kelvin waves. These can be seen in the sea level error field. The fastest Kelvin waves seen in the error field travel from the central Pacific to the eastern boundary in just over a month. There is some error reduction during this eastward propagation and some errors in the west Pacific do not propagate but gradually decay in situ with time. At the end of the three months some errors remain, of up to 1-2 cm in both the east and west Pacific. Off the equator some of the errors propagate westwards as Rossby waves.

Figure 4b shows the evolution of the sea level error from IA1, where a single sea level assimilation step was performed at the beginning of the model integration. The increments to the density field are not balanced by increments in the velocity field and this leads to the generation of high frequency gravity waves, which are particularly pronounced in the central Pacific. Some of the imbalance projects onto the first Kelvin wave mode and propagates eastwards in a similar manner to fig. 4a but the magnitude of the wave is larger with the sea level errors reaching a peak of over 5 cm compared to only 3 cm for IM. This illustrates a problem of data assimilation along the equator when the model state is not in balance with the wind stress forcing. However after the initial increase, sea level errors decrease much more quickly in experiment IA1 than in IM: at three months the errors are almost everywhere less than 1 cm in IA1.

Experiment IA2, fig. 4c, is the same as IA1 except that sea level assimilation is carried out every 10 days as the model evolves, with the total thermal and salinity increments added immediately. Up until day 11 the evolution is the same as for IA1. The second assimilation on day 11, and subsequent ones, reduce the sea level errors further. In particular, the large Kelvin wave is removed before it reaches the eastern boundary. Each successive assimilation seems to generate its own set of high frequency gravity wave activity, although with decreasing magnitude; after two months amplitudes are less than 1 cm. After two months, the sea level errors along the equator are generally everywhere less than 1 cm, i.e. less than those in IA1.

In experiment IA3 (fig. 4d) the increments are added gradually between assimilations. This significantly reduces the gravity wave activity, and the errors after two months are of a similar magnitude to those in IA2. However the fast Kelvin wave is not removed as rapidly as in IA2 and manages to reach the eastern boundary. Its magnitude is still larger than in experiment IM in which no assimilation was performed. There are two reasons for this. First, while the sea level increments will correct the sea level error, the corresponding temperature and salinity increments produce a density field which is not in balance with the wind stress, and this imbalance projects onto the Kelvin wave mode. This is still true for IA3, where the increments are added slowly. Secondly, the increments in IA3 are added slowly after each assimilation, but the Kelvin wave moves around 30□ in 10 days, so by the time the increment has been fully applied, the wave has propagated away from the area. Applying the increments over a shorter period of time or doing the assimilation more frequently may reduce this problem but may excite further high frequency gravity wave activity.

Some of these features also occur in the 20°C isotherm depth, with wave propagation again prominent. All sea level assimilation experiments (IA1, IA2 and IA3) have similar Kelvin waves, mainly because the characteristics of the Kelvin waves are dependent on the errors in the initial state and hence depend on the balance between the initial state and the wind stress, which is the same for each assimilation. After three months, IA1 isotherm depth errors are generally less than 5 m, slightly less than in IM. Both IA2 and IA3 have significantly reduced propagation of Kelvin waves.

Sea level assimilation, especially when applied repeatedly, rapidly reduces the errors present in the initial state. This is particularly so for the thermal field, and to a lesser extent the salt field. Applying the assimilation increments in one application without making dynamical adjustments leads to the generation of gravity waves. This is reduced by applying the increments slowly so that the velocity field can adjust to the changes in the density field. However spreading the increments over 10 days may be too slow as the fastest Kelvin waves move out of the region before the increments have been fully added. Transient disturbances are most easily excited in the central Pacific, where the pressure and wind stress fields are delicately balanced.

3.4 Long term errors

The experiments described in table 1 were integrated for a period of two years. Most of the equatorial convergence occurs within the first three months, as seen in figs. 4a,b,c,d. After two years the errors with and without assimilation are relatively small compared to those in the initial state. The errors that remain in the longer term are studied in this section by looking at the time-mean errors over the second year of each integration. Figure 5 shows these mean errors in the 20°C isotherm depth, for IM and IA3. Without sea level assimilation (IM, fig 5a) errors in the sub-tropics are considerable, reaching over 15 m. These are greatest in the western Pacific and poleward of about 10°N/S. Within 10° or so of the equator the errors are significantly smaller, less than 3 m in the eastern Pacific though slightly over 3 m in the west.

With sea level assimilation, the 20°C isotherm depth error is reduced almost everywhere. By comparing figs. 5a and 5b one can see that errors within 10° of the equator are almost everywhere less than 3 m in the case of assimilation, compared to values up to 6 m in the no assimilation case. The main exception is in the far-eastern Pacific off the coast of central America (10°N) where assimilation of sea level has slightly increased the error from around 3 m without assimilation to around 6 m with assimilation. Poleward of 10° the errors have been significantly reduced in IA3. For example, between 10°N and 20°N in the western Pacific, errors of up to 9 m in IM are reduced to less than 6 m for IA3. Similarly, between 10°S and 20°S errors of up to 15 m are reduced to less than 12 m. The integration with only one assimilation, IA1, had lower errors than IM, although generally 1-3 m greater than when assimilation was carried out every 10 days.

In the longer term, assimilation has had a smaller impact in the southern hemisphere sub-tropics than in the northern hemisphere sub-tropics, similar to the very first assimilation, discussed in section 3b. The reason is again related to the salinity structure in these regions. The salt maximum at around 12°S near the date line, has salinity values of up to 36.2 ppt at a depth of 125 m. This means that there are strong vertical and horizontal gradients in both salinity and the S(T) relationship. It is extremely hard for sea level data alone to correct T and S errors under this circumstance because it is easy to have compensating errors in T and S which then disappear from the sea level signal. Despite this problem there is a small improvement in fig. 5b over fig. 5a even in the southern subtropics.

Figures 6a,b show sections along the equator of mean temperature error for IM (fig. 6a) and IA3 (fig. 6b) averaged over the second year of integration (1993). Without assimilation, IM, errors in the thermocline region in the central Pacific are generally less than 0.1°C, but reach between 0.1-0.2°C at depths of around 250 m. With sea level assimilation, IA3, errors at both these depths increase: to about 0.1-0.2°C around 100 m and to over 0.2°C around 250 m. The T errors in IA3 change sign with depth and therefore partly compensate in their contribution to the sea level error.

The errors in the yearly-mean salinity along the equator for IM and IA3 are shown in figs. 6c and 6d, respectively. Assimilation reduces errors in the salinity around 100 m depth in the central and western Pacific: for example, the errors at 100 m have been largely removed. However, like the temperature field, errors at depths of around 250 m have increased slightly with sea level assimilation and this salinity error provides additional compensation for the temperature errors.

The mean temperature and salinity structures for the control integration during 1993 are shown in figs. 6e and 6f. The temperature decreases monotonically with depth, the thermocline being at a depth of around 150 m in the west Pacific and 50 m in the far east Pacific. The salinity depth dependence is not monotonic with a salt maximum at a depth of about 125 m which weakens as it extends eastwards. In the far east there are two maxima, one at 75 m depth and one at around 225 m. In the central and western Pacific the initial sea level error is dominated by the positive salt errors at around 100 m and 250 m in fig 6c. Since the water column in too salty and dense, the CH scheme will lower the water column. At around 100 m depth salt is increasing with depth (see fig. 6f) so lowering the water column reduces the salt error, as seen in fig 6d. But around 200-300 m salinity is decreasing with depth so lowering of the water column increases both the salt and temperature errors. The problem can not be fixed in the current method if only sea level data are available. Sea level assimilation would be more useful if combined with independent data, for example, in-situ temperature profiles, which may help to distinguish the impact of temperature and salinity on the density structure.

4. CORRECTION FOR ERRORS IN THE SURFACE FORCING

4.1 Experiment set up

One of the major causes of error in a forced ocean model simulation, particularly in the tropics, is the forcing itself. In preparing initial conditions for ENSO forecasting, an ocean model is usually integrated using an observed estimate of the surface forcing regardless of whether or not data are being assimilated. The tropical Pacific is particularly sensitive to wind stress forcing. In this section model errors due to surface forcing (in heat fluxes, fresh water fluxes and wind stress) are introduced by using a different set of forcing fields to the control integration. The aim of the sea level assimilation is then to correct the simulation towards the control. Unlike experiment IM of the previous section, a twin experiment without assimilation will not converge to the control with time, as the surface forcing is always different from the control. This is a realistic case as in general there are always forcing errors, which assimilation would help to correct. The 1992-93 period was again considered, and the experiments are set out in table 2 (includes a description of the experiment naming convention).

The control experiment is the same as that described in section 2. In FM, errors due to the forcing were simulated for the period Jan '92 to Dec '93 using surface fluxes from the ECMWF operational archives (OPS). All the surface forcing fields, wind stress, heat flux and fresh water flux, were different from the control. The initial conditions were taken from the control run, i.e. no initial errors. The surface temperature was strongly relaxed to the observed SST as in the control. The control run was defined as truth and the differences in FM are errors. Sea level data from the control were assimilated into the FM run to form the FA run. The sea level assimilation was carried out every 10 days and the increments were split and added every time step as in IA3.

In addition, both experiments were repeated starting from erroneous initial conditions generated from a model run from 1990 using OPS fluxes (see section 3), until January 1992. These experiments, denoted FMI and FAI, allow an assessment of the relative importance of initial state errors versus forcing errors, and the ability of sea level assimilation to correct for them. In both cases the errors are realistic in that they are typical of those that would arise using different versions of the atmospheric forcing fields from the operational analysis/weather-forecasting system.

4.2 Correction for surface forcing errors

Without data assimilation the FM run shows errors in the surface height which grow for the first six months or so and then saturate. The OPS forcing produces a sea level in the west Pacific which is a few centimetres higher than the control using ERA forcing. In the central to east Pacific the error has a seasonal cycle with the FM sea level being up to 5 cm lower than the control in the DJF period, but slightly higher than the control in the JJA period. In the far-east the differences are similar in sign to those in the central Pacific, but around the JJA period, FM is up to 5 cm higher than the control.

Assimilation of sea level data does not totally constrain the sea level because the increments are added gradually on each time step. Between each assimilation time the model uses the incorrect OPS surface forcing so that the sea level slope will never be in balance with the wind forcing. Nevertheless the errors in the surface height are well constrained in the west and east Pacific, but not in the central Pacific. In the west the errors reduce from around 3-5 cm to generally less than 1 cm. In the east they reduce from around 3-5 cm to generally less than 2 cm (not shown). In the central Pacific the reduction in the surface height errors is much less. The ocean circulation in this area depends very strongly on the wind stress forcing since the strength of the upwelling and meridional circulation are driven by the wind forcing. Furthermore, imbalances between the pressure and the wind stress propagate rapidly eastwards as Kelvin waves, as seen in the previous section. Adding the assimilation increments slowly means that the error has propagated eastwards before the increments to correct it are fully applied.

These sea level errors are mirrored by errors in the slope of the 20□C isotherm. Assimilation reduces the errors in the depth of the 20°C isotherm along the equator, as shown in figs. 7a and 7b. Without data assimilation (FM, fig. 7a), errors grow for the first six months or so. After this there is a repeating pattern with a strong seasonal cycle in the east Pacific, with the thermocline in the west Pacific up to 10-15 m deeper than when using the ERA forcing. In the east Pacific, the thermocline is around 5-15 m too deep in JJA but 10-20 m too shallow in DJF, consistent with the sea level differences noted above.

Assimilating sea level (run FA, fig. 7b) reduces the errors in the thermocline depth. In the west Pacific errors of up to 15 m are reduced to less than 5 m, while in the east errors are reduced to less than 10 m. However, in the central Pacific there are periods, for example, month 16, when the error in the 20°C isotherm depth increases from 5 m to around 10 m. This occurs concurrently with the larger sea level errors described above.

Off the equator data assimilation has had an even greater impact on the thermocline depth. Figures 8a and 8b show the errors in the 20°C isotherm depth as an annual mean for the second year of the integration for FM and FA respectively. The errors in the isotherm depth have been significantly reduced by assimilation. For example, errors in the subtropics of over 15 m in FM are in general reduced to less than 3 m in FA. Only in the equatorial band of the central/west Pacific, and in the South Pacific Convergence Zone, do significant errors of 3-6 m remain.

Figures 9a and 9b show the 1993 annual mean errors in the salinity at 50 m depth for FM and FA respectively. Again, except for the equatorial Pacific near the dateline, sea level assimilation leads to a decrease in the error almost everywhere. For example, at around 10°N in the east Pacific, errors of around 0.6 ppt are reduced to less than 0.2 ppt with assimilation. By contrast, in the central equatorial Pacific the error has increased from around 0.2 ppt to 0.4 ppt. This error is found to be mainly confined to the mixed layer, extending down to around 100 m depth. This is a problem that the assimilation scheme does not correct. If the mixed layer is too salty, then the water column is too dense and the assimilation scheme will lower the profiles but leave the mixed layer salinity the same. The availability of sea surface salinity analyses would eliminate most of this problem.

4.3 Correction for surface forcing errors and initial state errors

Errors arising from using both the wrong surface forcing, and an erroneous initial state (the most likely situation) are examined by performing twin integrations with both these errors present. Experiment FMI is the same as FM except that it also has wrong initial conditions for January 1992, i.e. the same initial conditions as for IM described in section 3. An assimilation integration was also performed, FAI, which is the counterpart of FA.

In the tropics (within 20° of the equator) errors in IM are significantly less than those in FM and FMI (for example, when looking at the annual mean of the 20□C isotherm depth error for the second year of integration; fig. 5a for IM, fig. 8a for FM (and FMI, not shown), indicating that errors after a year due to using a different forcing are much greater than those arising from starting with the wrong initial conditions with the correct forcing. Errors in IM are up to 2 m in the equatorial west Pacific, while in FM they reach over 10 m. In the subtropics the errors attributable to the forcing are comparable to those produced from the wrong initial state. Not surprisingly the largest errors are produced when using both wrong forcing and wrong initial state (FMI). It appears that errors from the initial conditions (IM) and errors due to the wind forcing (FM) are more or less additive.

When sea level assimilation is carried out, the errors in the 20°C isotherm depth in FAI are again a combination of those in IA3 and FA. In the equatorial strip (within 10° or so of the equator) errors are mainly due to using different surface forcing, fig. 8b, while in the sub-tropics (poleward of 20°N/S) they are dominated by errors in the initial state, fig. 5b. Between these latitudes both initial state errors and surface forcing errors contribute significantly to the total error in the isotherm depth over the second year of integration.

4.4 Correction for wind forcing errors only

The experiments described in section 4.3 consider the ability of sea level assimilation to correct for errors in the surface forcing. These errors arise from the wind stresses, and the heat and fresh water fluxes. Since the surface temperature in all integrations is strongly relaxed to the observed SST, the heat flux differences will have little impact as the relaxation will compensate for differences in the imposed fluxes. But surface salinity is only relaxed weakly on a one month time scale to the Levitus climatology, and so fresh water flux differences between operational, OPS, and ERA forcing may introduce significant differences in the surface salinity fields. To investigate this, an experiment similar to FM but using the same heat and fresh water fluxes as the control, was performed, denoted FMW, see table 2. Similarly, an assimilation experiment, denoted FAW, was carried out using the same wind stress as FA, but with the heat and fresh water fluxes from the control.

The resulting annual mean errors in the second year, 1993, in the 20°C isotherm depth for FMW are similar to those for FM, fig 8a. The difference is generally less than 3 m, the only exception being small parts of the south eastern tropical Pacific. This shows that most of the error in the 20°C isotherm depth came from the surface wind stress. When sea level was assimilated the 20°C isotherm depth errors in FAW were dramatically reduced, and over much of the tropical Pacific they were less than 3 m. For FA the errors were in general 1-3 m higher than those for FAW. The sea level assimilation therefore corrects most of the error resulting from wind forcing, and some due to flux forcing errors.

Figure 10 shows the 1993 annual mean errors in the salinity field at 50 m depth for FMW (fig. 10a) and FAW (fig. 10b). These can be compared to the corresponding plots when all the forcing was erroneous: FM in fig. 9a and FA in fig. 9b. Using the same fresh water and heat fluxes, leads to a reduction in the salinity field errors (FMW compared to FM) by as much as a factor of two, although similar patterns are seen in both. For example, in the equatorial Pacific around the date line, errors of 0.2 ppt are reduced to around 0.1 ppt. One feature that has not been reduced is the large error of up to 0.7 ppt. centred around 110°W and 10°N. It appears that wind error is the major cause of salinity error at this location.

When sea level from the control is assimilated in FAW, a significant reduction in the salt error at 50m depth results, fig. 10b. Only in the equatorial Pacific around the date line, and the far eastern Pacific, do errors of over 0.1 ppt. remain. For FA, which used erroneous heat and fresh water fluxes as well as stresses, there are errors exceeding 0.2 ppt. away from the equator (fig. 9b). The error that remains in FAW, on the equator near the date line, is larger when sea level is assimilated and is associated with the mixed layer being too salty.

Sea level assimilation has been very effective in controlling the growth of model errors due to errors in the forcing fields. It has had a significant impact on the 20°C isotherm depth, particularly so in the tropics. Away from the central equatorial Pacific, errors in the salinity field have also been reduced. Errors that arise from using the wrong surface forcing lead mainly to temperature and salt errors in the thermocline, which the CH scheme corrects rather well. However, errors affecting the mixed layer, mainly but not exclusively from erroneous heat and fresh water fluxes, are not well corrected by the CH scheme. Because of the large precipitation in the central Pacific, the mixed layer salinity is sensitive to both the fresh water flux and the wind stress forcing. Errors in these, lead to errors in the mixed layer salinity, which are not corrected. Mixed layer salinity observations would be of great benefit in this area.

5. SUMMARY AND DISCUSSION

Idealised twin experiments were used to assess the potential benefit of assimilating altimeter data to correct model errors in the tropical Pacific ocean. The ability of sea level assimilation to correct for two types of errors were studied: (a) model errors due to starting from an erroneous initial state and (b) errors arising from the forcing.

Errors in the model initial state were significantly reduced after just one sea level assimilation, and repeating the assimilation every 10 days reduced the errors in the model evolution still further. The method of applying the increments was also important. Applying the increments in full without making balancing velocity adjustments (difficult near the equator) led to the generation of gravity waves. This problem was reduced by adding increments slowly between each assimilation, however spreading the increments over a 10 day period, at least in the central Pacific where Kelvin waves travel relatively fast, meant that the signal being corrected propagates away before the increment is fully applied. Performing the assimilation more frequently, or adding increments over a shorter period may reduce this problem.

The performance of the CH scheme for altimeter assimilation varies with the salinity structure, particularly where the sea level errors were dominated by the salinity field. The scheme did well where the error was dominated by temperature errors. Where there were strong vertical salt gradients, particularly where these compensate temperature gradients to reduce the density gradients, for example where there are intermediate salt maxima, altimeter assimilation with the CH scheme occasionally made the T/S errors worse. Intermediate salt maxima are found around the thermocline region in the sub-tropical south Pacific and equatorial west Pacific, and the scheme performed less well in these regions. The scheme had its biggest impact in the sub-tropical north Pacific. The CH scheme assumes the conservation of water mass properties and thus cannot correct errors in the vertical temperature/salinity water mass characteristics in any single assimilation.

Sea level assimilation was shown to effectively correct for errors due to the surface forcing, particularly those due to the winds. With sea level assimilation, errors in the 20°C isotherm depth due to using wrong forcing were in general less than half those when there was no assimilation, particularly away from the equator. However just in the central equatorial Pacific the error in sea level was not corrected. The possible reasons are two fold: first, this area is simply too sensitive to the wind stress forcing; and second, by only assimilating every 10 days and adding the increments slowly over this period, the features being corrected have moved at the relatively fast Kelvin wave speed before the increment is fully applied.

The central equatorial Pacific is also a region where there are relatively large differences in the fresh water flux between the two sets of forcing fields used. This, as well as using different wind stresses, leads to differences in the salinity field in the mixed layer. The CH scheme is not able correct for such errors since the principle of water property conservation is least appropriate in the mixed layer. Constraining the surface temperature and salinity with data would be of great benefit when assimilating altimeter data. In the experiments described here the surface temperature was strongly relaxed to the observed values but the surface salinity was only weakly relaxed to climatology as there are no analyses of surface salinity available. Observations of surface salinity would also be of great benefit for sea level assimilation.

The experiments which combined both initial state and forcing errors showed that these errors were more or less additive. Forcing errors dominated in the equatorial strip (within 20° of equator) and initial state errors dominated further away from the equator.

In this idealised twin model study we have concentrated on the vertical projection of the sea level data onto the temperature and salinity fields. Thus sea level observations from a twin experiment were used on the full model grid. The problems of spreading information in the horizontal (for example, when assimilating along track altimeter data) and taking account of observation and model background errors and their horizontal covariance patterns are ongoing research (see *Segschneider et al,* 2000 for preliminary results).

This study has shown that there is potential benefit from assimilating altimeter data in the tropical Pacific and this is best done in combination with a strong constraint to the observed SST. The benefits would increase if enough surface salinity observations were also available. Combining the sea level data with sub-surface in situ data has not yet been considered and may provide even greater benefit from assimilating altimeter sea level data. This is an area of ongoing investigation.

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Experiment	Initial State I January 1992	Assimilation frequency	Assimilation increment
Control	Control	None	None
IM	OPS forced run	None	None
IA1	OPS forced run	Day 1 only	Instantaneous
IA2	OPS forced run	Every 10 days	Instantaneous
IA3	OPS forced run	Every 10 days	Split each step

Table 1. Experiments to investigate the ability of sea level assimilation to correct for initial state errors. Experiment names follow a convention. The first letter refers to the type of error in the twin, with 'I' for initial state error. The second letter indicates whether the twin is a model only 'M' or an assimilation 'A'. An optional third character shows either '1', '2' or '3' indicating different ways of assimilating increments. OPS forced run is an integration using ECMWF operational forcing up until 1992 (as opposed to ERA forcing used in the control) to provide a different initial state.

Experiment	Initial State I January 1992	Assimilation frequency	Assimilation increment	Forcing
Control	Control	None	None	ERA
FM	Control	None	None	OPS
FA	Control	every 10 days	Split each time step	OPS
FMI	OPS forced run	None	None	OPS
FAI	OPS forced run	every 10 days	Split each time step	OPS
FMW	OPS forced run	None	None	OPS Wind stress ERA heat and fresh water
FAW	Control	Every 10 days	Split each time step	OPS Wind stress ERA heat and fresh water

Table 2. Experiments to investigate the ability of sea level assimilation to correct for errors in the surface forcing. ERA is the ECMWF Re-Analyses forcing and OPS is the ECMWF operational forcing. Experiment names follow a convention. The first letter refers to the type of error in the twin and is 'F' for forcing error. The second letter indicates whether the twin is a model only 'M', or an assimilation, 'A'. An optional third character shows either 'I' indicating both wrong initial conditions and erroneous forcing, or 'W' indicating erroneous wind stress forcing only, with heat and freshwater fluxes correct.

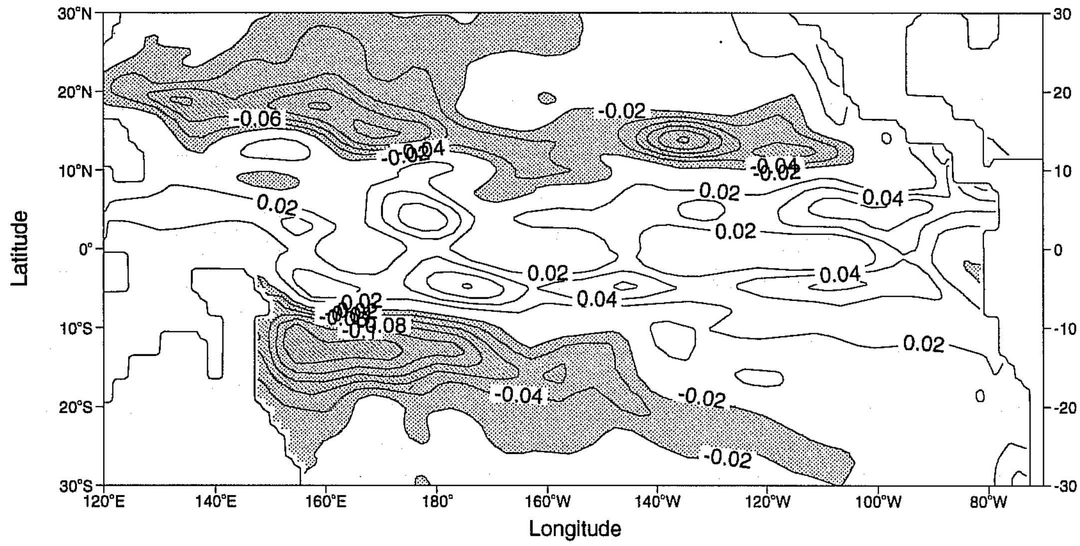

Figure 1. Plot of sea level difference between experiments IM and control at the start of the experiment (Jan 1992). The contour interval is 2 cm, but with the zero contour suppressed, areas less than –2 cm shaded. In the equatorial zone errors are small. However a few degrees off the equator they are ~8cm and between 10° and 20° from the equator are up to 15 cm.

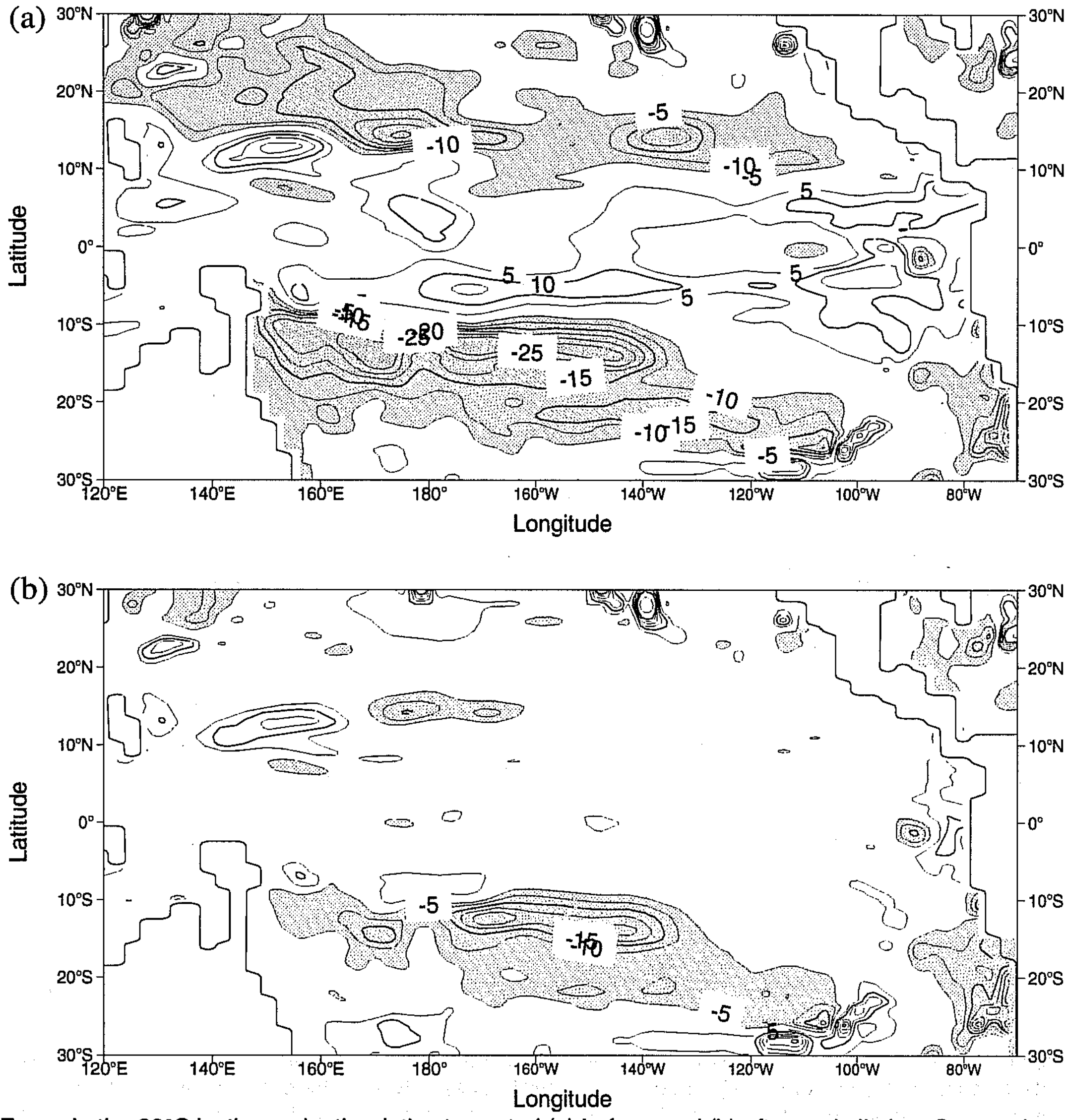

Figure 2. Errors in the 20°C isotherm depth relative to control (a) before and (b) after assimilation. Contour interval is 5 m, with the zero contour not plotted. Shading indicates areas less than -5 m.

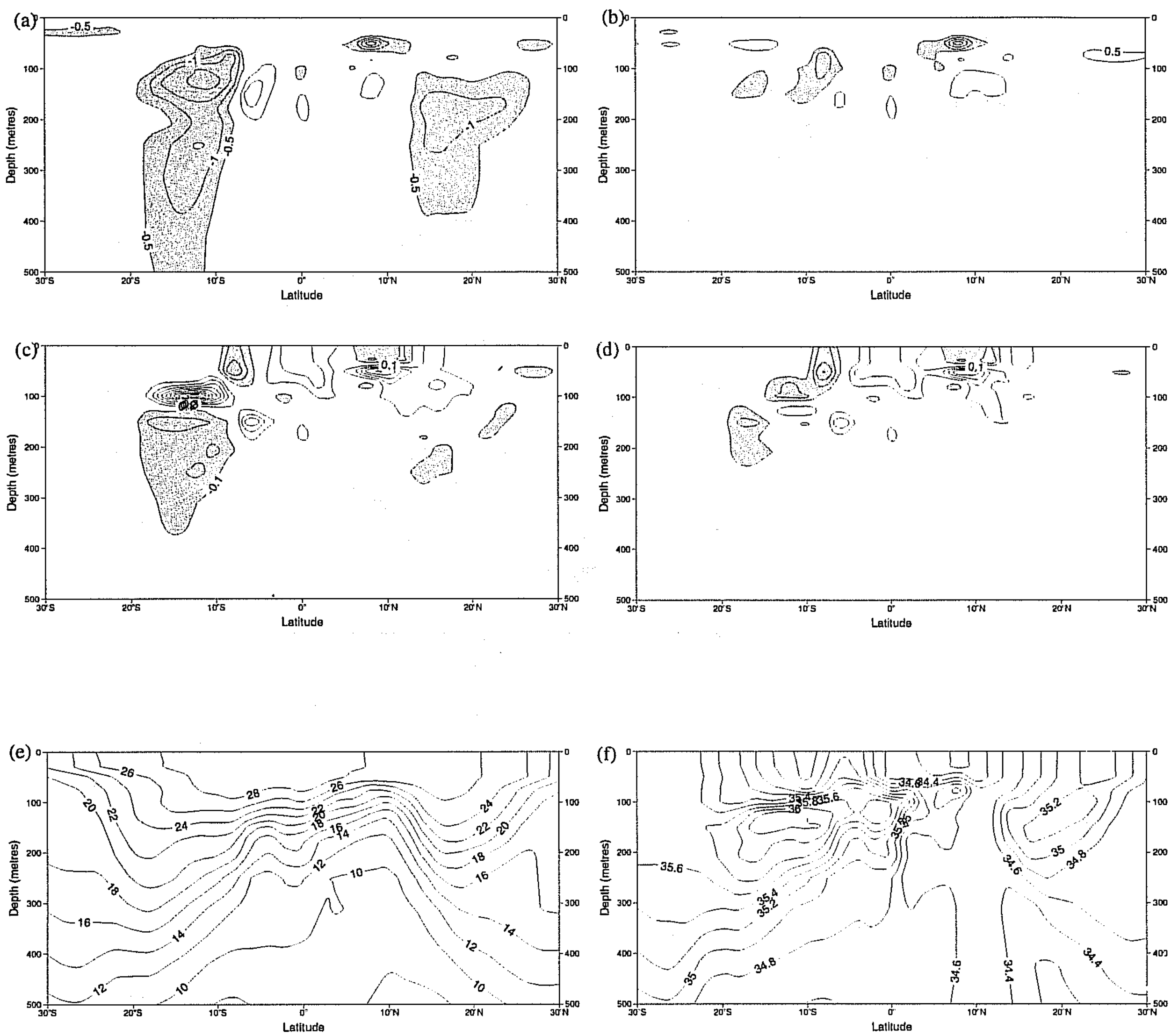

Figure 3. Meridional sections at 165°E before and after assimilation: (a) temperature error before assimilation, IM-Control, (b) temperature error after assimilation IA1-Control, (c) salinity error before assimilation, IM-Control, (d) salinity error after assimilation, IA1-Control, (e) Control temperature and (f) Control salinity. Temperature error contours are every 0.5°C, with values less than -0.5°C shaded and no zero contour. Salinity error contours are every 0.1 ppt, with values less than -0.1 ppt shaded and no zero contour. Temperature contours are every 2°C and salinity contours are every 0.2 ppt.

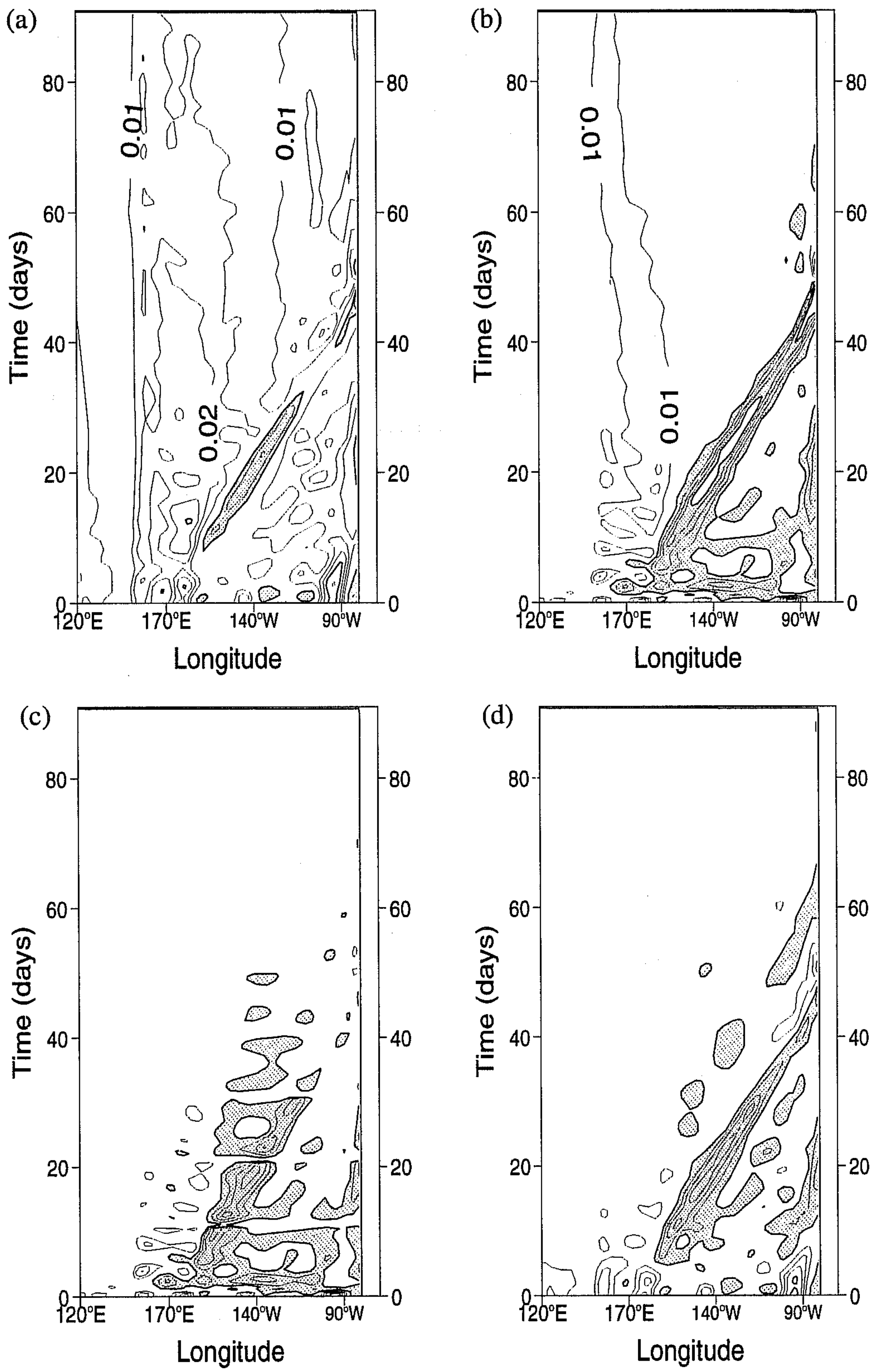

Figure 4. Time evolution of sea level errors relative to control for the first three months of integration: (a) IM (b) IA1 (c) IA2 (d) IA3. Contour interval is 0.01 m, with the zero contour not plotted. Shading indicates areas less than -0.01 m.

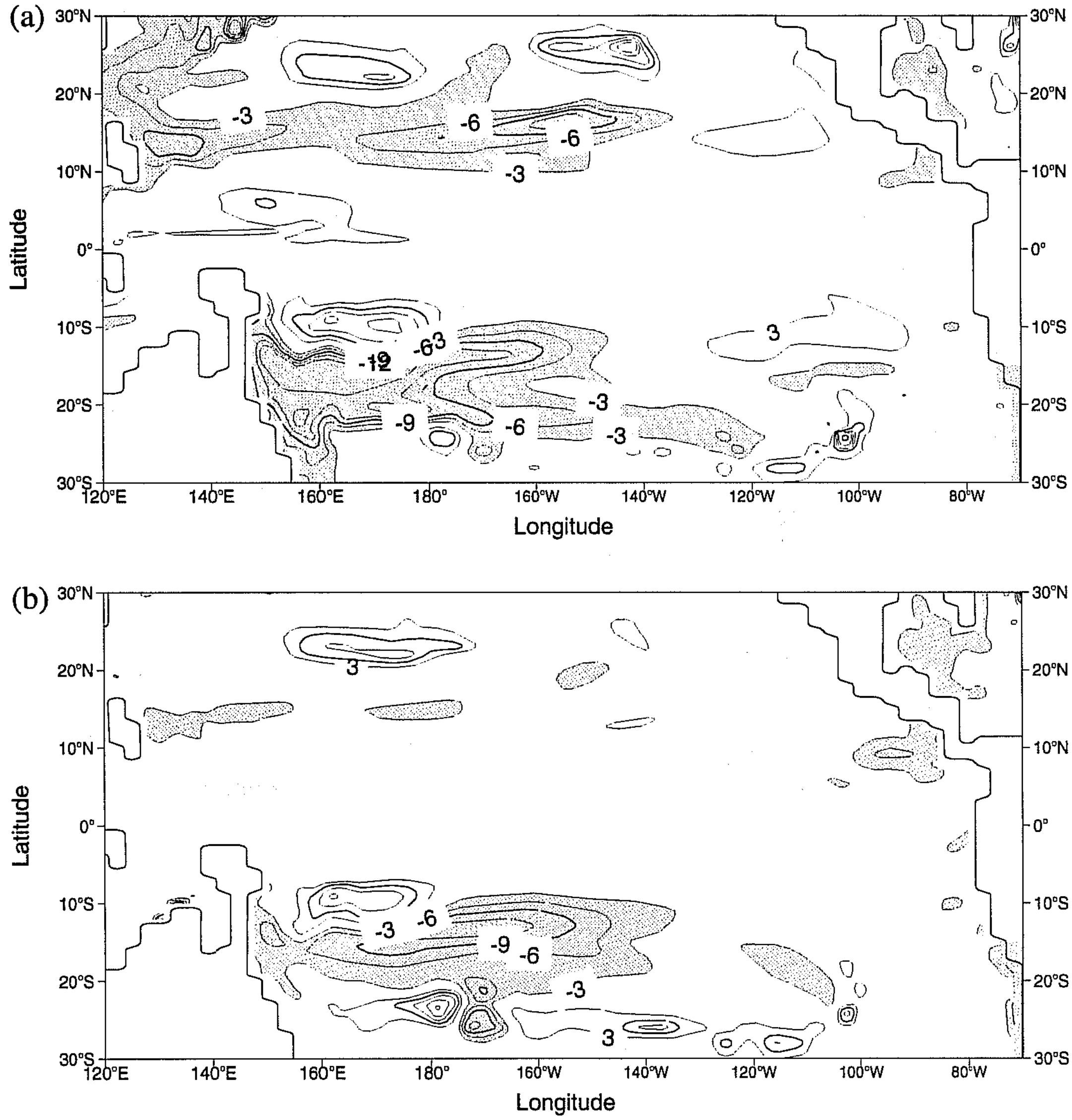

Figure 5. Errors in the 20°C isotherm depth for the second year of integration (mean over 1993) relative to the control: (a) IM and (b) IA3. Contour interval is 3 m, with the zero contour not plotted. Shading indicates areas less than −3m.

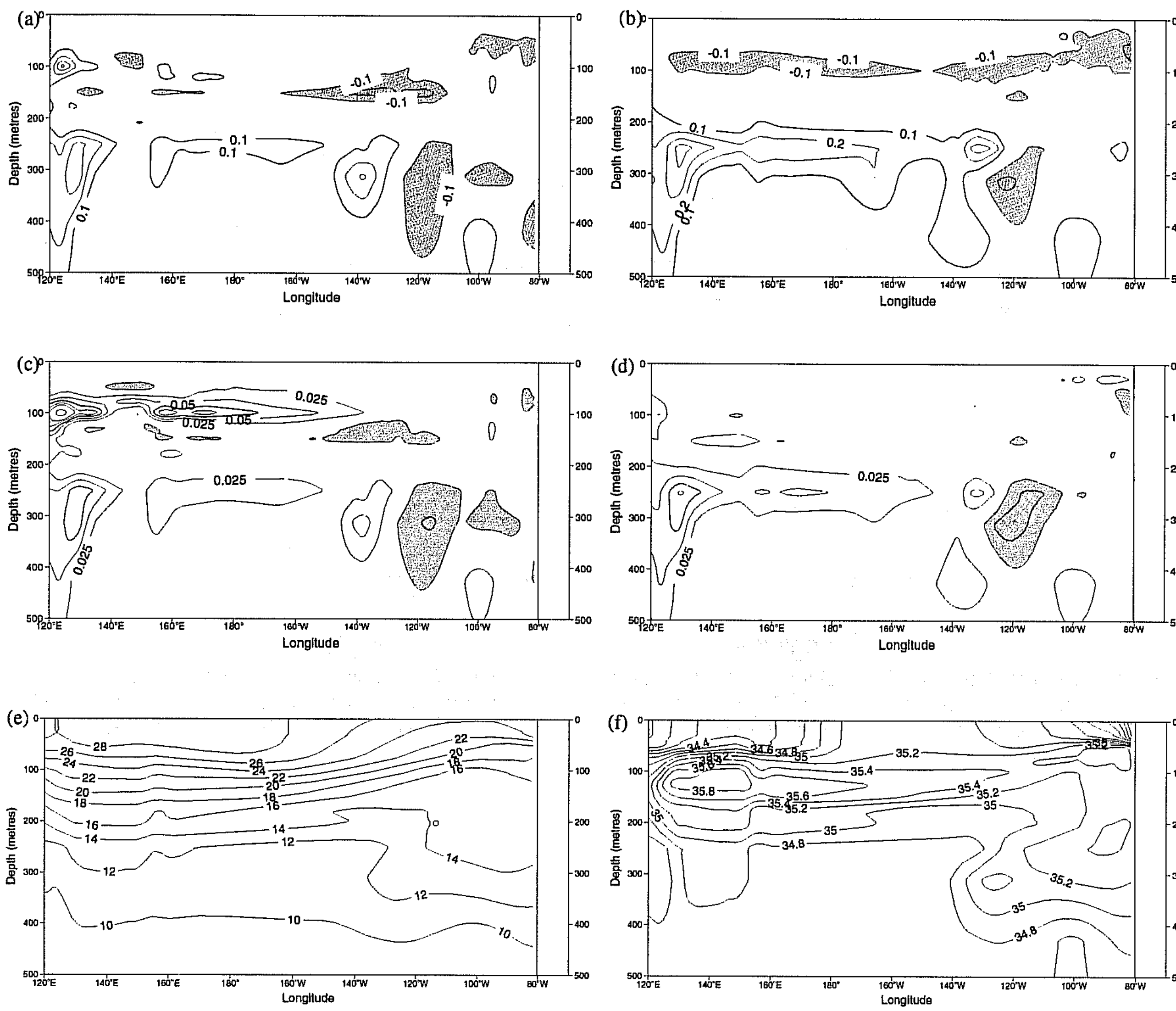

Figure 6. Zonal section along the equator averaged over 1993: (a) temperature error, IM-Control, (b) temperature error, IA3-Control, (c) salinity error, IM-Control, (d) salinity error, IA3-Control, (e) Control temperature and (f) Control salinity. Temperature errors have contours at ±0.1, ±0.2, ±0.3, ±0.4, ±0.5°C, with values less than -0.1°C shaded. Salinity errors have contours at ±0.025, ±0.05, ±0.075, ±0.1, ±0.125, ±0.2 ppt., with values less than -0.025 ppt shaded. The temperature field has contours every 2°C and the salinity field every 0.2 ppt.

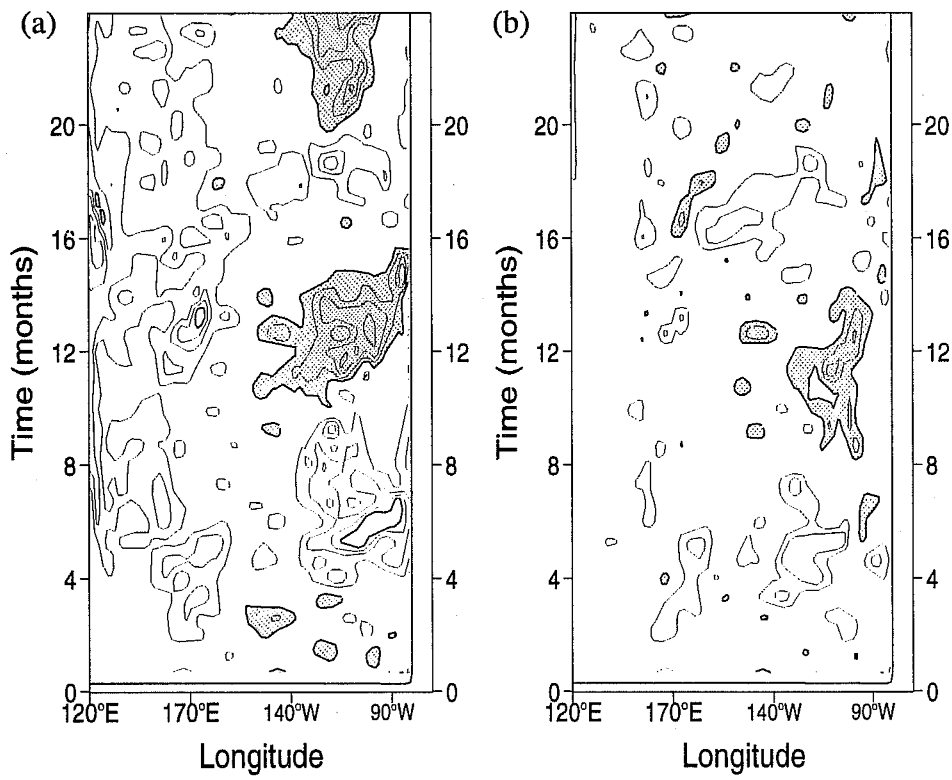

Figure 7. Time evolution of 20°C isotherm error when using an "erroneous" OPS wind stress (a) without assimilation, FM, and (b) with assimilation, FA. Twenty-day averaging has been used to remove some of the noise due to high frequency variability. Contour interval is 5 m, with the zero contour not plotted. Shading indicates areas less than -5 m.

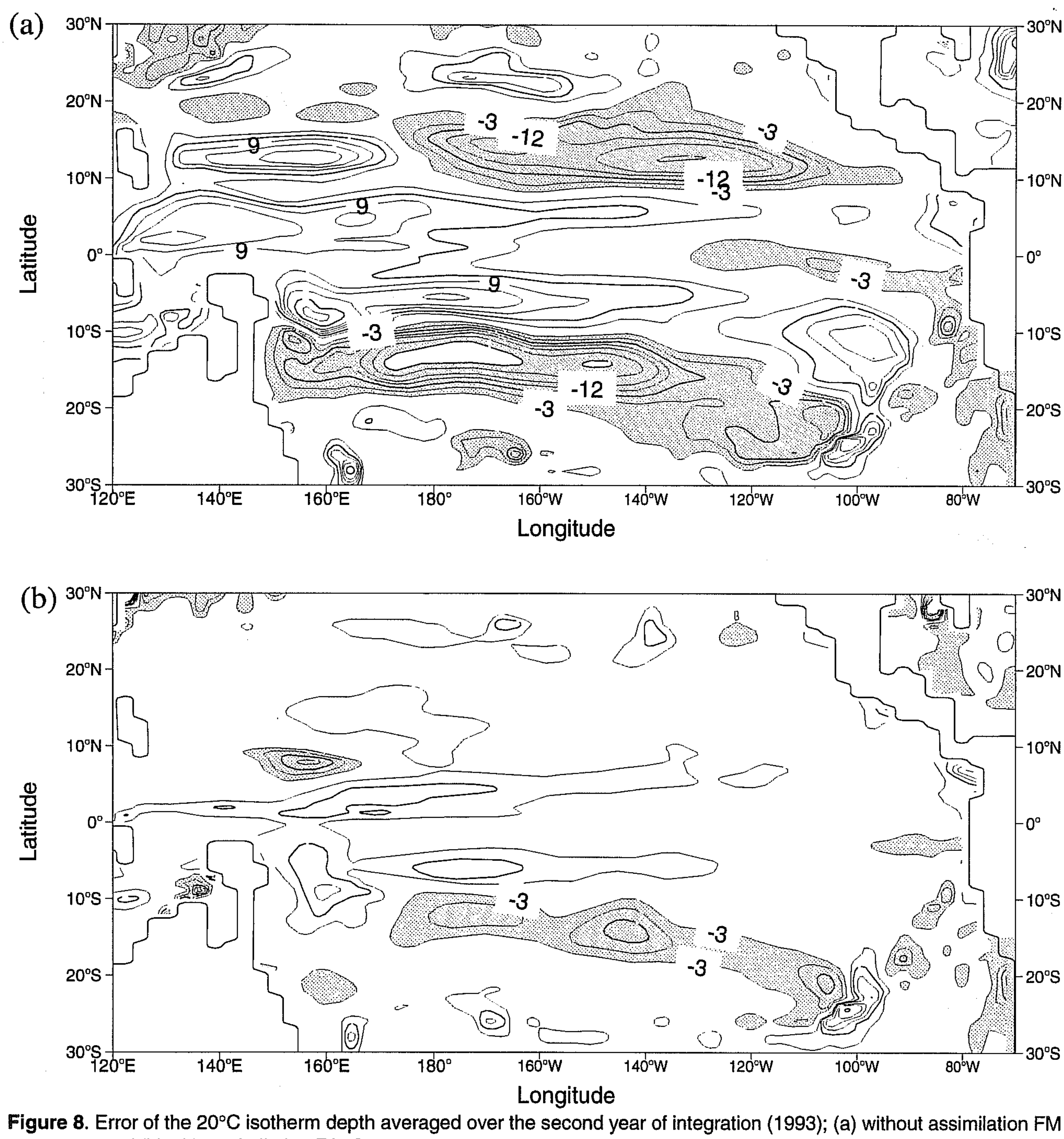

Figure 8. Error of the 20°C isotherm depth averaged over the second year of integration (1993); (a) without assimilation FM and (b) with assimilation FA. Contours are at ±3, ±6, ±9, ±12, ±15, ±20, ±25 m. Shading indicates areas less than -3 m.

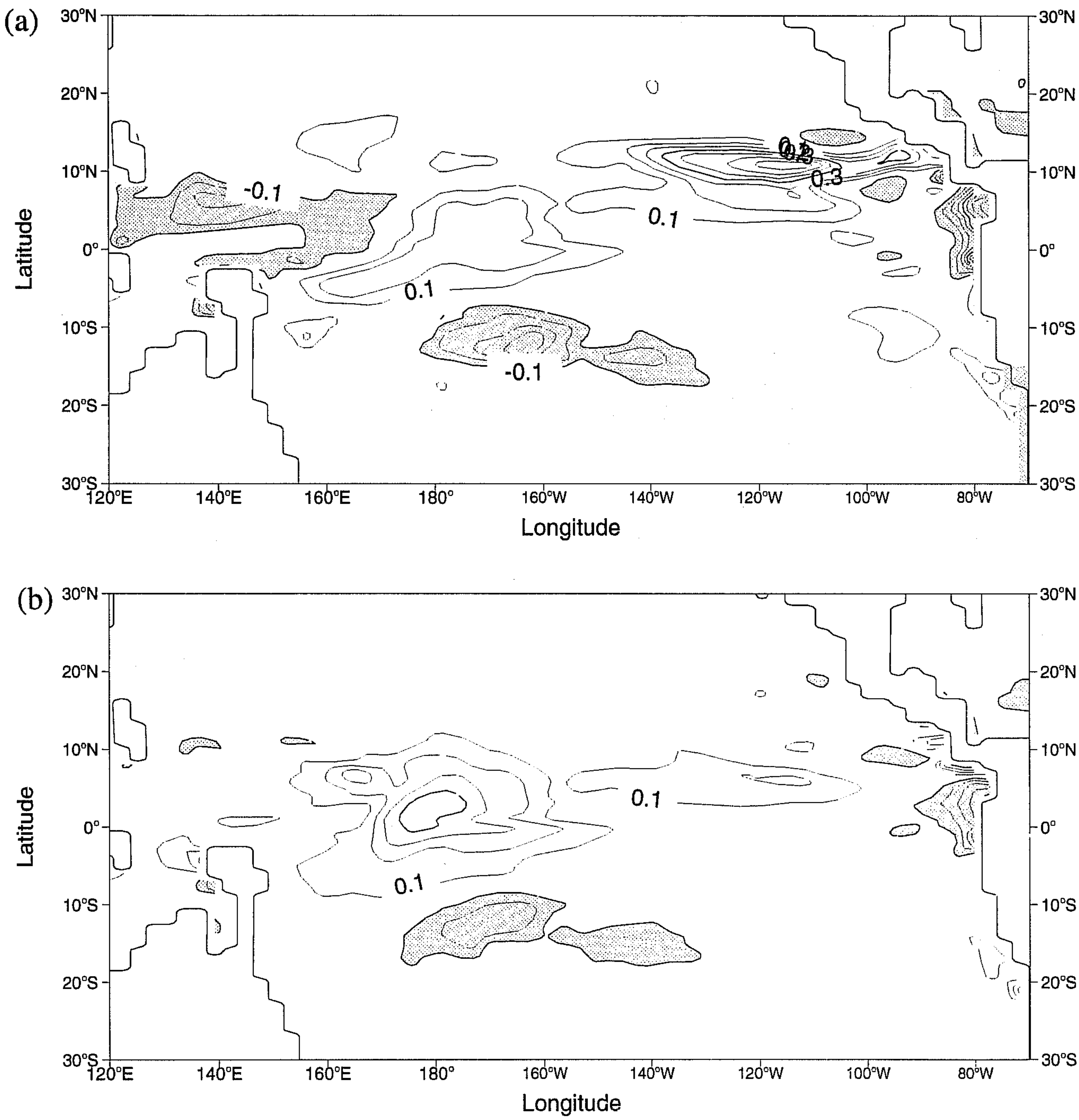

Figure 9. Error in salinity field at 50 m depth averaged over 1993 (a) without, FM, and (b) with data assimilation, FA. Contour interval is 0.1 ppt, with the zero contour not plotted. Shading indicates areas less than -0.1 ppt.

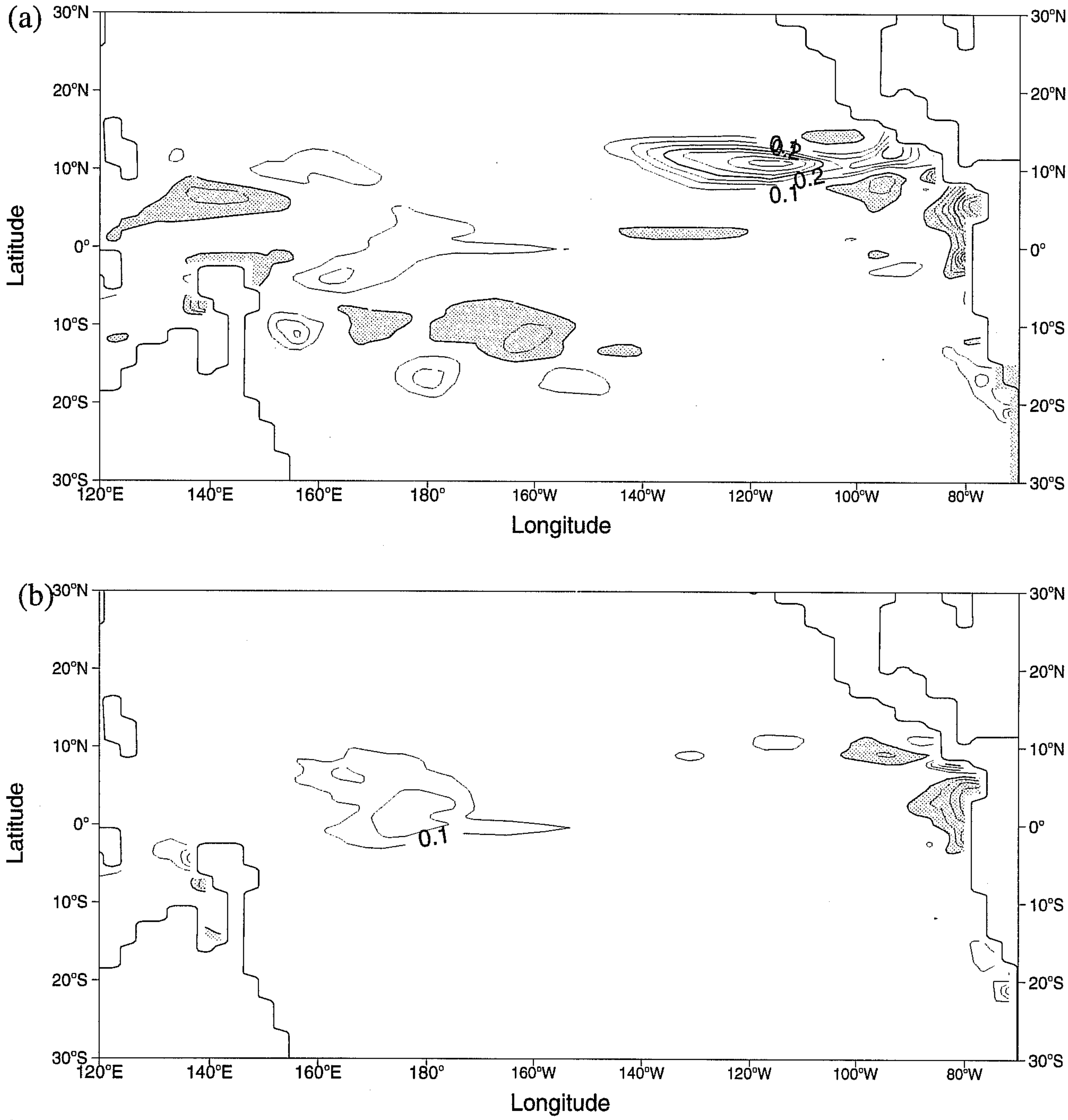

Figure 10. Error in salinity field at 50 m depth averaged over 1993 for the experiments where only the wind stress had errors (a) without data assimilation, FMW, and (b) with data assimilation, FAW. Contour interval is 0.1 ppt, with the zero contour not plotted. Shading indicates areas less than -0.1 ppt.