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Altimetric assimilation in a Mediterranean general circulation model

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Abstract. Identical twin experiments are performed with a realistic Mediterranean general circulation model, assimilating surface pressure at the rigid lid (simulating altimeter data) in order to reproduce the three-dimensional circulation and water column structure. The first twin pair is forced with monthly varying wind stress and surface buoyancy, and the assimilation method of *Cooper and Haines* [1996] is used. Water columns are displaced vertically by an amount calculated to ensure that the surface pressure change required at assimilation time is reduced to zero at the seabed. The assimilation is successful with T, S and velocities having an error reduction of about 40% after 1 year, assimilating surface pressure every 15 days. A modification to this assimilation method is introduced in which a change to bottom pressure is calculated based on a balance between relative vorticity and stretching in the vertically displaced water column. A second identical twin experiment pair is forced with daily varying wind stress (from the European Centre for Medium-Range Weather Forecasts in 1992-1993) but still monthly varying surface buoyancy forcing. The variability in surface pressure is much greater when daily winds are used, with much of the high-frequency current signal being a barotropic response (with a strong signal at the seafloor). Applying the correct daily winds without any surface pressure assimilation can reduce the T, S and velocity errors by about 50% after a year, suggesting that much of the additional barotropic variability is deterministic. The additional assimilation of surface pressure every 20 days, now with the bottom pressure updated, leads to a further 30% error reduction. A further twin experiment with daily winds, in which the barotropic mode is allowed to converge first, before surface pressure assimilation begins, shows that a bottom pressure update during assimilation may be unnecessary if the correct wind stresses are known.

1. Introduction

Sea surface height from satellite altimeters is now available as a continuous stream of data revealing phenomena on the mesoscale and up to global scales. To use this data effectively it is vital to combine it with other data sets and to study the ocean evolution over a period of time. This can only be done by combining the data with numerical ocean circulation models. Perhaps one of the most difficult problems inherent in using altimeter data is the projection of the surface data downward to modify deeper currents and water properties which are needed for studying all but the fastest external waves. Over the past decade, several methods for vertical projection have been introduced.

One general approach is to assimilate surface current data into models without changing subsurface properties or currents in the expectation that the numerical model will transmit the data downward in some physically sensible manner. *Hurlburt* [1986], *Berry and Marshall* [1989], and *Holland and Malanotte-Rizzoli* [1989] all take this approach to assimilating surface pressure data (the models all have a rigid lid) with differing success, as determined by identical twin experiments. If the update is based on surface currents alone [e.g., *Hurlburt,* 1986; *Berry and Marshall,* 1989], the method only works for two vertical layers where large bottom friction can help the convergence of the second layer. On the other hand, if the update is based on the surface potential vorticity, as in the nudging method of *Holland and Malanotte-Rizzoli* [1989], then the subsurface flow is recovered for multilevel models because the effect of a surface potential vorticity change is felt immediately at depth through the invertibility principle.

A second general approach to vertical projection has been to use some statistical relationship to derive changes in subsurface quantities from the change in sea surface height. A statistical treatment is usually needed anyway in order to provide an error analysis on the incoming data, so this approach is appealing as it allows a great deal of physical information about the system to be subsumed into large sets

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of correlation coefficients. *De Mey and Robinson* [1987] developed the approach for vertical projection with a data set from the Gulf Stream region, correlating surface and subsurface pressure data. *Mellor and Ezer* [1991] and *Ezer and Mellor* [1994] developed a direct correlation between surface height anomalies and subsurface T, S properties in a model of the Gulf Stream. *Oschlies and Willebrand* [1996] base their analysis of a North Atlantic model on the vertical current correlations from which a new density structure is derived based on the thermal wind relation. These methods have all shown satisfactory results in twin experiment conditions when recovering subsurface currents. The biggest drawback is that the statistics are usually derived from a numerical model since observed data are not sufficient to provide reliable correlations. In addition, statistics can never function effectively in the more unusual situations or when trends are present which are often the most interesting periods to study. A review of some of the results obtained by these methods is presented by *Haines* [1994].

Haines [1991] introduced a new approach to vertical projection. The subsurface potential vorticity was conserved at the time when sea surface height data are assimilated. All surface current changes are then introduced through changes in surface potential vorticity alone. In some respects this is similar to nudging the surface momentum or vorticity equations as shown by *Haines et al.* [1993], although carrying out the change directly gives greater control of the physics of the model adjustment. The method worked particularly well in a quasi-geostrophic four-layer model and in a three-layered shallow water model *Haines et al.* [1993].

Cooper and Haines [1996] (hereafter CH) developed the conservation method for assimilating sea surface height data into a 21-level, eddy-resolving, Cox-Bryan general circulation model (GCM), [*Cox,* 1985]. The potential vorticity method above was extended to conserve all water properties by calculating a vertical displacement of water columns based on the update to surface heights. The result is a physically based assimilation method suitable for the large quantities of altimeter data which are now available from satellites. The method also does not resort to uncertain statistics to modify subsurface water properties.

In this paper the work by CH is extended in three important ways:

1. A much more complex model is now used, which involves seasonal variations in wind and buoyancy forcing and full salinity and temperature variations. This model, of the Mediterranean Sea, has been used recently to perform a variety of realistic modeling studies, e.g. *Pinardi and Navarra* [1993], *Roussenov et al.* [1995], *Haines and Wu* [1995], and *Wu and Haines* [1996].
2. A comparison study is made of assimilation in the presence of high- and low-frequency wind forcing. With low-frequency (monthly) forcing the CH technique works well. With the high-frequency (daily) forcing the ocean response has a much larger barotropic component. A modification to the CH method is introduced to permit a more barotropic (and hence a reduced baroclinic) adjustment to the model fields.
3. However, it is also shown that the high-frequency barotropic response may be deterministic which reduces the need for assimilating the barotropic component.

Section 2 introduces the assimilation by the vertical water displacement. Section 3 describes the GCM of the Mediterranean and presents a control run and twin assimilation experiment using monthly varying wind forcing over a 1-year period. Section 4 presents a modification to the CH assimilation technique to allow for deep pressure changes. This brings the method closer to the original total potential vorticity conservation method of *Haines* [1991] but updated to a primitive equation framework. Section 5 presents a second set of experiments in which the Mediterranean model is forced with daily varying winds taken from the European Centre for Medium Range Weather Forecasts (ECMWF) over the period 1992-1993. A "control" and a "parallel" run are compared with the same wind forcing starting from different model initial conditions without data assimilation. Then two experiments with surface height assimilation, one using the original CH technique and the other using the modifications of section 4, are reported. Section 6 provides discussion and conclusions.

2. Assimilation by Water Displacement

CH proposed that when changes to the sea level are required within a numerical model due to observations from a satellite altimeter, they be introduced by vertically displacing the incumbent water column at each location by a certain amount Δh which will vary horizontally over the domain. Such an assimilation technique is clearly consistent with conserving water properties, T, S, and any tracers, on isopycnal surfaces. It is also consistent with conserving the linear component of potential vorticity, which is simply the stratification, when considered as a function of potential density. Alternatively, it is also a local conservation of the amount of water with given water properties through most of the water column, except at the top and bottom. Figure 1 shows a schematic of this displacement process as it affects

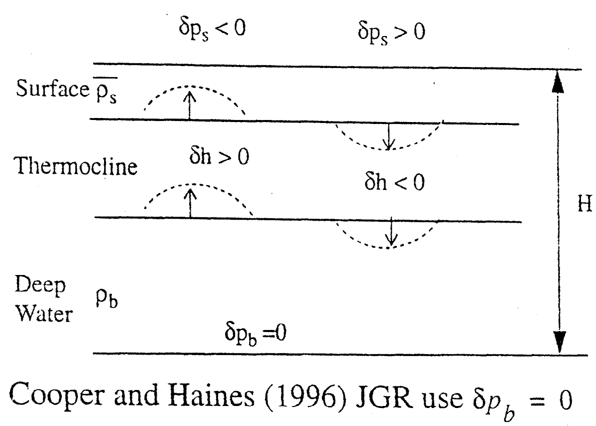

Figure 1. Illustration of the *Cooper and Haines* [1996] assimilation method for altimeter data. Water columns are displaced vertically in response to observed surface pressure changes. The deep pressure field is kept unchanged, which uniquely determines the displacement.

the thermocline, and possibly the surface water properties, in the presence of an observed sea level change. In order to calculate the vertical displacement field, it is necessary to consider the hydrostatic balance condition:

$$\Delta p_s + g\int_{-H}^{0} \Delta\rho dz = \Delta p_b \quad (1)$$

where Δp_s and Δp_b are the changes in surface and bottom pressure, respectively, and $\Delta\rho$ is the change in the water column density as a function of depth. In the above equation, $\Delta\rho$ could be an arbitrary three-dimensional (3-D) field; however CH chose to calculate $\Delta\rho$ within each water column from a two-dimensional (2-D) vertical displacement Δh. The above equation can then be written approximately as follows (refer to Figure 1):

$$\Delta p_s + g(\rho_b - \overline{\rho_s})\Delta h = \Delta p_b. \quad (2)$$

Now $\overline{\rho_s}$ is the mean potential density of the surface waters which are added or removed from the water column, and similarly, ρ_b is the potential density of the bottom waters which are effectively exchanged in the vertical displacement process. The overbar on the surface density reflects the fact that there will likely be vertical structure at least within the surface waters. The second term represents the change in the weight of the entire water column after a vertical displacement of an amount Δh. It is clear that, except in cases of a neutrally stable water column, a positive Δh, or lifting of the water column, will increase the weight (more deep dense water) and a negative Δh will decrease the weight (more light surface waters). In practice, the change in weight of a water column is determined more accurately than implied by (2) because each water column (T, S) structure is first splined in the vertical and any change in surface water properties associated with the vertical displacement is accurately assessed; see CH for details.

In order to close the problem and to calculate Δh (and hence $\Delta\rho$), CH make the assumption that $\Delta\ p_b\ =\ 0$. In this case the vertical water column displacement entirely compensates for the change in the sea level such that no trace of the new surface pressure signal reaches the deep ocean. This was described as choosing a "level of no change to the motion" since if pressure does not change at depth, there are also no geostrophic current changes. With this assumption the problem of how much vertical displacement is required for each water column is uniquely defined, except when the required surface pressure change is very large. In these cases the mean water density cannot be changed enough simply by vertical displacement. Such situations can occur in cases of weak stratification, for example, toward polar regions, and may imply that the deep pressure field should be changed. Finally, the modified density structure is used to calculate a geostrophic update to the current fields.

CH show that this assimilation method works well in a twin experiment with a 21-level eddy-resolving, rigid-lid, primitive equation model in a box. One year of assimilation of surface pressure every 9 days permits good recovery, even of the deep subthermocline currents (which show 60% error reduction after 1 year although they are not improved by any single assimilation step). The model used by CH has rough bottom topography which tends to reduce deep currents [*Böning,* 1989], and the surface wind stress was also constant in time, which reduces the possibility of a more variable barotropic circulation. In the next section we show how this assimilation scheme performs in a much more complex model of the Mediterranean Sea.

3. A Mediterranean Twin Experiment With Monthly Winds

To test the displacement assimilation technique out in a more realistic model, a twin experiment was performed with a Cox/Bryan model of the Mediterranean Sea; for model details, see *Roussenov et al.* [1995], and *Wu and Haines* [1996]. The model has 1/4° horizontal resolution and 19 vertical levels with fairly realistic topography. It is forced with monthly varying wind stress with a repeating yearly cycle taken from 5-year averaged ECMWF 10m winds for the period 1988-1993. The surface forcing is a Haney relaxation to T^*, S^* taken from National Oceanographic Data Center (NODC) data, with some modifications in the water formation regions; see *Wu and Haines* [1996]. This version of the model has a *Gent and McWilliams* [1990] scheme introduced to provide subgrid scale dissipation; see *Haines and Wu* [1997].

The model has been spun up from *Levitus* [1982] data for 46 years. For the following year the twin experiment truth/control run is performed with surface pressure at the rigid lid dumped every 15 days. This surface pressure data will be used for assimilation and the full history files used for judging the assimilation success in reproducing currents and water properties at deeper levels. Figure 2 shows the surface pressure (Figure 2a), and the salinity on the 28.95 isopycnal surface (Figure 2b), at the beginning of the truth run. Some mesoscale features can be seen both in the surface pressure and in the water property distributions, particularly along the path of the Algerian current where baroclinic instability occurs. Such features vary from year to year, despite the repeated forcing, and assimilation of surface pressure is required to correct not only the surface pressure and currents but also the subsurface property distributions such as in Figure 2b.

To perform an assimilation experiment, the initial conditions are taken at year 51 of the same run as described above. Assimilation of the truth pressure should therefore cause year 52 to begin to reproduce the truth year 47. An additional problem was encountered when trying to assimilate by vertical displacement during the summer months in the presence of a seasonal thermocline. To avoid lowering the sharp gradients from the seasonal thermocline into the water column, only the water below the third level, i.e., at 80 m and below, is displaced by assimilation. Water properties above are not altered. This is done for all experiments at all times of year because it will not influence the results in winter in any case. Initially, we present results from the first assimilation step before the model integration is continued

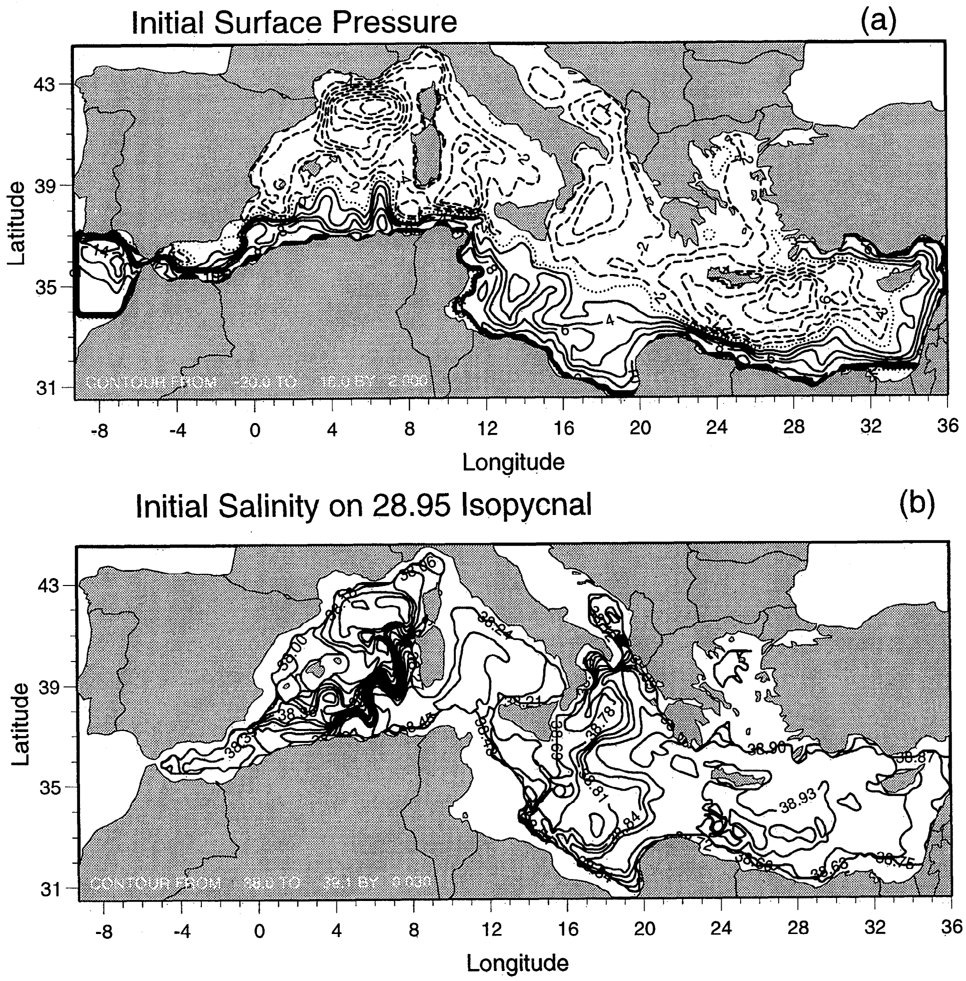

Figure 2. (a) Initial surface pressure distribution at the start of the monthly forced assimilation experiment. Units are 100 Pa; thus contour interval is 2-cm equivalent sea surface height. (b) Salinity distribution on the 28.95 isopycnal surface. Units are practical salinity units.

in order to discuss changes introduced by the assimilation scheme. This is followed by results from repeated intermittent assimilation over a 1-year period.

3.1. Single Assimilation Step

Figure 3a shows the surface pressure difference between the truth and the assimilation run at the start. This field, Δp_s, is to be assimilated according to section 2 (no errors are involved in the twin experiment). Figure 3b shows the vertical displacement of each water column in response to Figure 3a. Note, for example, the changes SW of Sardinia where a large Algerian current meander (Figure 2a) needs to be removed by the assimilation and a tripole of surface pressure and vertical displacement updates are introduced. Figure 4 shows the temperature (Figure 4a), salinity (Figure 4b), and velocity (Figure 4c) errors as functions of depth. The solid lines show the initial error profiles, the dashed lines show the profiles after a single assimilation step, and the dotted lines show error profiles after 1 year of intermittent assimilation. It can be seen that within the thermocline, there is a considerable improvement, especially in the velocity error, immediately after a single assimilation. There is even a small improvement in velocity error at depth. The property errors show a mixed response within the thermocline. Temperature errors get slightly worse at 120-m depth but improve below that. Salinity errors improve down to 200 m but get slightly worse below that. In an RMS sense over the whole basin the density error decreases by 20% at this time. This reflects about the best that can be achieved by limiting the assimilation to a vertical displacement. Undoubtedly, many differences between the model fields have come about due to changes in horizontal water redistributions. To allow these to be corrected, the model must be run forward.

An additional comment on current updating is now necessary. All current updates are geostrophic; however, direct finite differencing from the pressure field changes can lead to boundary problems in a complex basin such as the Mediterranean. Updating the barotropic stream function, ψ, is particularly hazardous and eventually problems were overcome by solving the following equation with a relaxation method:

$$\nabla.(\rho_0 f \nabla \Delta\psi) = \nabla.(H \nabla \Delta p_s) + \nabla. \int_{-H}^{0} g(z+H) \nabla \Delta\rho dz.$$

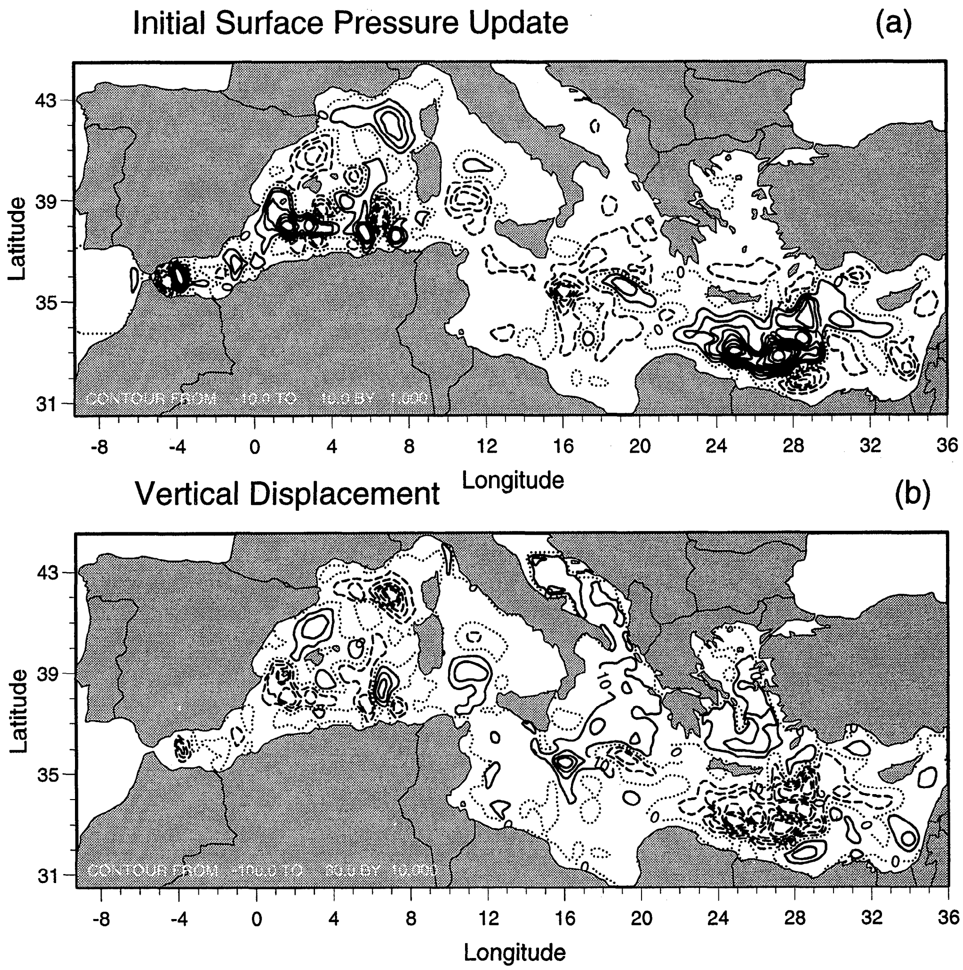

Figure 3. (a) Initial surface pressure update from monthly wind-forced assimilation run. Contour interval is 1-cm sea surface height equivalent. (b) Vertical displacement field with contour interval of 10 m. Positive is a raising of the water columns.

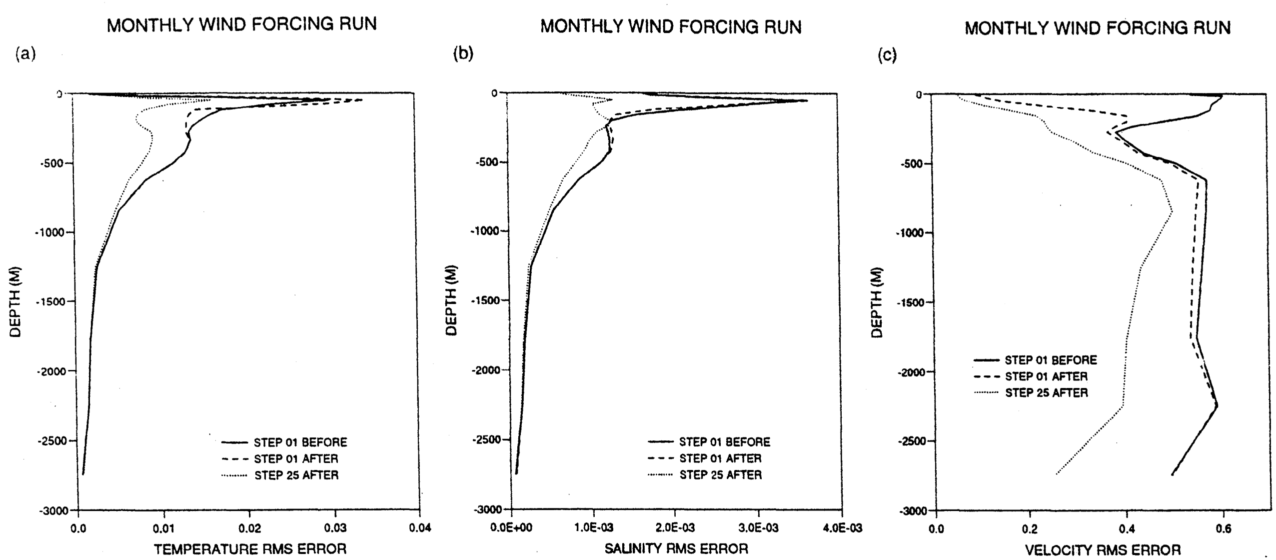

Figure 4. Vertical error profiles of (a) temperature, (b) salinity, and (c) velocity. Before any assimilation (solid lines), after single assimilation (dashed lines), and after 1 year of assimilation (dotted lines). Units are degrees Celcius (Figure 4a), practical salinity units (Figure 4b), and centimeters per second (Figure 4c).

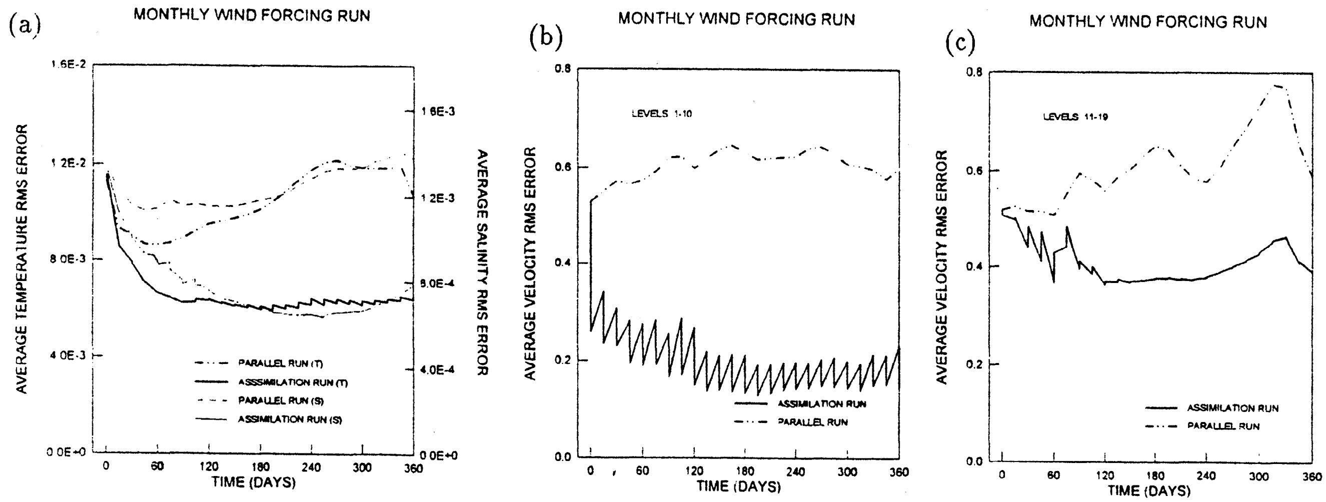

Figure 5. Solid lines show, for the monthly wind-forced twin experiment, the errors in basin-averaged (a) T, S, (b) velocity above 300 m, and (c) velocity below 300 m. Each plot also shows parallel run results (broken lines) from the same initial conditions with no data assimilation. Units are degrees Celcius and practical salinity units (Figure 5a) and centimeters per second (Figures 5b and 5c).

This allows for changes in circulation around islands and satisfies all appropriate boundary conditions. Since total current updates can be calculated from the surface pressure and the thermal wind equation, the barotropic currents from $\Delta\psi$ can be subtracted to give the baroclinic currents. The relationship between stream function and surface pressure has been discussed in detail by *Pinardi et al.* [1995] from which the above equation, which assumes geostrophy, has been adapted.

3.2. Intermittent Assimilation

The assimilation procedure was repeated every 15 days over the period of 1 year. Each 15 days the model is stopped, the new surface pressure field from the truth run is assimilated, and the model is restarted. To assess the convergence of the model run, RMS errors are calculated before and after each assimilation step. Figure 5a shows the evolution of the RMS T and S errors over the whole model domain. Levels have not been weighted according to depth, which emphasizes errors in the upper water column. The thick continuous line shows T errors, with the scale on the left, and the thin continuous line shows S errors, with the scale on the right, for the data assimilation run. The broken lines are T, S errors from a parallel run which show the year 52-error evolution without any assimilation of surface pressure. The small reduction in the parallel run errors that occurs in the first 50 days is due to the onset of winter when the mixed layer is deeper and its properties are strongly surface controlled; however, there is no significant trend for convergence or divergence of the parallel run fields over the 1-year period. The assimilation run, however, shows a very rapid decrease in errors for both T and S which persist through to the end of the experiment.

Figure 5b shows the RMS velocity error for levels above 300 m (levels 1-10), and Figure 5c shows the velocity errors below 300 m (levels 11-19). Both show a clear reduction but with the subthermocline velocities (Figure 5c) mainly showing improvement during the model run rather than at assimilation times. Again, the broken lines show the parallel run errors with no assimilation. Referring back to Figure 4, the dotted lines show the final errors in T, S and velocity as a function of depth. Note particularly that the velocities well below the thermocline are improved by about 30-50% by the end of the 1-year run.

Figure 6 shows fields of salinity error on the 28.95 isopycnal surface before the beginning of assimilation and after 1 year of assimilation. Because the assimilation occurs with only vertical water displacements, this error will not change at assimilation time. However, the fact that it has reduced by the end of the 1-year assimilation run shows that the improvement in currents induced by the pressure assimilation is able to redistribute water properties horizontally within the basin and along isopycnal surfaces, leading to an eventual improvement in horizontal property distributions. The RMS salinity errors on this isopycnal are reduced by 53% by the end of the year. This result emphasizes the distinctive role that the GCM itself plays in the convergence process, which other assimilation methods often fail to demonstrate. Given the correct encouragement to produce accurate velocity fields, especially at the mesoscale, the tracer fields of the model can actually converge rather than diverge with time.

A problem may arise in the use of the above assimilation technique when the observed sea surface height anomalies reflect pressure changes throughout the water column. Of course, this is commonly encountered when considering the tides but it is assumed that these can be corrected for, and removed from the sea level anomaly signal, before assimilating into a circulation model. However, large-scale barotropic pressure changes may also be generated by nontidal forcing such as atmospheric storms. In the next section we consider modifying the CH method to permit larger barotropic changes.

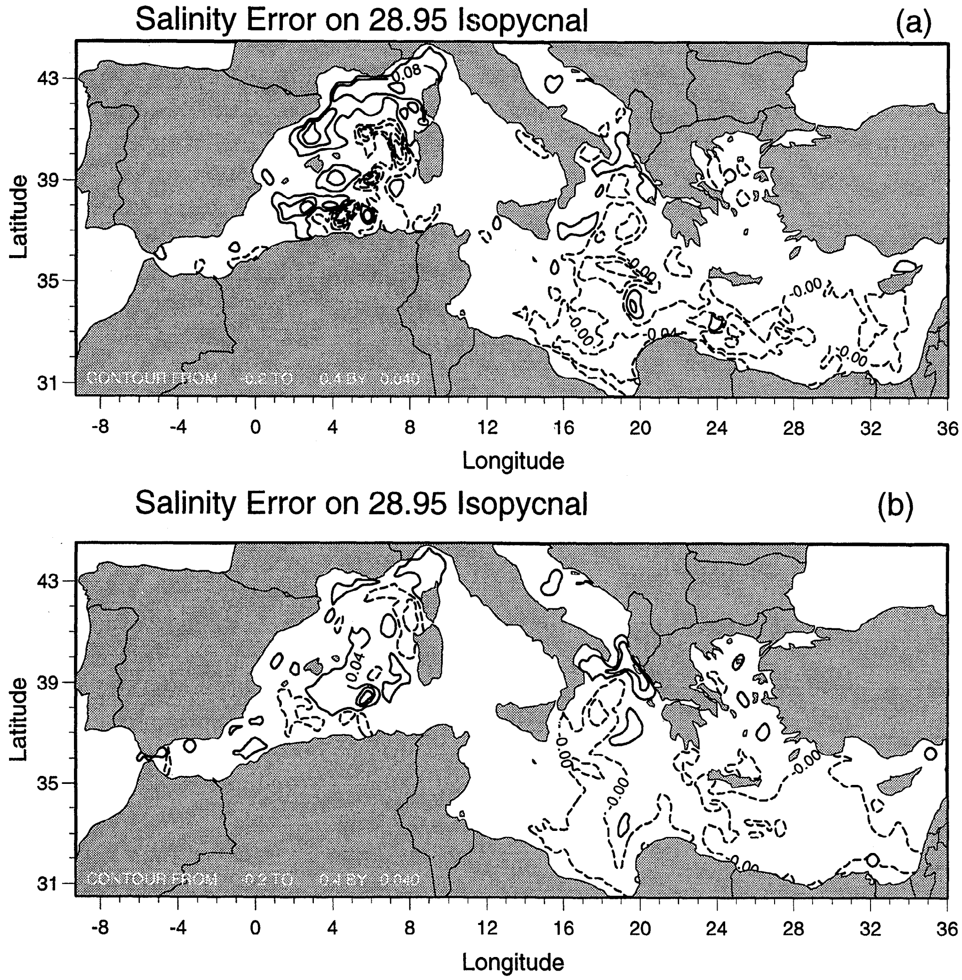

Figure 6. (a) Initial salinity errors on the 28.95 isopycnal from monthly wind-forced assimilation run. (b) Final salinity errors on the 28.95 isopycnal after 1 year of intermittent surface pressure assimilation every 15 days. Contour interval is 0.04 psu.

4. Deep Pressure Corrections

There are problems with the assimilation method described in section 2. One in particular was noted by CH, that the condition of no change to the deep pressure may not be realistic because it greatly restricts changes to the barotropic mode. Imposing this condition may produce unrealistic vertical displacements and hence unrealistic adjustments to the baroclinic modes. Incorrect modification to the baroclinic modes by vertical displacement is a serious problem because these modes take longer to adjust and efforts should be made to ensure that they are not adversely affected by assimilation. Despite this drawback the results of the previous section, using a much more complex model than that of CH and with seasonal forcing, indicate that the CH method still works well. The reason is that the model run with only monthly varying winds always displays very weak bottom velocities which are kept small by the topography. Hence no serious adjustment to bottom pressure is required at assimilation times. This is not always the case as experiments in section 5 will show.

To make the assimilation more general we wish to relax the constraint on changes the deep pressure. Figure 7 illustrates one method of achieving this. In the layer models of *Haines* [1991] and *Haines et al.* [1993] the full potential vorticity, including relative vorticity, was conserved for all layers below the top layer during data assimilation. The CH work does not consider relative vorticity changes at all and

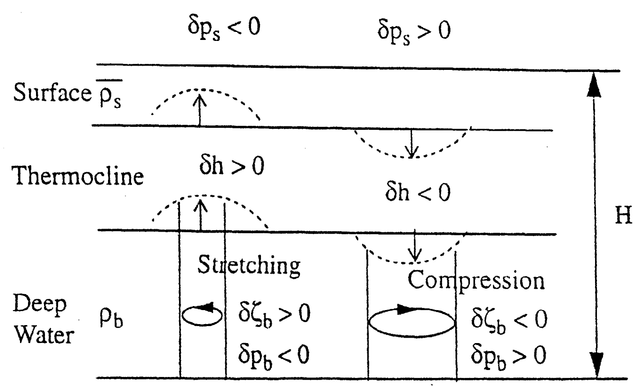

Figure 7. Illustration of the modified *Cooper and Haines* [1996] assimilation method. Water columns are displaced vertically as in Figure 1, but the deep pressure field is modified according to the relative vorticity induced by the stretching/compression of water columns.

only preserves isopycnal thickness. It is difficult to account for changes in relative vorticity on all isopycnal layers; however, the vertically integrated relative vorticity change can be related to the overall compression or extension of the water columns determined by the vertical displacement field Δh.

Figure 7 illustrates this compressing or stretching of water columns which occurs when a vertical displacement Δh is introduced. The appropriate equation relating the changes in relative vorticity, $\Delta\overline{\zeta}$, with Δh is given by,

$$\Delta\overline{\zeta} = \frac{f}{H}\Delta h. \tag{3}$$

The overbar indicates a vertical average and H is the total column depth. This change in relative vorticity occurs throughout the water column and therefore also in the deep ocean. For simplicity we consider that this can be related geostrophically to the deep pressure change (rather than to the vertically averaged pressure change) and so;

$$\Delta\overline{\zeta} = \frac{1}{\rho_0 f}\nabla^2\Delta p_b. \tag{4}$$

It is simple to combine (2), (3) and (4), to eliminate Δh and obtain an expression giving the deep penetration of the surface pressure signal:

$$\frac{gH(\rho_b - \overline{\rho}_s)}{\rho_0 f^2}\nabla^2\Delta p_b - \Delta p_b = -\Delta p_s. \tag{5}$$

The approximations above are acceptable because we are really only looking for a first-order correction to the $\Delta p_b = 0$ condition of CH which is reasonably robust and can be calculated based on dynamical rather than statistical grounds. Calculation with a more accurate version of (4), using vertically averaged pressure changes, does not make any significant difference to the results.

This Helmholtz equation can be solved to determine Δp_b. Consider some examples. (1) In cases of weak stratification such that $\rho_b - \overline{\rho}_s$ is small, the Laplacian term in equation (5) will be small and the assimilation will be more barotropic with $\Delta p_b \sim \Delta p_s$. (2) In cases of large-scale Δp_s the response will also be more barotropic. (3) In cases of strong stratification and/or small-scale Δp_s, then $|\Delta p_b| < |\Delta p_s|$. The expression

$$R^2 = \frac{gH(\rho_b - \overline{\rho}_s)}{\rho_0 f^2}$$

is essentially a squared Rossby deformation radius for each water column. Typical values in the Mediterranean are $R \sim 50 - 100km$.

Having solved for Δp_b, it remains to determine Δh which controls the baroclinic response. This involves using the vertically integrated hydrostatic equation (1) for each water column. No approximations are made and the vertical displacements are determined accurately with the hydrostatic equation being integrated with the same finite difference scheme used in the model. A careful splining is used to determine the water column density profiles as described by CH and *Cooper* [1995]. The essential aspect of assimilation by water displacement introduced in CH is thus preserved. The only difference is that the changes are permitted to be more barotropic, according to the original stratification present in the assimilation region and the scale of the assimilated surface pressure anomalies. The vertical water column displacements are reduced while maintaining independence from the need to use statistical correlations.

The above derivation is mainly useful for two reasons. It offers a more complete primitive equation version of the original *Haines* [1991] full potential vorticity conserving assimilation method. It also provides a spatial "filter" which permits a sensible reduction in the baroclinic (water column displacement) adjustments during assimilation. We do not claim that it will permit a really accurate estimation of the barotropic mode component of the sea surface height signal. It has been shown by *Cummins* [1991], for example, that for the vertically integrated flow, local relative vorticity changes are not normally the response to imposed local stretching, as implied by the balance in (3). Instead, stretching is more likely to be related to horizontal motion in a topographic Sverdrup balance. However, this result may change if the forcing is separated more distinctly by frequency. More work is clearly needed in this area, with altimetric assimilation particularly in mind. Nevertheless, in the next section we consider an experiment with daily varying winds where a stronger barotropic mode signal must be accounted for in order to make a successful assimilation, and the above method proves to be useful.

5. A Twin Experiment With Daily Wind Forcing

If the new method for calculating bottom pressure changes is implemented in the twin assimilation run described in section 3, it is found that the changes to the bottom pressure calculated from (5) are much too large, suggesting that the relative vorticity/stretching relationship is a poor assumption for the slow variations in this run and it is better to ignore relative vorticity as done by CH. If the new bottom pressure updates are nonetheless used, convergence is still obtained despite the large barotropic errors introduced at the early assimilation times. However, in order to test the new assimilation method in a more realistic experiment, we used the same model but with a daily varying wind stress.

The parallel run described in section 3, where no assimilation is performed, was run on further until August 1 in year 53. From this time the model was forced by the daily varying ECMWF wind stresses from August 1 1992, until December 31 1993. The surface buoyancy forcing was not changed and still reflects climatological conditions varying on a monthly timescale. The period from September 10 onwards was used as the truth/control run for assimilation with data stored every 20 days. The switch to 20 days from 15 in section 3 was due to parallel experiments with TOPEX/ERS 1 data where maps were available every 20 days throughout this period *Drakopoulos et al.* [1996].

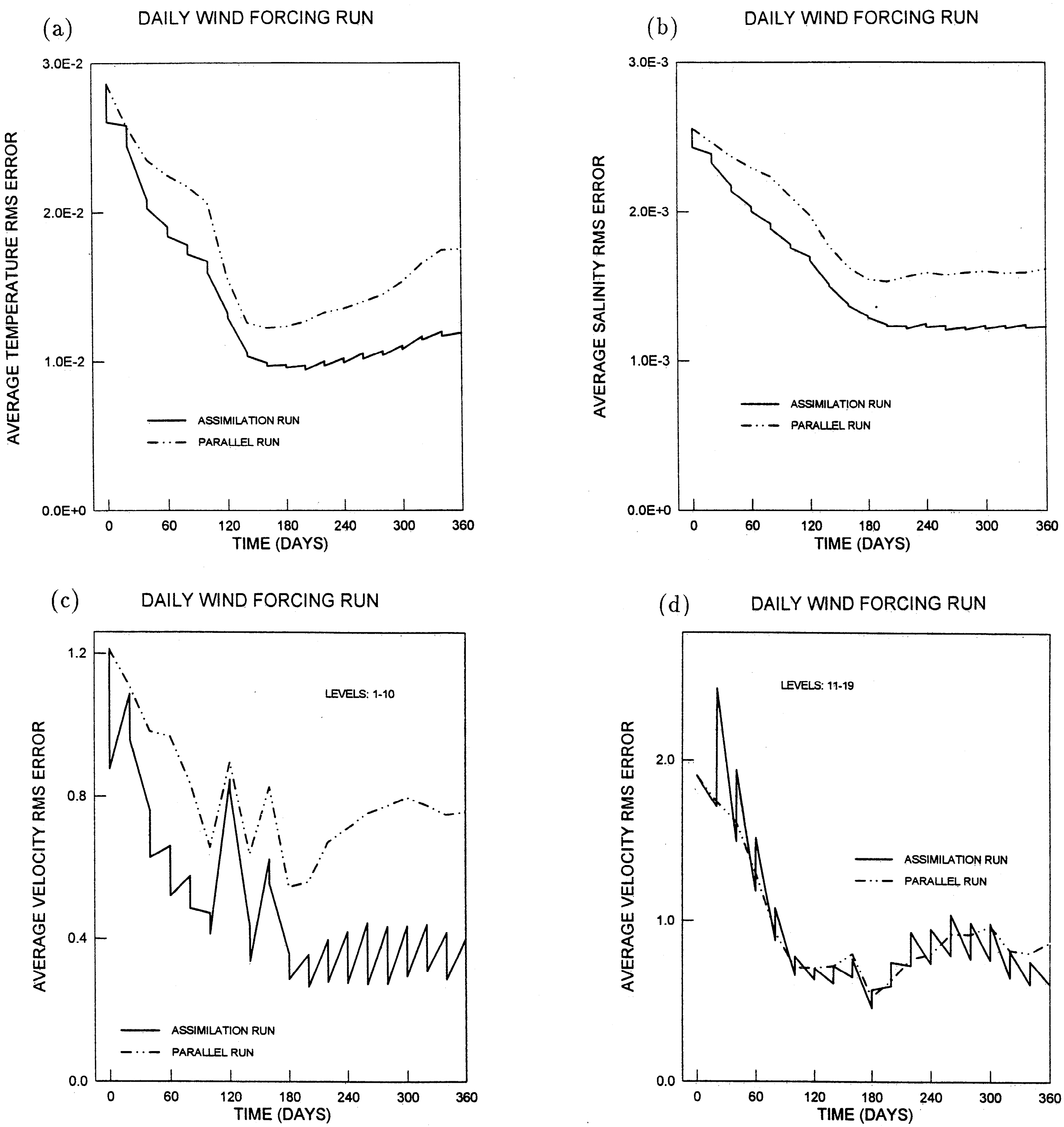

Figure 8. Solid lines show, for the daily wind-forced twin experiment 2, the errors in basin averaged (a) T, (b) S, (c) velocity above 300 m, and (d) velocity below 300 m. Each plot also shows parallel run results (broken lines) from the same initial conditions with no data assimilation. Units are degrees Celcius (Figure 8a), practical salinity units (Figure 8b), and centimeters per second (Figures 8c and 8d).

The modeled variability in sea surface height over the Mediterranean domain is much greater with daily wind forcing and is more consistent with the amplitude of the observed sea level anomalies in the Mediterranean; see *Larnicol et al.* [1995]. The kinetic energy of this control run is 2-3 times greater than with monthly wind forcing. To obtain the initial conditions for the assimilation experiments, the model was run on for 2 further "years", each of 360 days, using daily winds from September 10 1992 to September 4 1993, inclusive. The September field at the end of this time provides the new initial conditions. The sudden switch in the winds each September will turn out to be important because the ocean "remembers" the wind forcing for much longer than the few days decorrelation time of the winds themselves.

Three experiments will be described. The first is a "parallel" run in which the model is simply run for another year, with no surface height assimilation, to test the convergence time of the model fields due to the winds alone. The other two are assimilation experiments, with and without modifications to the deep pressure.

5.1. Experiment 1: Parallel Run

The broken lines in Figures 8a and 8b show the evolution of the RMS T and S differences (i.e., errors), respectively, over the whole model domain. The broken line in Figure 8c shows the RMS velocity error for levels (1-10) above 300 m and the broken line in Figure 8d shows the velocity errors below 300 m (levels 11-19).

Note that the initial errors in all the fields are considerably greater (by a factor of 2-3) than in the previous experiment (Figure 5) due to the high-frequency wind forcing. Unlike Figure 5, there is a considerable reduction in the errors in all quantities between the beginning and the end of the parallel run, by about 40% for T, S and the upper water column velocity and by about 50% for the deeper velocities. These reductions occur mainly within the first 100 days at the start of the parallel run, suggesting that this is the length of time required for the model to undergo barotropic adjustment to adapt to a new wind forcing (the time may be longer for a larger basin than the Mediterranean). The smallest density errors are reached in the winter period around day 180, due to the surface buoyancy forcing, after which the errors rise a little. It is interesting to note that the final errors after the 1-year parallel run are only a little higher than the constant parallel run errors from the monthly forced experiment depicted in Figure 5. This suggests that the signal component excited by the high-frequency winds may be being reasonably well reproduced and may be quite insensitive to the initial model flow provided the correct winds have been applied for long enough (about 100 days).

5.2. Experiment 2: Assimilation With Deep Pressure Updates

The initial state for these assimilation runs has been forced most recently with winds from September 1993, but there is a sudden switch to September 1992 winds at the beginning of the run. Figure 9a shows the surface pressure difference at the start of the assimilation. This is the Δp_s which

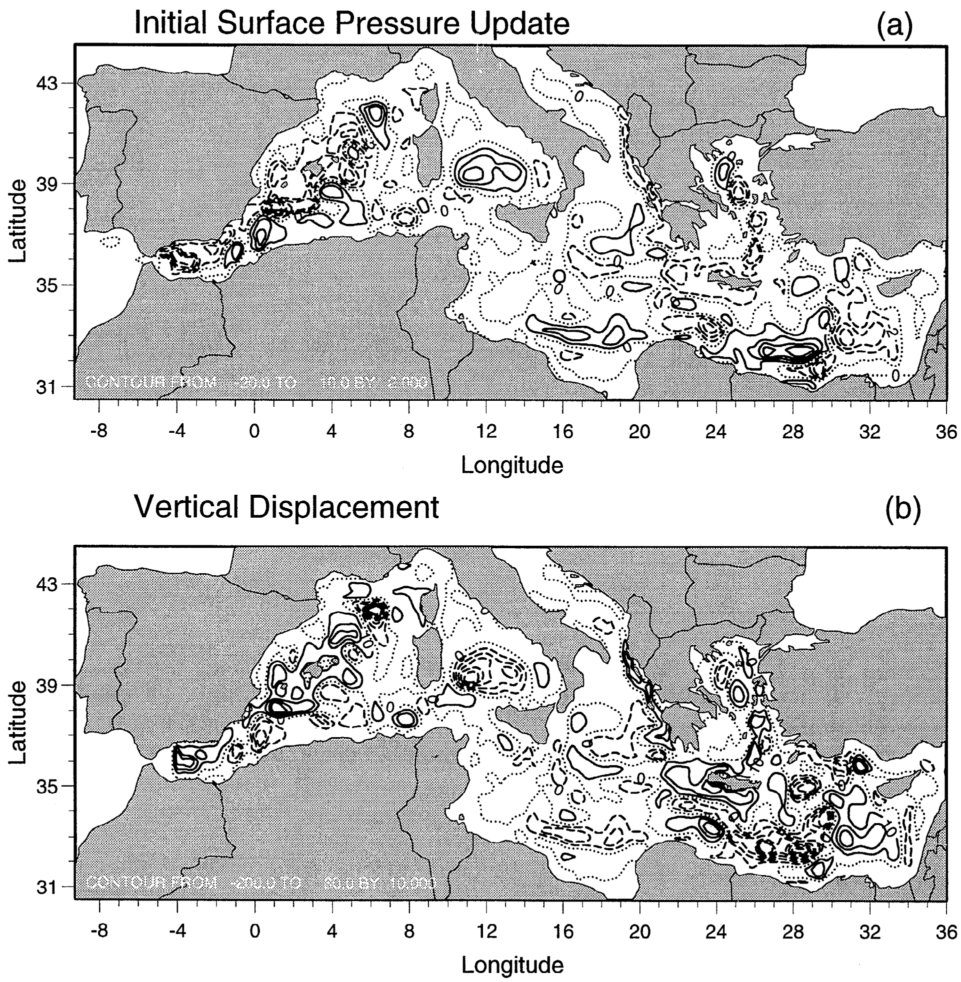

Figure 9. (a) Initial surface pressure update from daily wind-forced assimilation run. Contour intervals are 2-cm sea surface height equivalent. (b) Vertical displacement field with contour interval of 10 m. (c) Initial transport stream function update. (d) Initial streamfunction error before assimilation. Contour intervals in Figures 9c and 9d are 2 Sv.

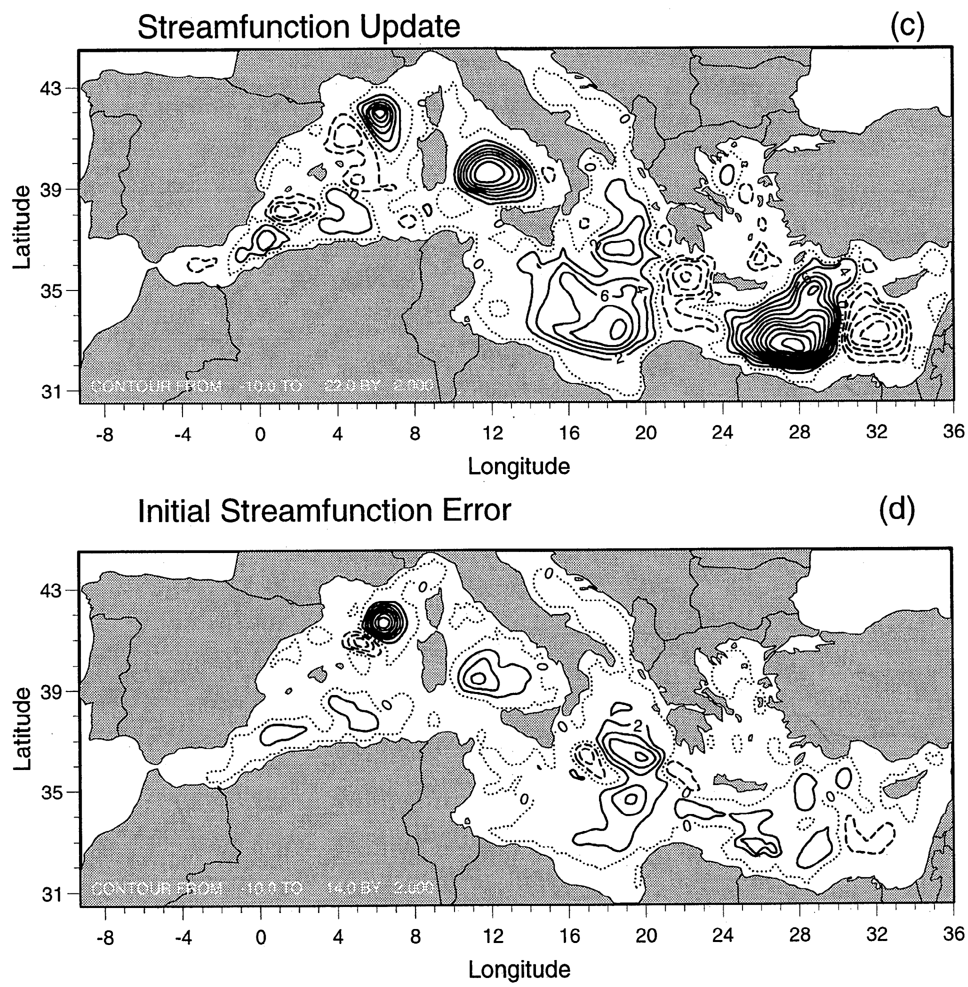

Figure 9. (continued)

will be assimilated according to (1),(2), and (5), but note that it is considerably larger than in the previous twin experiment (Figure 3a), reflecting the increased variability in surface pressure when daily winds are used.

Figure 9b shows the vertical displacement field Δh. Note particularly that this field is of a similar magnitude (up to 200 m) to the first twin experiment (Figure 3b), despite the much larger changes in surface pressure. This reflects the fact that a good deal of the additional surface pressure variation is deflected to the barotropic mode and does not require displacement of water columns. This "filtering" of a barotropic component is necessary because an attempt to reduce the surface pressure updates in Figure 9a to zero at the bottom would produce totally unrealistic water displacements.

To emphasize the barotropic component, we look at the update to the barotropic mass transport stream function. Figure 9c shows the update to ψ, while Figure 9d shows the actual error in ψ before the assimilation. It can be seen that in most areas the sign of the stream function correction is right but the magnitude is a little too large (although it is too small in the large eddy west of Corsica). The very worst area is the Levantine where large changes are introduced which are unnecessary. In this region the baroclinic current errors must be more prevalent at this time, reflecting the large changes in water column structure that can be present from year to year depending on the intermediate water dispersal route.

The tendency for the barotropic current updates to be, on average, too large was noted by *Haines* [1991]. Any mechanism which tends to reduce the currents in the deep bottom layers will lead to this result. In the work by *Haines* [1991], Ekman friction was the cause, while here it is the rough bottom topography and the pressure torques which tend to reduce deep currents. Equation (3) relating relative vorticity to stretching does not account for this. However, as shown in Figure 9, the method works effectively as a filter because, by overestimating the barotropic mode, unrealistic changes to the baroclinic structure are avoided. If the constraint of no change to bottom pressure used in CH were introduced in this run, the vertical displacements would be considerably greater than in Figure 9b, especially during the winter months, which would lead to unrealistically noisy hydrography.

For intermittent assimilation every 20 days the solid lines in Figures 8a and 8b show the RMS T and S errors, respectively, over the whole model domain, for comparison with the parallel run errors, (broken lines). The assimilation run errors fall more rapidly than the parallel run errors, and by the end of the assimilation year this run has a 25-30% T, S

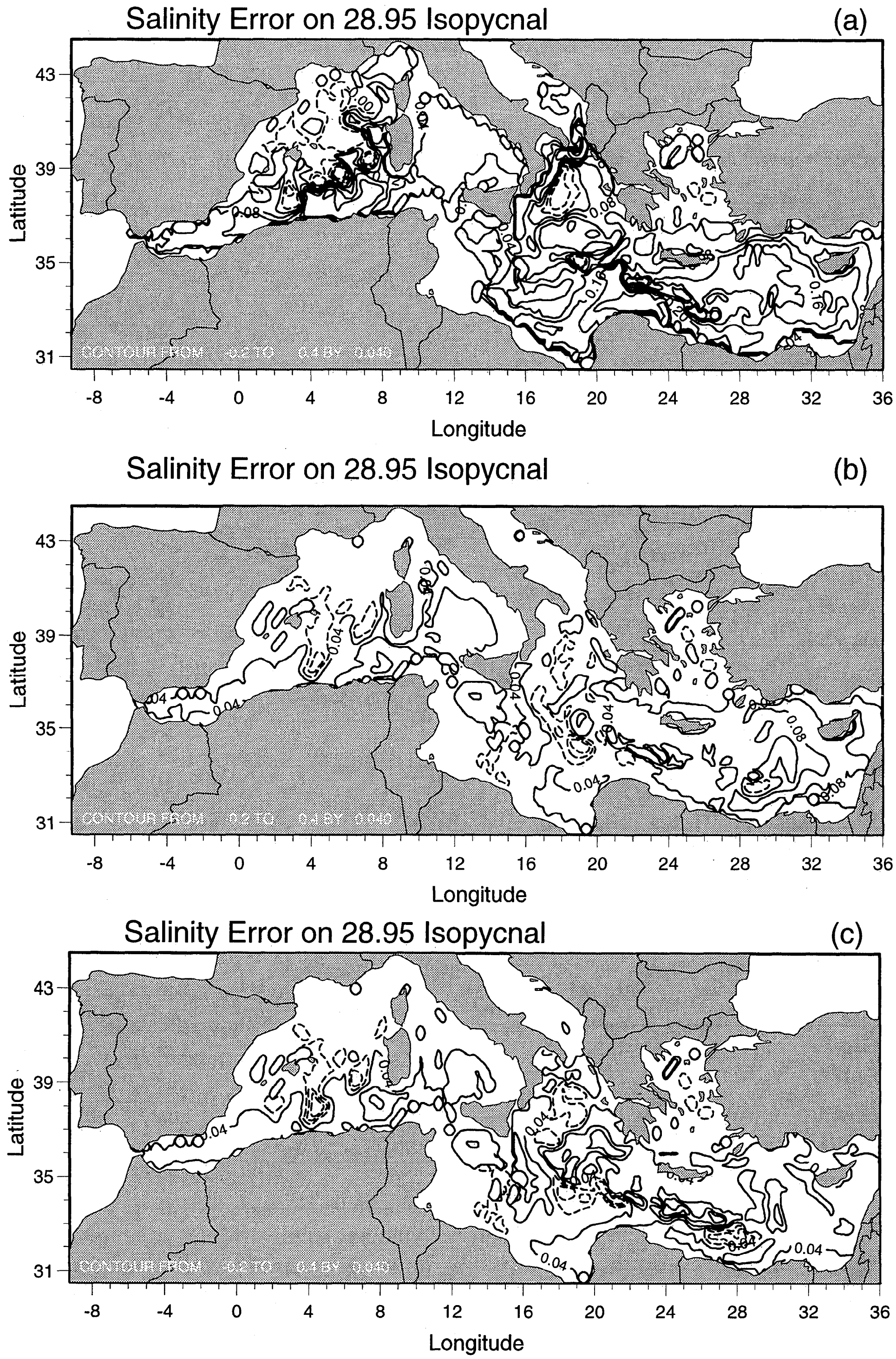

Figure 10. (a) Initial salinity errors on the 28.95 isopycnal from daily wind-forced assimilation run. (b) Final salinity errors on the 28.95 isopycnal after 1 year of intermittent surface pressure assimilation every 20 days. (c) Salinity errors on the 28.95 isopycnal after 1 year of parallel run with no data assimilation but forced with the truth daily winds. Contour interval is 0.04 psu.

error reduction over the parallel run and a 60% improvement over the initial conditions.

The solid line in Figure 8c shows the RMS velocity error for levels (1-10) above 300 m, and the solid line in Figure 8d shows the velocity error below 300 m (levels 11-19). Figure 8c shows a 60% reduction in the velocity errors within the thermocline, over that attained by the parallel run, or a 75% reduction over the initial conditions. The large velocity errors in the upper water column in the middle of the assimilation are localized in the Aegean Sea and are due to a large storm which passed through in January 1993. The model was not able to cope with this extreme event well, producing very noisy fields of barotropic stream function. Nevertheless, this does not appear to have disrupted the longer-term convergence. The subthermocline velocity errors, in Figure 8d, are less convincing in comparison to the parallel run errors, although some improvements are seen. Again, they mainly show improvement during the model run rather than at assimilation times, as in Figure 5c.

Figure 10 shows salinity error on the 28.95 isopycnal surface before the beginning of assimilation (Figure 10a), and after 1 year of assimilation (Figure 10b). The errors at the end of the parallel run without assimilation are also shown (Figure 10c). The reduction by the end of the 1-year assimilation run again shows that the improvement in currents induced by the surface pressure assimilation is able to redistribute water properties horizontally within the basin, leading to an eventual improvement in property distributions. The parallel run shows a 75% error reduction, while a further 15% reduction occurs with data assimilation. The modified CH method permits successful assimilation, even in cases of very large variations in surface height, by filtering off a barotropic signal. However, given the rapid reduction in the barotropic errors seen in the parallel run it is worth investigating if an alternative assimilation procedure is possible.

5.3. Experiment 3: Assimilation After Deep Pressure Convergence

The deep velocity (and pressure) fields converge well without any assimilation after about 100 days of forcing with the correct daily winds (Figure 8d, broken line). We performed an experiment assimilating surface pressure from the control run only after 100 days of forcing with the true winds. The original CH assimilation method was used on the assumption that the barotropic response should already have converged to a large extent. Figures 11a, 11b, 11c, and 11d (solid lines) show the evolution of the T and S and upper and lower level velocity errors, respectively, for this run. The parallel run of Figure 8 is also shown and the two runs are coincident for the first 100 days before assimilation is performed. The surface pressure differences to be assimilated in this case are more comparable in magnitude to those in the monthly wind forced run and the original CH assimilation method is clearly successful with convergence of all the error fields at the end of the run. The results here are slightly better than in experiment 2. On this evidence it may be that the real advantage of the assimilation method in section 4 is that it permits assimilation even in cases when the barotropic mode has not converged and it may be useful, for example, where there are errors in the wind field which prevent full convergence of the wind-forced signal.

6. Discussion

This paper describes the results of twin experiments with surface pressure assimilation using a sophisticated ocean circulation model of the Mediterranean with seasonally varying surface forcing of winds and buoyancy. The first experiment is carried out with monthly varying wind stresses and uses the *Cooper and Haines* [1996] surface pressure data assimilation method in which each water column is vertically displaced so as to leave the bottom pressure unchanged at assimilation time. The twin experiment is successful with thermocline and subthermocline currents improving and water property (T, S) distributions improving within the thermocline. When viewed on level surfaces, there is an immediate change in T, S as the assimilated pressure is used to deduce a vertical displacement which varies across the model. However, when viewed in isopycnal coordinates, the horizontal distribution of T, S is not changed at all on the assimilation step. However, when intermittent assimilation is performed, the new advecting currents lead to improved T, S (and would perhaps for other water property distributions) on isopycnal surfaces after a year of model integration.

A modified assimilation technique is introduced which allows the bottom pressure to be updated, which still uses the vertical displacement technique for the baroclinic component of the assimilation. The update to the deeper pressure, and hence current, structure is derived by considering changes to the vertically averaged relative vorticity which are related to deep pressure updates geostrophically.

Two further assimilation experiments were then performed with ECMWF daily wind forcing from 1992 to 1993. With daily wind forcing the surface pressure signal is much larger, but much of this is barotropic variation. It is shown that the modified assimilation procedure, allowing bottom pressure changes, provides an effective filter which allows successful assimilation into this run by reducing the amount of baroclinic water displacement during assimilation. Without the modifications the vertical water displacements necessary to cancel out changes in surface pressure are unrealistically large. The new assimilation technique is tested over a 1-year period and shows similar success to the previous twin with water properties and currents converging to a degree both within and below the thermocline.

A parallel run without data assimilation but with correct daily winds used for forcing shows that the high-frequency component of the wind-forced variability is largely deterministic and converges after 100 days. This suggested a third assimilation experiment in which the high-frequency component of model variability is first allowed to converge after 100 days of forcing with the true winds. Thereafter assimilation is performed using the original CH procedure with no updates to deep pressure. This experiment also worked well, leading to convergence of all fields after 1 year of intermittent assimilation.

Modifications to the CH assimilation procedure may be unnecessary if the high-frequency barotropic component of ocean variability can be successfully modeled; however, this requires a good barotropic modeling capability and an accurately known wind field. In the Mediterranean this component of the surface signal is large and appears to be comparable in magnitude to the baroclinic variability. It is unclear to what extent this is true in the rest of the world oceans. Results from TOPEX and ERS 1 suggest that there are only a few areas in the world with significant large-scale high-frequency sea surface height variations [*Chao and Fu,* 1995]. In these areas, some filtering of the barotropic mode may be useful to prevent unrealistic baroclinic changes dur-

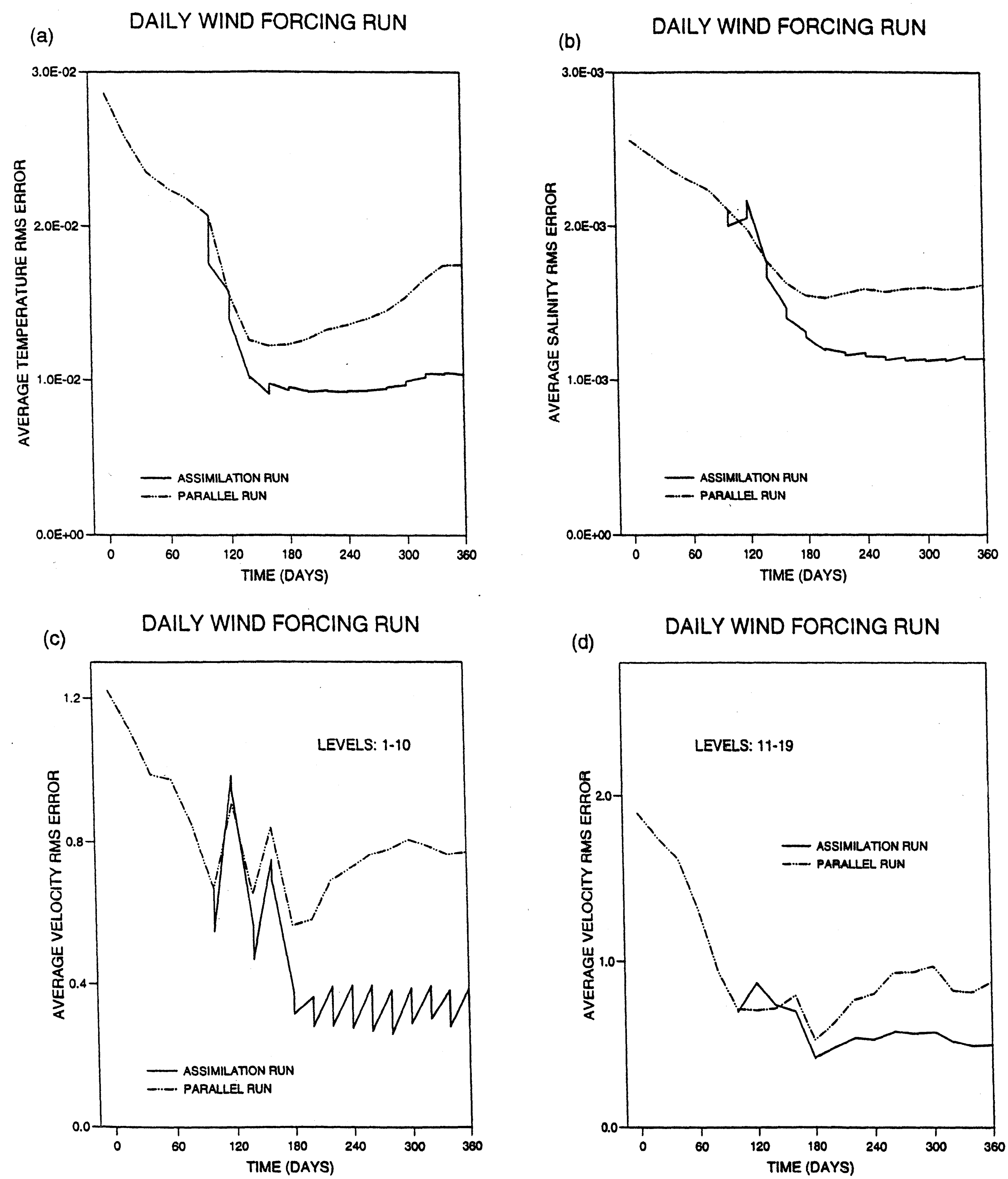

Figure 11. Solid lines show, for the daily wind-forced twin experiment 3, the errors in basin averaged (a) T, (b) S, (c) velocity above 300 m, and (d) velocity below 300 m. Assimilation only begins at day 100. Each plot also shows parallel run results (broken lines) from the same initial conditions with no data assimilation. Units are degrees Celcius (Figure 11a), practical salinity units (Figure 11b), and centimeters per second (Figures 11c and 11d).

ing assimilation if the CH method is used. This might be achieved using statistical correlations although the method outlined in section 4 appears to be robust and to provide a reasonably effective solution. Work is currently underway with a global ocean model in which these possibilities will be further examined.

The next step is to use these assimilation techniques with observed satellite data and a time-evolving GCM flow, in order to improve the synoptic circulation patterns and water property distributions over a particular time period and to test the possibility of making hindcast predictions.

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