OCEAN DATA ASSIMILATION FOR SEASONAL FORECASTING

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1 Introduction

Oceanography is somewhat behind meteorology in the use of observational data, either in situ or remotely sensed. That situation has largely arisen because the ocean is a harder environment in which to make real-time measurements. In the past, research vessels were the only source of reliable sub-surface data. However, in the last twenty years, largely as a result of the pioneering foresight of the TOGA programme, that situation is changing. It is now routine to dial up measurements of the ocean thermal structure from the middle of the equatorial Pacific - one of the remotest places on earth- on a daily basis. The work of TOGA has been augmented with data from a sister programme WOCE and more recently by the CLIVAR programme. Despite the valiant efforts of these programmes to get in situ data, the coverage is sparse when viewed in a global context. Remotely-sensed data provide one possible way of building on the in situ observing system to give greater and/or better coverage. Since this seminar series is about satellite data, I will dwell mainly on the use and prospects for such remotely sensed data. Before that, however, I will address what is available by way of in situ data. That perspective is necessarily tainted by the problem to be addressed, the purpose for which the data are wanted. At ECMWF our interest is in seasonal forecasting. That immediately focusses attention on the tropical oceans, especially the Pacific ocean.

In the late 80's an observing array based on buoys anchored to the ocean floor started to emerge. This has been established and developed in the 90's into a very useful observing network. Approximately 60 buoys are moored from 8S to 8N at roughly 15 degree intervals across the Pacific. They are in the water for about 1 year measuring temperature (T) at 10 levels in the upper 500m which are relayed on a daily basis via satellite back to land where they are distributed by ftp and GTS to the oceanographic community. Salinity (S) is now measured at a few spots but not with a density that allows a salinity analysis. We will return to salinity and the problems it creates later. Suffice to say now that it is a much more difficult quantity to measure than temperature, hence our poor knowledge of the salinity field and its variability. In addition to the TAO array and its developing expansion into the Atlantic and hopefully Indian Oceans, knowledge of the ocean comes from ships of opportunity (usually merchant ships plying trade routes) which collect thermal data from XBTs (eXpendible Bathy Thermographs). These XBTs measure the temperature of the upper ocean, some to a depth of 500m, some to 700m. Despite many attempts to develope a salinity sensor for XBTs, no cheap and reliable intrumentation has been possible. A small, yet important addition to the observations described above comes from high quality data from research ships.

There are almost no real-time observations of the velocity field in the ocean apart from velocities inferred from drifting buoys drogued at $\sim 10m$ depth. An exciting development is PALACE floats or an extension, called ARGO floats. These can measure T and S every few days (typically 10) from a depth of 1000m to the surface. Some data of this type are available for the Atlantic in near real time and are already used in the ECMWF ocean assimilation system.

It is against this in situ observing system that we must now consider remotely-sensed data. Electromagnetic radiation does not penetrate the ocean to any great depth: for microwave radiation the penetration depth is just a few mms while for radiation in the visible, the penetration is a few meters depending on the clarity of the water, (which is largely determined by the biological activity).

The most important ocean measurement is SST (Sea Surface Temperature), mainly coming from infrared satellite instruments and ship-based observations. More recently the microwave sensors, less influenced by clouds, are gaining in popularity. The second most important surface measurement is surface stress which can be measured from scatterometers as discussed by Stoffelen this volume. The oceans, particularly the tropical oceans can be viewed as a forced damped system, not generally given to internal instability. Thus a great deal of information about the current ocean state can be obtained from the past history of the forcing fields. The most important of these is the stress although heat and fresh water fluxes are also of interest. At ECMWF, the measurements of sigma0 from the scatterometer are used to modify the atmospheric analysis. The surface stress from this analysis is then used to force the ocean model to provide initial conditions for the ocean component of our coupled ocean-atmosphere model of seasonal forecasts. Improvements in observing the surface wind then feed through the atmosphere giving rise to a better ocean analysis as the surface forcing fields improve. For some purposes, wind fields are derived directly from the scatterometer, without any modification by an atmospheric GCM-based analysis system but these will not be discussed here.

The next most important quantity of interest is measurement of the sea surface height. Although we can not see into the ocean, if changes take place even deep inside the ocean they may have a signature at the surface. For example if warm water replaces cold, then the top surface of the ocean will go up. The altimeter can detect changes in the top surface of the ocean to an accuracy of $\sim 2cms$ from the Topex/Poseidon mission.

A fourth quantity of interest is surface wind (no direction) from the altimeter. This data is just available at nadir and is not assimilated into any model, either of the ocean, of the atmosphere or of the wave field but it can be used to monitor the quality of the atmospheric analysis system as discussed by Jannsen. The fifth quantity is significant wave height. This is not of direct use to seasonal forecasting, nor indeed to open-ocean analysis, but it is of indirect interest as it gives another method of checking the quality of the forcing fields and indeed of improving the way momentum is transferred between the atmosphere and the ocean. This has been covered by P Jannsen (this volume) and I won't say more.

Two other fields are of potentially great use: salinity and gravity. The first will be measured by the SMOS mission. Surface emmisivity is slightly sensitive to surface salinity and therefore there is some possibility of measuring salinity from space. It is not an easy measurement, however, and the objective accuracy for salinity is about .2 psu in 10-day-average fields with a space scale of $\sim$ 200 kms. The gravity field needs to be known to great accuracy in order to use the altimeter to deduce the mean state of the ocean. Gravity missions are in progress as discussed later.

In the rest of this talk I will try to give a review of the way we use altimeter data in the context of seasonal forecasting. Altimeter data is not easy to use as it requires a solution to a difficult inverse proble. Given that the top surface of the ocean has changed, how can you decide how that information should be projected onto the subsurface field. There is clearly an infinity of ways in which this can be done. Selecting the most likely way(s) is a tough problem, not yet solved. Not only will we look at the impact of altimetry on ocean analyses, we will also consider its influence on the quality of the forecasts. Altimeter assimilation is still in its infancy. Although progress has in some sense been significant over the last few years, we are still far from an ideal solution. Ignorance of changes in salinity, both at the surface and at depth prevents a proper use of altimetry.

In the next section I will give some results from idealised experiments on how to project altimeter data into the subsurface. This will be followed by experiments using real data and assimilating altimetry in the presence of in situ thermal data. I have drawn material from Alves et al 2000, Segschneider et al 2000, Troccoli et al 2000 and Segschneider et al 2000.

2 Using altimeter data

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. One of the main aims of the ECMWF operational seasonal forecasting system is the forecasting of ENSO up to six months ahead 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 (Stockdale et al. 1998). 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. Here we 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. 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. One recent study which addressed the more realistic situation of both salinity and thermal contributions to the density field was that of Vossepol and Behringer, (1999)who 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. Here we take a different approach to using altimeter data to correct both temperature and salinity.

At ECMWF we have used the Cooper and Haines (1996) (hereafter referred to as CH) scheme to project the altimeter data onto the thermal and salinity fields. First we will consider an ideal setting in which some of the practical problems of using altimeter data are sidestepped by using an identical twin set-up, based on the ECMWF seasonal forecast system. This allows us to investigate the potential benefit of sea level assimilation, in particular in the tropical Pacific. Both the ability of sea level assimilation to correct for errors in the initial state and to correct for errors due to incorrect surface forcing are investigated. The latter is an important application as inaccurately known winds are a serious source of error in the tropical oceans. We will show that in this ideal situation, assimilation of altimeter data provides a very useful improvement in ocean analyses, largely correcting errors due to wind forcing. This is followed by assimilation of sea level anomalies (SLA) from Topex/Poseidon data and then the *combined* assimilation of sub-surface temperatures (T_{sub}) and sea level anomalies.

2.1 The ocean model and altimeter assimilation system

2.1.1 The 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.

2.1.2 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 may be written;

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

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.

Beneficial properties of the above method are that it is local, is derived purely dynamically, i.e. is model independent, and is easy to apply. It preserves the T/S relationship of the water column and the volume of each water mass and hence the stratification on potential density surfaces, except at the top and bottom and does not initiate convection. 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 partial increments are added each time step, so that by the next assimilation the full increments have been added. This removes the need to make geostrophic corrections to the currents, which cannot in any case be done near the equator.

In this idealised twin 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 discussed later.

2.2 Correction for Initial State and forcing field errors in identical twin experiments

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 to climatology 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. Each integration was performed for 2 years starting on 1 January 1992 and was forced at the surface as for the control.

The sea level from the control is assimilated every 10 days throughout the integration, the increments being added smoothly during the 10 days following each assimilation time 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.

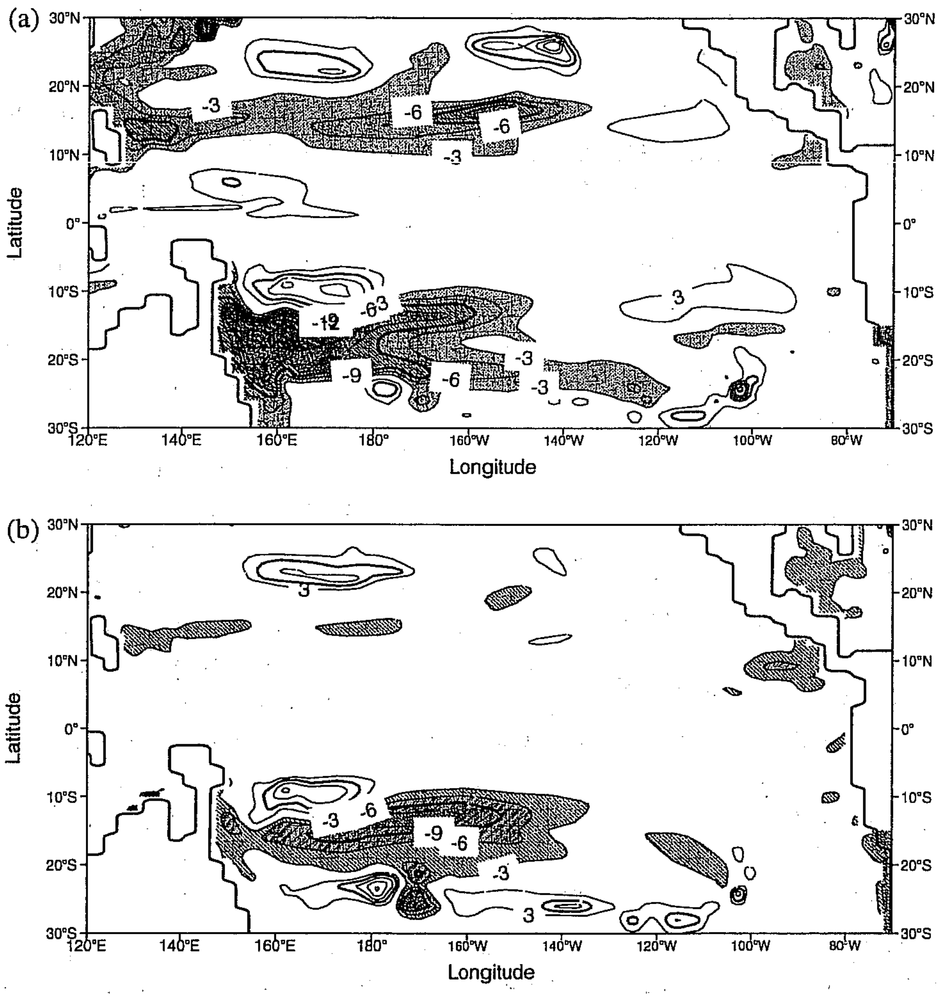


Figure 1: Errors in the 20C isotherm depth for the second year of integration (mean over 1993) relative to the control: (a) IM and (b) IA3. Contour interval is 3 m, with the zero contour not plotted. Shading indicates areas less than -3 m.

The experiments were integrated for a period of two years. Most of the equatorial convergence occurs within the first three months. 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 seen in Figure 1 which shows these mean errors in the 20° C isotherm depth. Without sea level assimilation (fig 1a) 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. 1a and 1b one can see that errors within 10 degrees 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 near 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 are significantly reduced in the assimilation run. For example, between 10° N and 20° N in the western Pacific, errors of up to 9m are reduced to less than 6m. Similarly, between 10° S and 20° S errors of up to 15 m are reduced to less than 12 m.

In the longer term, assimilation has had a smaller impact in the southern hemisphere sub-tropics than in the northern hemisphere sub-tropics. The reason is 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 do not affect the sea level signal. Despite this problem there is a small improvement in fig. 1b over fig. 1a even in the southern subtropics.

The errors in the yearly-mean salinity along the equator at 100m are reduced by sea level assimilation in the central and western Pacific but at depths of around 250 m are increased slightly. 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 central and western Pacific the initial sea level error is dominated by the positive salt errors at around 100 m and 250 m. Since the water column is too salty and dense, the CH scheme will lower the water column. At around 100 m depth, salt is increasing with depth and so lowering the water column reduces the salt error. 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.

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 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 the experiment 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 should help to correct. The 1992-93 period was again considered. The control experiment is the same as that described earlier. 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 as errors. The sea level assimilation was carried out every 10 days using data from the control and the increments were split and added every time step as in earlier assimilation experiment.

Assimilating sea level reduces the errors in the thermocline depth (fig 2). 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 when the error in the 20C isotherm depth increases from 5 m to around 10 m. This occurs concurrently with the larger sea level errors. Off the equator data assimilation has an even greater impact on the thermocline depth. The errors in the isotherm depth are significantly reduced by assimilation. For example, errors in the subtropics of over 15 m are in general reduced to less than 3 m. 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.

For salinity, sea level assimilation leads to a decrease in the error almost everywhere other than for the equatorial Pacific near the dateline. 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 increases 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 and 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 e.g. from SMOS should reduce this problem.

2.2.1 Summary and Discussion of identical twin experiments

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, viz errors due to starting from an erroneous initial state and errors arising from the forcing were considered.

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, altimeter assimilation with the CH scheme occasionally made the T/S errors worse. The scheme performed less well in regions where intermediate salt maxima are found around the thermocline region, in the sub-tropical south Pacific and equatorial west Pacific, for example.

Sea level assimilation was found to effectively correct for errors due to the surface forcing, particularly those due to the winds. 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

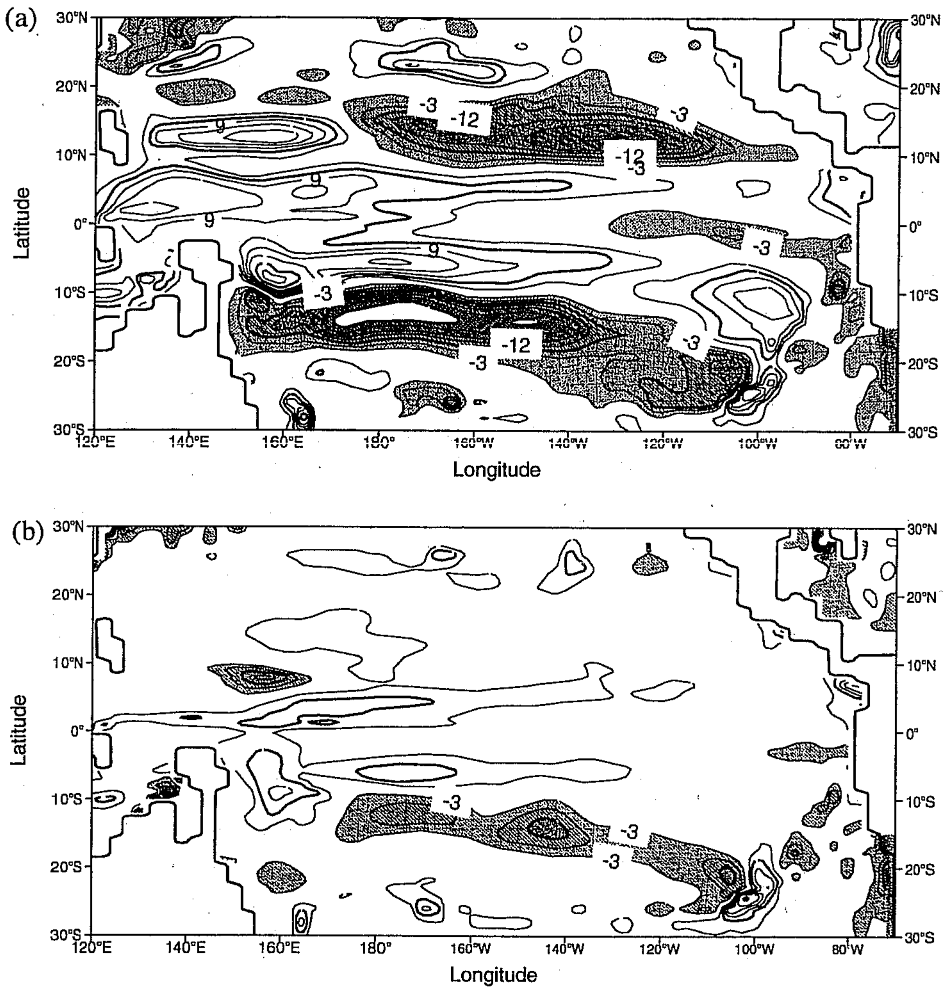


Figure 2: Errors in the 20C isotherm depth for the second year of integration (mean over 1993) (a) without assimilation and (b) with SLA assimilation. Contour interval is 3 m, with the zero contour not plotted. Shading indicates areas less than -3 m.

the salinity field in the mixed layer. The CH scheme is not able to 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 further experiments which combined both initial state and forcing errors it was found that 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 section we have shown that there is potential benefit from assimilating altimeter data in the tropical Pacific and suggested that the benefits would increase if enough surface salinity observations were also available.

3 The use of altimeter data from Topex/Poseidon

3.1 The altimeter data set

In the previous section, we showed that the assimilation of altimeter data through the CH scheme could improve the ocean analysis, at least when using perfect altimeter data and in the absense of in situ data. We now want to relax these constraints. First we will consider real data, but still in the absense of other in situ observations. Later we will consider methods of assimilating altimetry in the presence of in situ data.

Every 10 days, maps of sea level anomalies relative to the 1993 to 1995 period are produced by Centre Localisation Space, located in Toulouse, France, denoted HH (homogeneous historique). From May 1998, we use maps which are available weekly and in real-time. The latter maps are derived from the TOPEX/Poseidon and ERS-1/2 satellites, and have a spatial resolution of 0.25° x 0.25° (LeTraon et al., 1995, 1998). At ECMWF the maps are smoothed because the ocean model has too coarse a resolution to resolve the eddies contained in the data. The maps are then interpolated to the model grid, where a mean sea level is added to the anomalies. The mapped data are assimilated every 10 days using CH, and the derived temperature and salinity increments are spread over the following 10 days to allow a smooth adjustment of the density field.

The above method requires the initial interpretation of SLA in terms of subsurface temperatures and salinities. The CH method that we use requires a mean sea level, which is not yet provided with satellite observed sea level. We have taken this mean sea level, for the years 1993-1995, from a former subsurface T assimilation experiment with the ocean model (denoted OI-3 in Segschneider et al., 2000a). Having to use a model-derived mean sea level in the assimilation process is a weakness in present attempts to assimilate altimeter data, preventing its full potential from being realised. This deficiency should be overcome once a sufficiently accurate geoid is available to provide absolute sea level observation from space. Such an accurate geoid should be available from gravimetric satellite missions within the next few years. The successful launch of the CHAMP (Catastrophes and Hazard Monitoring and Prediction) mission in July 2000 represents a useful start to this process which will continue with the GRACE (Gravity Recovery and Climate Experiment, launch 2002) and GOCE (Gravity and Ocean Circulation Experiment, launch 2004) missions. In the meanwhile we accept the limitation of not having a measured mean sea level. This problem exists in all present attempts to assimilate altimeter data, although it can be disguised if only temporal information is used.

In order to use the altimeter data operationally to provide oceanic initial conditions for the coupled forecasts, we use high quality HH (Historical Homogeneous) data until April 98 and then use near real time data. This experiment is denoted ALT-HHNRT.

3.2 Assimilation of subsurface temperature observations

Subsurface temperature observations are presently assimilated into the ocean analysis via an optimum interpolation scheme (Smith et al., 1991). For almost all the experiments us-

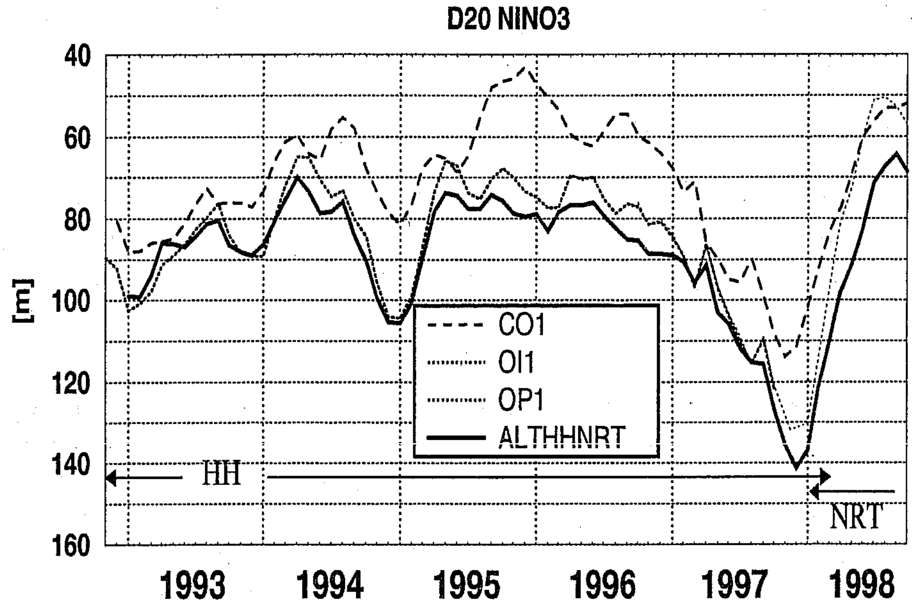


Figure 3: Time series of simulated D_{20} averaged over Niño-3 for the control experiment (i.e. no data assimilation), indicated by the dashed curve, the OI assimilation of only in situ thermal data shown dotted (OI-1 upto 9/97 and OP-1 from 4/96), and the altimeter assimilation ALT shown as solid. Arrows indicate the periods covered by HH data and NRT data, respectively.

ing in situ data, manually quality controlled observations including XBT, CTD and TAO data are used. The exception is OP-1 which uses auto quality control. Every 10 days, observations are collected over a centered 10-day period. The observed profiles are interpolated onto model levels. The optimal interpolation scheme is then solved independently for each model level. Individual observations and the modelled background are given the same weight. The background and observation error spatial structures are modelled using gaussian functions. Background error decorrelation scales are nonisotropic and vary with latitude. At the equator the zonal scale is 1500 km and the meridional scale is 200 km. As latitude increases the zonal scale decreases and the meridional scale increases to 400 km in both directions at 20°. The observation error decorrelation scale is homogeneous and isotropic, with a space-scale of 200 km and time-scale of 3 days. Because the TAO moorings provide daily data at the same location, relatively strong weight is given to TAO observations compared to XBT observations. The latter are mainly along ship tracks and therefore more than 2 or 3 XBT observations are seldom found within the background error decorrelation radius over a 10 day window. Once the analysis has been computed, the derived temperature increments are applied over the next 10 days to minimize adjustment problems. In this original temperature OI scheme, there is no correction to salinity, which can give rise to unreasonable salinity profiles as discussed by Troccoli et al. (2000).

The main difference compared to the assimilation of sea level data is that assimilation of temperature observations is capable of correcting the vertical temperature structure, while sea level assimilation using CH can only correct misplacements of the temperature profile in the vertical. The major drawback is the sparseness of data, especially outside the domain of the TAO array. On the other hand, satellite observations span a near global domain and have better spatial coverage even within the equatorial region, though their temporal resolution is worse than for TAO.

The Niño-3 averaged time series of D_{20} from experiment ALT is compared with the control experiment and the in-situ temperature assimilation experiment OI in Fig.3. The OI-1 experiment is continued using results from the real-time ECMWF ocean analysis system denoted OP-1. The control experiment tends to simulate too shallow a thermocline, whereas the assimilation of altimeter data replicates the 'observed' D_{20} quite closely, although the thermocline is too deep at the peak of the 97/98 El Niño, relative to the operational analysis, OP-1. The subsequent rise of the thermocline is well captured in the near real time altimeter analysis, but D_{20} is never as shallow as in OP-1 and in fact the time of minimum depth is slightly delayed. Whether the larger discrepancies, which occur during the phase in which NRT data are assimilated, are caused by lower quality of the NRT data or whether salinity-effects, for example, are playing a significant role can only be assessed when the HH data for the same period have been processed. They are not available yet.

3.3 Assimilation of altimeter data in the presence of in situ observations

In this section we seek to assess the importance of altimetric measurements when there are other in situ observations. Although the altimeter data will be assimilated based on the CH scheme, there are several ways of doing it. We will consider only two. They differ in the way that salinity is corrected.

In the following, we aim to combine the advantages of using altimetry and in-situ temperature observations. In so doing, we first assume that the directly-observed temperature profiles are the more direct and hence more reliable source of information about the subsurface temperature field wherever such observations exist. The information from the altimeter is then mainly used in areas where few direct temperature observations exist, such as in the tropical Atlantic and Indian ocean and outside the TAO array in the Pacific. The sea level observations are also used to update the salinity field over the whole model domain using the information about salinity provided by the CH scheme.

An optimum method to combine the altimeter and thermistor data has yet to be found, but we feel that our approach is useful for our global ocean analysis. We first analyse temperature and salinity from the altimeter data for every gridpoint using CH as described above. The altimeter-derived temperatures are not directly used to update the background field, however. Instead, the altimeter-derived temperatures are used as *pseudo* or *synthetic* temperature observations which are combined in the OI with the *in-situ* temperatures from TAO/XBT and the model background. We thin the *pseudo* temperature observations before the OI. Every second latitude band is dropped in an initial step. In a second step, we check for each of the remaining gridpoints, whether an *in-situ* temperature observation is present in the surrounding 2° x 2° box. If so, the respective *pseudo* observation is dropped. Because the 'exclusion-box' is smaller than the decorrelation scales used for the OI, in particular close to the equator where the zonal decorrelation scale is 1500 km, there is still influence from the altimeter at *in-situ* temperature observation points, but the weight given to the altimeter data is smaller because of the greater distance.

3.3.1 Correction of sub-surface salinity

As outlined above and described in detail in Troccoli et al. (2000), the assimilation of only temperature data without correction of the model salinity field can introduce first order errors in the temperature field. These errors can be substantially reduced when salinity is updated together with temperature. Furthermore, recent studies have shown that subsurface salinity variations impact strongly on sea level variations in the Western Equatorial Pacific (e.g. Ji et al., 2000) and in the Western Equatorial Atlantic (Segschneider et al. 2000b). Because salinity observations are sparse, in particular those available in near real time, it is not possible to analyse S based on direct measurements. However, an indirect approach that locally uses the T-S relationship of the background field to find the salinity that matches the newly analysed temperature at each level has been developed by Troccoli and Haines (1999, henceforth TH99).

The improvements from using TH99 on the temperature fields when only temperature was assimilated have already been described in Troccoli et al. (2000). The ocean analysis was improved in particular in regions of pronounced salinity maxima at intermediate depth, such as in the western equatorial Pacific and Atlantic oceans. However, most of the improvements of the temperature fields were achieved at depths of more than 200m, and it is thus not immediately clear what the impact on seasonal forecasts might be. Here we apply TH99 in the presence of temperature and altimeter data and we will examine not only the ocean analyses, but also the impact on coupled forecasts. Adverse changes in T or S below the pycnocline may not directly affect the SST and propagation of upper ocean heat content anomalies and hence the forecasts. They influence the sea level, however, and through that, prevent the optimum use of altimeter data. In the next section we describe more fully how the salinity is corrected in various experiments.

3.3.2 Set-up of experiments

In the following, we will briefly describe the set-up of two further ocean analyses that are used to gauge the impact of the different assimilation approaches. The experiments are run for the same period as the ocean analyses in the previous section. As before, the ocean model is forced by daily averages of momentum, precipitation minus evaporation, and heatflux derived from the ECMWF re-analysis (ERA-15) for 1993, and from the operational numerical weather prediction system for 1994 to 1997. The model SST is relaxed to weekly-averaged analyses of SST (Reynolds and Smith, 1995) with a time-scale of three days, and the sea surface salinity is relaxed to climatological data from Levitus and Boyer (1994) with a time scale of one month. Subsurface temperature and salinity are both relaxed to climatological data with a time scale of one year. The OI is applied down to model level 15 (approximately 1000m).

To recap, in the first experiment, called T-OI, only subsurface temperatures are assimilated. In a second experiment, called ALT, only altimeter data are assimilated. In the third experiment, called T+A, both altimeter and temperature observations are assimilated as described in section 3.3. In experiments ALT and T+A, the salinity increment from the sea level assimilation is used to up-date salinity on all model grid points. In the fourth experiment, called TA+TS, temperature and sea level are assimilated as in T+A, but salinity is corrected using the scheme described in Troccoli et al 2000.

3.3.3 Results from the ocean analyses

In this section, results from the ocean analyses assimilating altimeter data in the presence of in situ thermal data will be described. In panel a we show sea level in the Nino3 region from various expewriments. The ALT experiment is taken as a control and shown by a heavy solid line. All experiments reproduce sea level rather well. Even experiment TOI which has no altimeter data agrees well with ALT implying that the assimilation of sub surface data does a good job of reproducing sea level changes. There are areas in which this is less true, as discussed by Segschneider et al 2000.

Our main emphasis is on upper ocean heat content as represented by D-20. We shall show results for the tropical Pacific and Atlantic oceans. In the following figures, D-20 of experiment T-OI will be drawn as solid black lines, as it will be taken as reference for the combined assimilation experiments.

We will first consider the tropical Pacific. In the Niño-3 region, the deviations of D-20 between experiments are generally smaller than 5m, though on occasion they can reach 10m, as in experiment ALT in 1995 for example (Fig. 4b). By combining the two observation types, the departures of D-20 of experiment ALT from the reference can be further reduced in experiments T+A and TA+TS as seen in Fig. 4b. The most pronounced feature in Fig. 4 is the strong deepening of the thermocline in 1997. The good representation of D-20 in experiment ALT, where only altimeter data are assimilated, underlines the ability of the model to reproduce the time-varying temperature structure and the usefulness of the CH scheme which actually takes the vertical structure of the model temperature field into account.

In the Niño-4 region the agreement for the D-20 time series is again excellent, and what little deviation there is in the altimeter-only assimilation experiment (ALT) is further reduced by the additional assimilation of temperature observations in experiments T+A and TA+TS, as one would expect. For the sea level in Niño-4 where salinity variations contribute more to sea level changes than in Niño-3, the agreement between the experiments is slightly worse than for the Niño-3 region. Beginning during the second half of 1995 and through until mid 1997 the sea level is too high in experiment T-OI compared with the reference experiment ALT, with differences in the range of 5 cm. This period has been discussed at some length by Ji et al. (2000) who noted the importance of salinity. One would expect that the additional correction of salinity in experiment TA+TS should improve the sea level, but in fact the sea level is more similar to that of T-OI than to ALT over the relevant period, whereas experiment T+A agrees quite well with experiment ALT.

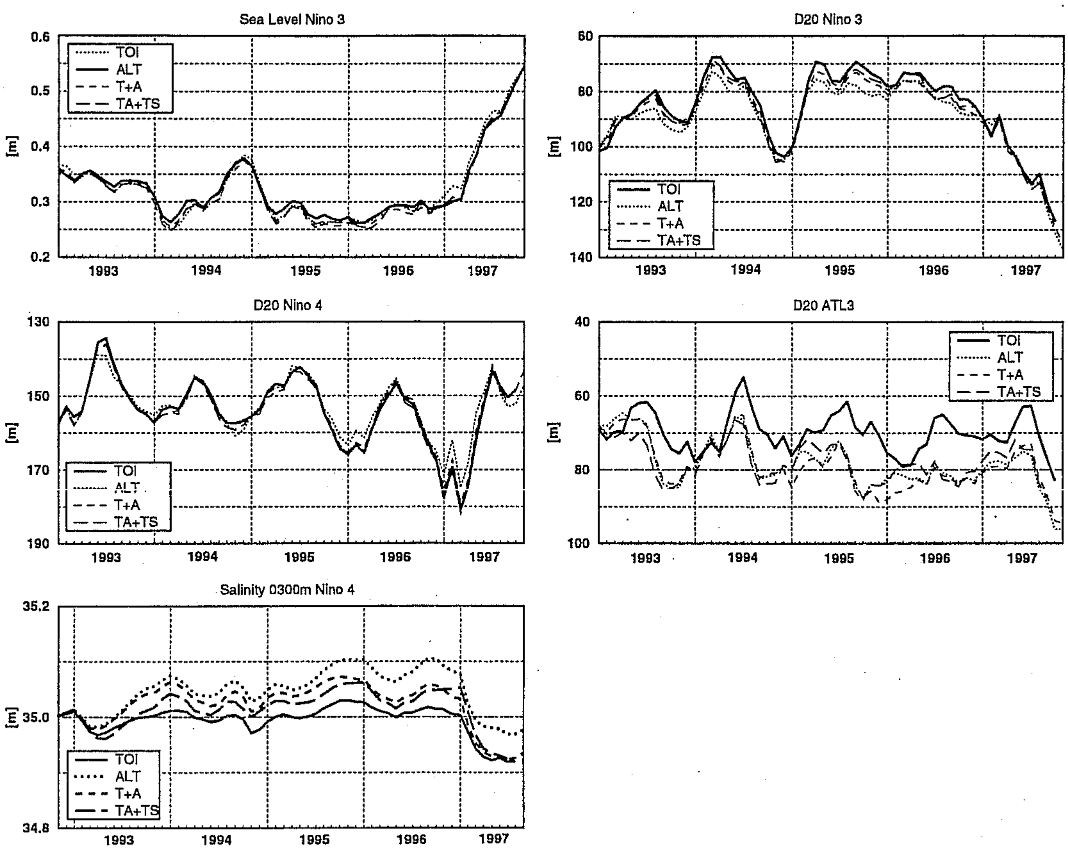


Figure 4: Sea level (a) and D-20 (b) averaged over the Niño-3 region, D-20 in the Niño-4 region (c) in the ATL3 region (d) and the average salinity in the Niño-4 region (e), from experiments T-OI, ALT, T+A, and TA+TS.

Kessler (1999) argues that salinity in the western Pacific changed due to the horizontal advection of saltier water from the east. The TH99 scheme, that searches for the matching salinity in the vertical, can not correct the salinity error of the model, and as a consequence the sea level is similar to that of experiment T-OI. An improved version of TH99 that also searches in the horizontal is under development, but has not yet been tested in a tropical application.

In the Atlantic, standard regions are less commonly used than in the Pacific. We will consider ATL3 which spans the region between +/-3° N/S, 0 - 20° W. In the ATL3 region both the seasonal cycle and interannual variability are weak. The variations in D-20 show larger differences than in other regions of the tropical Atlantic. The three altimeter experiments are reasonably consistent but differ by up to 20m from D-20 in the T-OI experiment. The pattern of variability in D-20 in T-OI is less well correlated with that in the altimeter experiments than is the case in other regions.

The role of salinity in the framework of temperature assimilation at ECMWF is discussed by Troccoli et al. (2000) and a brief discussion on the role of salinity when altimeter data are assimilated is given in Segschneider et al. (2000a, 2000b). Here we will briefly discuss salinity for the experiments in which both sea level and subsurface temperatures are assimilated. We show average salinity integrated over the upper 300m (S300), where most of the salinity variations occur, for the Niño-3 and Niño-4 regions in Fig.4e. Differences in the integrated salinity result in differences of sea level and therefore impact on the altimeter data assimilation. The treatment of salinity during the assimilation is as follows: in T-OI salinity is not updated at all, in experiment ALT and T+A salinity is updated using vertical shifts of the model background profiles based on sea level only, and in experiment TA+TS the TH99 scheme is used to correct salinity based on the analysed temperature. When temperature observations are assimilated together with sea level, the vertical structure of the temperature fields can be altered, and the salinity fields of experiment T+A (using salinity from CH) and TA+TS (using salinity from TH99) can be different.

Differences in S300 between the experiments are present but it is not straightforward to decide which of the 4 experiments performs best, as there is little salinity data for verification. In the Niño-3 region (not shown) the integrated salinity of experiment T-OI is lowest, and quite similar in experiments ALT, T+A, and TA+TS. From 1995 onwards, experiments T+A and TA+TS are still quite similar, whereas experiment ALT has slightly higher salinity. In the Niño-4 region T-OI shows less variability than the three other experiments. The differences between the averaged salinity for all experiments are less than 0.1psu for both areas. This seems small but assuming an average temperature of 19° C and salinity of 35psu, this translates into a sea level difference of slightly more than 2cm (based on Table A3.1 in Gill, 1979).

4 Coupled Forecasts

In this section we estimate the impact of altimetry on seasonal forecasts. Climate forecasts over six months are performed by coupling the HOPE ocean-model to the ECMWF atmospheric Numerical Weather Prediction model at T63 resolution by use of the OASIS (Ocean-Atmosphere-Sea-Ice-Soil) coupler. Because a fully coupled system is used, the coupled model drifts. To couple only the anomalies of the subsystems is one possible means of reducing such a drift, but there are disadvantages in anomaly coupling. So at ECMWF the two systems are fully coupled and subsequently an estimate of the drift is computed, which can then be subtracted from the model results a posteriori (Stockdale et al 1998).

Forecasts are started every three months (i.e., on 1 January, 1 April, 1 July, and 1 October of each year) from 1.1.1993 to 1.10.1997 and integrated forward in time for 184 days. The atmospheric initial conditions are taken from the operational analysis/forecast system run at ECMWF for weather forecasting. Oceanic initial conditions are obtained from the control ocean analysis (the coupled experiment is denoted C-CO), the subsurface temperature analysis OI-1 (coupled experiment denoted C-OI, and altimeter analysis (coupled experiment denoted C-ALT) and two sets of forecasts based on analyses combining altimetry and in situ data, denoted C-T+A and C-TA+TS.

Because of the chaotic nature of the atmosphere, extensive ensembles of forecasts are required. For each forecast-date we generate an ensemble by perturbing oceanic initial conditions in the equatorial Pacific, in particular SST-perturbations of 0.01° C amplitude are applied between 5° N and 5° S to the regions EQ-1 (90° W - 130° W) or EQ-2 (130° W -

170° W). The chaotic response of the atmospheric model to the applied SST-changes causes a spread of the ensemble members of the forecasts. Here, each ensemble consists of five members. Since there are 20 start dates this amounts to 100 coupled forecasts for each experiment and a total of more than 300 years of coupled model integration.

The model drift, based on the years 1993 to 1997, was computed for each of the coupled experiments. The SST drift is computed separately for forecasts starting in January, April, July, and October from the difference between the predicted and the observed climatologies for the five year period. Because the model drift is not spatially uniform, it is estimated over subdomains in the equatorial Pacific, namely Niño-12, Niño-3, Niño-4, and EQ-1, EQ-2, EQ-3. The drift is largest for the control forecasts except in the near coastal region Niño-12. The drift is frequently larger than the signal we are seeking to predict, however. Although it is in general desirable to have a reduced drift, and in that sense the assimilation of altimeter data has been beneficial, there is not a simple relationship between drift and forecast skill. In experiments in which we inserted a correction to the heat flux of $15W\ m^{-2}$ in the tropical strip the overall drift to cooler temperatures was largely removed, but there was little impact on the forecast skill. A more detailed description can be found in Stockdale (1997). As will be shown below, the reduced drift of C-ALT relative to the forecasts using in-situ temperatures does not result in better forecasts.

We estimate the relative performance of the coupled experiments by comparing predicted SST anomalies to observed values (Reynolds and Smith, 1995). Although comparing only SST anomalies in the equatorial Pacific does not make full use of the information that is provided by the forecasts, the comparison is restricted to these regions in the tropical Pacific because they are so important for seasonal forecasting. In the first method, we compare SST anomalies in the EQ-2 region and compute rms-errors and anomaly correlations. We first compare the C-OI and C-ALT results against the control C-CO.

Fig.5 shows a) the rms-error between predicted and observed SST anomalies and b) the anomaly correlation coefficient. Rms-error is *first* computed for each of the five forecast sets making up the individual experiments whereas for the computation of the anomaly correlation coefficient the five ensemble members for each start date are first averaged and then the anomaly correlations of the ensemble averages are computed. The vertical bars indicate two times the standard deviations of the rms-error. The rms-error is largest for the control forecasts. In particular for the first three months the error of the control forecasts grows quickly and is larger than for the persistence forecast. In experiments C-OI and C-ALT the rms-errors grow more slowly than for the persistence forecast. Rms-errors appear to be slightly larger in the experiment using altimeter data, but differences between the two experiments are small. We applied a Wilcoxon-Mann-Whitney test to the 100 ensemble members to investigate whether the mean of the pdf (probability density function) of the absolute error was significantly shifted between C-OI and C-ALT. It turned out that the shift was only significant at the 40% level. This means that based on 100 forecasts the relative performance of experiments C-OI and C-ALT was not statistically different. By contrast, the control forecasts differ from the two other experiments with a significance greater than 99%.

We will now investigate whether the combined assimilation of subsurface temperatures

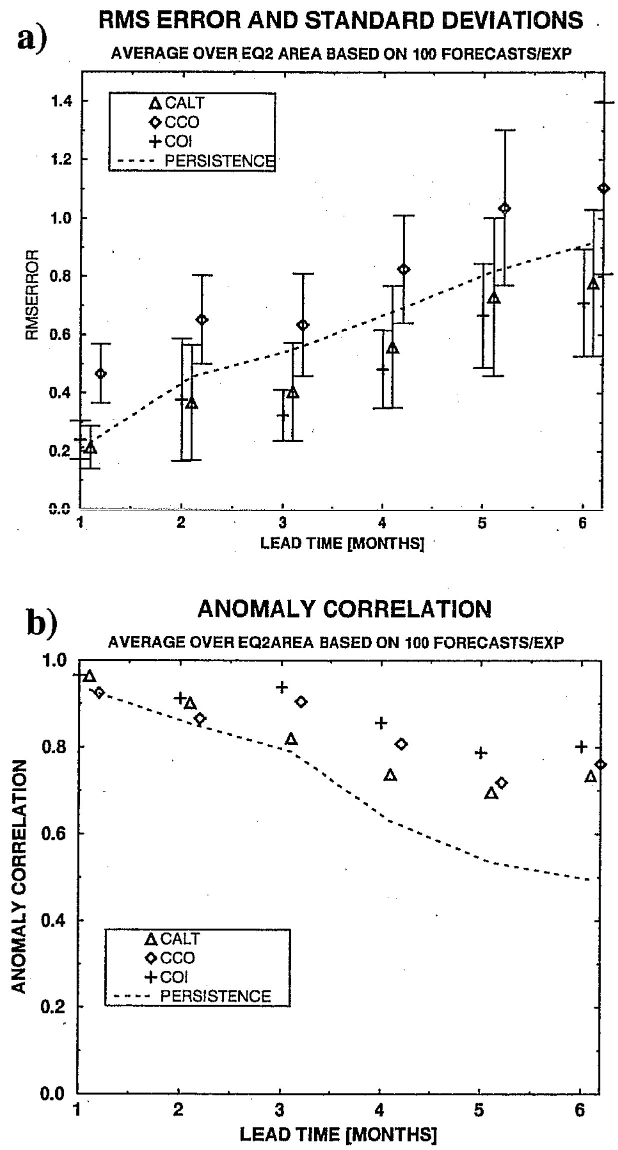


Figure 5: a) rms-error between predicted and observed SSTAs as average over all ensemble members and forecast dates (1. January, 1. April, 1. July, and 1. October) in the EQ-2 region in the central equatorial Pacific (130° W to 170° W, 5° S to 5° N), vertical bars indicate two times the standard deviation, centered at the ensemble mean value. b) anomaly correlation coefficient of the ensemble means for EQ-2. Shown are experiment C-CO (diamonds), C-OI (plus-signs), C-ALT (triangles), and the persistence forecast (dashed line).

and sea level, and the additional correction of the salinity field using the TH99 scheme, can further improve the forecasts. We will compare four sets of forecasts started from the experiments T-OI, ALT, T+A, and TA+TS. The traditional measures of forecast skill, rms-error and anomaly correlation coefficient (ACC), are shown in Fig 6 for experiments C-OI, C-T+A, and C-TA+TS as averages over the EQ-2 (a,b), Niño-3 (c,d), and Niño-4 (e,f) regions. Fig 5 showed that experiment C-OI performed best in terms of rms-error and ACC compared to the control forecasts and the altimeter initialized forecasts. Any improvement of the skill from the combined assimilation of temperature and sea level in C-T+A, and the additional application of the TH99 scheme in C-TA+TS, should therefore be measured against C-OI.

For the rms-error in the EQ-2 region (panel a) such an improvement is not evident: all coupled experiments give very similar results. The deviations between them are less than 0.05° C even for a lead time of six months and are much smaller than the individual standard deviations which are on the order of $\pm.25^\circ$ C (shown by the vertical bars). The ACCs of all coupled experiments are on the order of 0.8 for six months lead, and are clearly better than for the persistence forecast (thin dashed line). The ACCs are very similar for all experiments (panel b). In the Niño-3 region the differences between the experiments are slightly larger for both rms-error and ACC. Experiment C-TA+TS gives higher correlations than C-OI and C-T+A by almost 0.1 for lead times of 4-6 months. The rms-error is also smallest in C-TA+TS, but improvements are less clear than for the ACC. In the Niño-4 region where SST-variations are smaller than in the eastern equatorial Pacific, deviations between the experiments are again small for both rms-error and ACC. According to a Wilcoxon-Mann-Whitney test, however, none of the differences is significant at a level of confidence of more than 70%.

Another skill measure is what we define as the relative performance of pairs of forecast sets. A set of forecasts is 'better' if it is closer to the observed SSTA by more than a specified threshold value, which here is chosen to be 0.3K. The basic motivation to define this skill-measure is that for some applications of El Niño forecasts it might be more important to know which forecast system is closest to the observations on most occasions rather than how the rms-error for the whole set of forecasts performs.

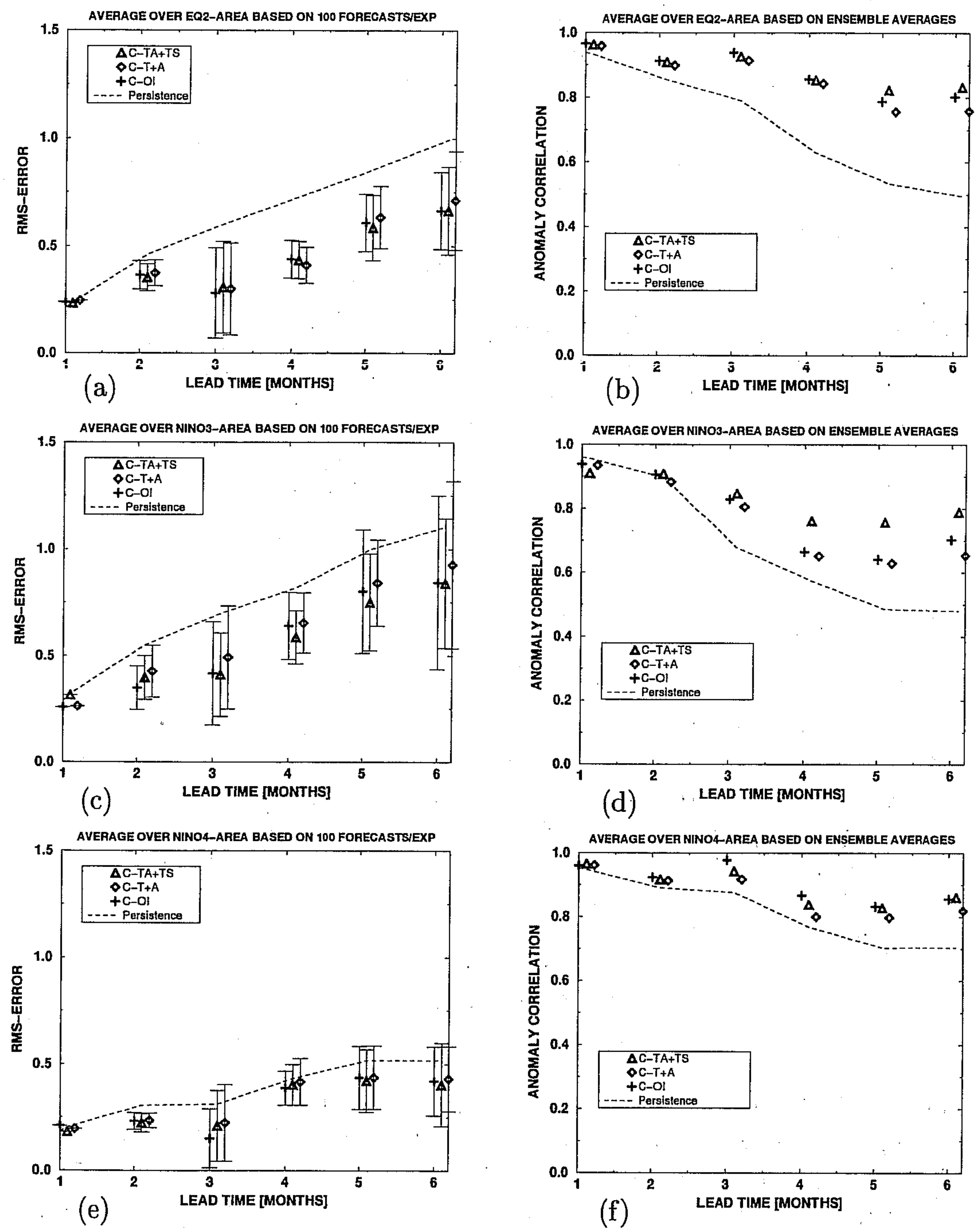
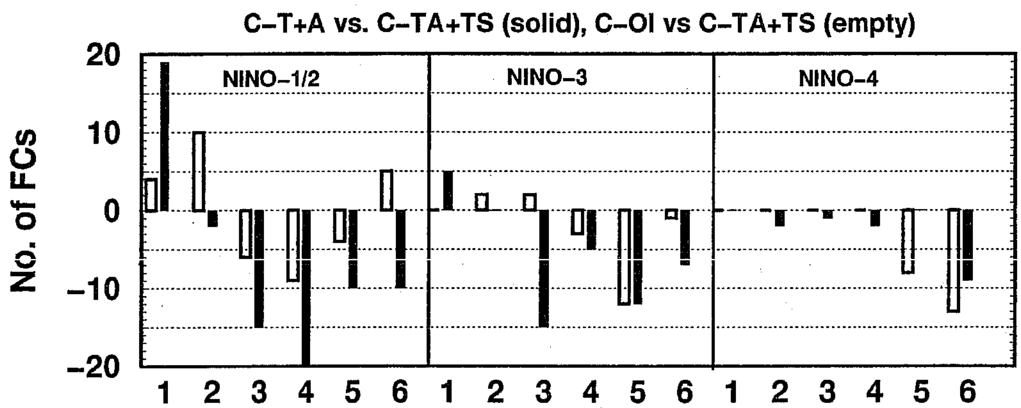
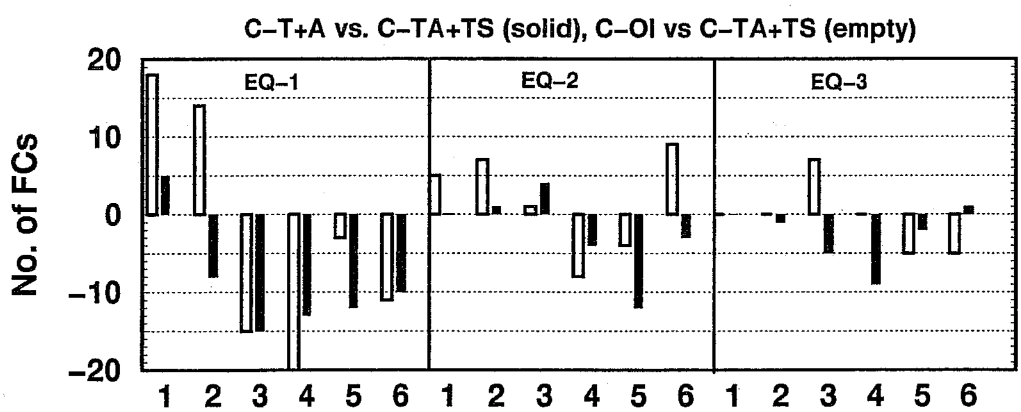


Figure 6: Ensemble average of the rms-error (left column) and the anomaly correlation coefficients of the ensemble mean (right column) averaged over (a,b) the EQ-2 region (130° W to 170° W, 5° S to 5° N), (c,d) the Niño-3 region (90° W to 150° W, 5° S to 5° N), and (e,f) the Niño-4 region (150° W to 170° E, 5° S to 5° N) for experiments C-OI (plus signs), C-T+A (diamonds), and C-TA+TS (triangles). The thin dashed line shows the persistence forecast. The symbols are offset to allow better readability



(a)



(b)

Figure 7: Relative performance of the coupled experiments for the areas (a) Niño-1/2 (80° W to 90° W), Niño-3 (90° W to 150° W), Niño-4 (150° W to 160° E), and (b) EQ-1 (90° W to 130° W), EQ-2 (130° W to 170° W), and EQ-3 (170° W to 150° E). In all regions the latitudinal extent is 5° S to 5° N except in Niño-1/2 where it is 10° S to the equator. A forecast is considered better if it is more than 0.3K closer to the observed SSTA. Solid bars compare experiment C-T+A to C-TA+TS, and empty bars C-OI to C-TA+TS. Shown is the number of better minus the number of worse forecasts. Bars in the negative range mean that C-TA+TS performed on average better, bars in the positive range that it performed worse.

Such a measure is given in Fig.7 for 6 regions in the equatorial Pacific. It is of interest that, while the Niño regions and the EQ regions cover almost the same equatorial belt, longitudinal shifts of the boxes can sometimes result in significantly different skill estimates. First, experiments C-OI and C-TA+TS are compared (empty bars), that is the impact from the additional assimilation of sea level observations and the correction of salinity compared to the assimilation of temperature data only. On average C-TA+TS performs better than C-OI for lead times of more than 3 months, with exceptions for a lead time of 6 months in EQ-2 and Niño-1/2. On shorter lead times, C-OI performs relatively better compared to the longer lead times (e.g. C-OI is better than C-TA+TS in the EQ-1, EQ-2, and EQ-3 regions for up to 2 months ahead). Second, experiments C-T+A and C-TA+TS are compared (solid bars), that is the impact of the TH99 correction of salinity is estimated. Fig.7 a,b show that for almost all areas and lead times, C-TA+TS performs better than C-T+A. This implies that a good simulation of the salinity field is required to make optimum use of the altimeter data.

5 Conclusions

In this paper we have described our first attempts to assimilate altimeter observations in the framework of seasonal forecasting. The first experiments considered the ideal case of perfect model and perfect data, when altimetry was shown to make a marked improvement to the ocean analyses in some parts of the ocean but had a weak influence in others. The underlying ocean thermal and salinity structures were sometimes such that changes in the thermal field were compensated by changes in salinity in such a way as to induce only a weak signal at the surface. In such situations altimetry provided little useful information in isolation of other in situ data.

We then considered the use of real data. Four sets of ocean analyses and coupled forecast experiments using real data were considered. The first used altimetry in the absense of other in situ data and was compared with an existing assimilation scheme using only in situ thermal data. The altimeter was seen to provide useful information when compared to a control with no data assimilation. In the next two experiments we condered two different ways of combining altimetry with in situ data. Our aim in these two experiments was to assess the added value from combining altimeter data with our existing in situ temperature data assimilation system. The two new ocean analyses using the combined observations differed only in the way that salinity was estimated. In one experiment salinity is corrected based on each new temperature analysis and a preservation of the local T-S relationship, in the other based directly on the altimeter data, using vertical shifts of the background profile. These two different combined analyses can be thought of as the addition of altimeter data to two different in situ analysis systems, one of which adjusts salinity based on the temperature analysis and the local T-S relationship, and one of which makes no adjustment. The desirability of correcting salinity when updating temperature has nothing to do with altimeter data per se. Errors in the salinity field, however, can have an impact on the model sea level, and through that prevent an optimum use of the altimeter data.

In the Pacific, the differences between the four ocean analyses are usually quite small as measured by D20 and sea level. For both Niño-3 and Niño-4 the differences in D20 are generally less than 10m, which is considerably less than the scale of interannual variability which can be 70m in Niño-3 and 50m in Niño-4. The differences between OI with and without altimetry are generally less than 5m. The sea level differences in Niño-4 are proportionately larger than the D-20 differences: 5cm between experiments compared to interannual changes of 15cm. These differences may reflect changes arising from the different ways salinity is handled in the various experiments.

In the equatorial Atlantic Ocean the differences of upper ocean heat content between the four ocean analyses are larger than in the equatorial Pacific. In the equatorial Indian Ocean, differences are of intermediate magnitude. Even if the altimeter provides good data coverage in the Indian and Atlantic ocean, the success of the assimilation of sea level observations relies on the subsurface temperature structure of the model that is used to project the sea level information onto the temperature field. In the equatorial Atlantic and Indian it is more difficult than in the Pacific to obtain even a sparsly observed state of the subsurface temperature field. This should improve in the near future: Observations from the PIRATA array already provide information of the subsurface temperatures in the tropical Atlantic, and observations from ARGO floats should become available in near real time soon.

We turn now to the impact on forecasts. To what extent does the assimilation of altimeter data impact the forecasts? To answer this, four extensive sets of coupled model forecasts were intercompared. The results show that the rms-error of the forecasts initialized from the combined assimilation of sea level and temperature data is comparable to that of forecasts that use only temperature data. This is only one measure of skill, however, and if one adopts other methods of assessment there are hints that the forecasts using altimeter and temperature data and the correction of salinity based on the preservation of the T-S relationship have higher correlation in the region Niño-3. Tests using hit and false alarm statistics and relative performance statistics also suggest improvement for this set of forecasts, especially at longer lead times.

The inclusion of altimetry leads to smaller improvement in forecast skill than we had hoped. Two caveats need to be borne in mind, however. It is likely that model error is a major contributor to forecast error. The same coupled model is used for all the experiments. If model error dominates, then improvements in the creation of initial conditions will not be allowed to show a proper impact on the forecasts. A second consideration is that basically the same OI scheme is used throughout for assimilating in situ data. If the decorrelation scales or other parameters or procedures of the OI are badly chosen then the potential impact of adding altimetry will be reduced. Other experiments at ECMWF suggest that while the TAO array does a very good job of constraining the thermal structure when using the decorrelation scales used in this paper, there is emerging evidence that these scales are too large. If they were reduced then the altimeter might have more impact than seen in this work.

Finally one might hope that the assimilation of sea level observations would improve the forecasts in the tropical Atlantic and Indian Oceans, where few direct temperature observations exist. We have not examined the SST forecasts in this region in detail but it appears that both predictability and skill are limited.

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