Impact of Ocean Mean Dynamic Topography on Satellite Data Assimilation

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Q1

The response of an eddy-permitting ocean model to changes imposed by the use of different mean dynamic topographies (MDT) is analyzed in a multivariate assimilation context, allowing the evaluation of this impact, not only on the surface circulation, but also on the interior ocean representation. The assimilation scheme is a reduced-order sequential Kalman filter (SEEK). In a first set of experiments, high resolution sea surface temperature, along-track sea surface height and sea surface salinity from climatology are assimilated into a 1/3° resolution North and Tropical Atlantic version of the HYCOM model. In a second experiment, in situ profile data are assimilated in addition to the surface measurements.

The first set of experiments illustrates that important differences in the representation of the horizontal model circulation pattern are related to differences in the MDT used. The objective of assimilation is to improve the representation of the 3D ocean state. However, the imperfect representation of the mean dynamic topography appears to be an important limiting factor with regard to the degree of realism obtained in the simulated flow.

Vertical temperature and salinity profiles are key observations to drive a general circulation ocean model toward a more realistic state. The second set of experiments shows that assimilating them in addition to sea surface measurements is a far from trivial exercise. A specific difficulty is due to inconsistencies between the dynamic topography diagnosed from in situ observations and that diagnosed from sea surface height. These two fields obtained from different data sources do not contain exactly the same information. In order to overcome this difficulty, a strategy is proposed and validated.

Q2

Data assimilation for oceanographic applications has entered a new era since the advent of satellite missions that allow accurate measurements with high resolution and global coverage. In particular, radar altimetry provides quasi-global observations of the sea surface height (SSH) measured with respect to the reference ellipsoid. For this reason, SSH measurements have become one of the most important contributions to ocean prediction systems. However, the physical quantity that reflects the dynamic state of the ocean is the absolute

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Q4

ocean dynamic topography, corresponding to altimetric SSH minus the geoid. The geoid is
an equipotential gravity surface that is different from the reference ellipsoid. Unfortunately,
until now geoid uncertainties have been too large, at the scale of major ocean circulation   **40**
features, to allow using them for oceanographic purposes, and in particular for the deriva-
tion of the absolute dynamic topography. As a result, for most oceanographic applications
in general and for data assimilation in particular, the use of altimetric signals has been
limited to sea level anomalies (SLA), variations of the sea level about the mean sea level
(computed over a certain time span). The CHAllenging Minisatellite Payload (CHAMP)   **45**
mission (http://op.gfz-potsdam.de/champ/), launched in July 2000, was the first of a se-
ries of Geodetic missions that should lead to a precise estimate of the ocean geoid height
and, consequently, to improved knowledge of absolute dynamic topography. In particu-
lar, the European Space Agency Gravity field and steady-state Ocean Circulation Explorer
program (www.esa.int/esaLP/goce.html), scheduled for 2006, and the Gravity Recovery   **50**
and Climate Experiment (GRACE) (http://op.gfz-potsdam.de/grace/index_GRACE.html),
launched in 2002, are designed to provide the data set required to obtain an accurate geoid
(around 10 cm error) at a high spatial resolution (around 80 km). However, while substantial
progress should be made in this field following these missions, the delivery of a product
with sufficient precision will still take several years.   **55**

   Since the quantity that can be assimilated in ocean circulation models is the absolute
dynamic topography, we need to add information about the mean ocean state to the altimetric
SLA. This distance between the geoid and the mean level of the ocean is called the Mean
Dynamic Topography (MDT). (Note that the MDT is usually called mean sea surface
height by ocean modelers, since most ocean models use a spherical geoid so that both   **60**
quantities are identical). Different studies have provided different MDT products, with
different characteristics and accuracy. A number of approaches exist. An inverse model
of the ocean circulation constrained by observations can provide an estimate of the MDT
(Mercier 1986). A synthetic geoid approach consisting of combining satellite altimetry
and in situ data can also be considered (Mitchell et al. 1990). Another approach is based   **65**
on ocean model sea surface representation, with or without assimilation (Killworth et al.
2001). These different methods provide descriptions of the mean ocean circulation that are
not necessarily the same.

   Because of geostrophic balance, ocean surface currents are proportional to the gradient
of sea surface height. The change in the slope of the sea level elevation imposed by the   **70**
assimilation is expected to modify the simulated surface circulation. Surface currents as
well as absolute dynamic topography contain the integrated signature of ocean processes
within the whole water column. Therefore assimilating this quantity also implies making
a correction to the ocean thermohaline structure. The use of imprecise MDT products to
compute the absolute value of the ocean dynamic topography will impact the simulated   **75**
oceanic features. In addition to the sensivity of surface currents, it is important to determine
to which extent the choice of a particular MDT has an impact on the interior ocean properties
estimated by the assimilation system. Both ocean model and data assimilation schemes have
now reached a degree of accuracy where the realism of the imposed MDT can constitute
a major limitation in forcing ocean models towards reality on regional scales. This is why   **80**
it appears important to study the impact of state-of-the-art MDT products on assimilation
hindcasts. A better knowledge of the sensitivity of ocean simulations to MDT uncertainties
can also lead to significantly improved estimates of MDT.

   With extensive hydrographic programs like ARGO, we have now entered a phase
where the challenge is to combine optimally surface and subsurface information into a   **85**
single estimation process. Through these observations, it is expected to better control the
representation of subsurface fields/water masses. This also leads to a number of relevant

and challenging questions. Some recent studies have been conducted to explore the impact of the assimilation of sparse in situ data compared to satellite data, in terms of spatial and
90   temporal resolution (Li et al. 2003). A specific question that needs to be addressed here relates to whether the dynamic topography from in situ observations is consistent with that obtained by the addition of SLA and the imposed MDT in a multivariate assimilation system. Indeed, there is no reason why these two fields, which are different in nature, obtained from different data sources and associated with different kind of errors, should
95   contain completely compatible information.

The present study aims to demonstrate the sensitivity of a system assimilating altimetric data to different available MDT products. The detailed validation of the MDTs, which can be addressed independently of any assimilative system, is not the objective here. Such a validation or quantitative comparison would be dependent on the model and assimilation
100   system used. Both the case where only sea surface data are assimilated and that where vertical temperature and salinity profiles are added are adressed. The next section describes the numerical model, assimilation method, and data sets. Following this, a section concentrates on problem definition and the experimental framework and another contains the results of the sensitivity study. Then a section deals with the specific question of the assim-
105   ilation of temperature and salinity (TS) profiles. Finally, a summary and conclusions are presented.

Numerical Model, Assimilation Method and Data Sets

Numerical Model

The circulation model used is HYCOM, the HYbrid Coordinate Ocean Model (Bleck 2002;   **Q5**
110   Chassignet et al. 2003; Halliwell 2004) developed at RSMAS. The computational domain is   **Q6**
the North and Tropical Atlantic Ocean basin from 28°S to 70°N, including all the Mediterranean Sea. The horizontal grid resolution is given by $1/3° \times 1/3°\cos(\phi)$, where ϕ is the latitude. The Northern and Southern boundaries are closed and relaxation to climatology is applied within a 30 grid-point zone adjacent to the wall. The bathymetry is interpolated
115   from the ETOPO5 database onto the model grid. Subgrid-scale mixing is parameterized using a biharmonic dissipation operator. Vertical mixing is governed by the K-Profile Parameterization (KPP) mixing scheme (Large et al. 1994).   **Q7**

The hybrid vertical coordinate is isopycnal in the open stratified ocean, z-level in the mixed layer and unstratified areas, and smoothly reverts to a terrain-following coordinate
120   in shallow coastal regions (Figure 1). In this version, the vertical density structure is represented by 26 hybrid layers.[1] If the density of a given layer deviates from its isopycnic reference, the interfaces bounding this layer are adjusted, so that density is restored to its target value (adding denser or lighter water if the water is too light or too dense). If the layer interface approaches a prescribed minimum thickness, isopycnic conditions are not
125   entirely restored, and a cushion function is used to recalculate a vertical or sigma coordinate location (depending on bottom depth). The lower layers are isopycnic and exist only if the corresponding model target density of each layer exists; otherwise they collapse to zero thickness. Hydrographic and momentum equations have been modified to handle a generalized vertical coordinate. In each layer, both density and salinity are advected. The
130   theoretical basis for implementing the hybrid coordinate can be found in Bleck and Boudra (1981).

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[1]The prescribed reference σ_β values are: 19.5, 20.25, 21.0, 21.75, 22.5, [1]23.25, 24.0, 24.77, 25.28, 25.77, 26.18, 26.57, 26.8, 27.03, 27.22, 27.38, 27.52, 27.64, 27.74, 27.82, 27.88, 27.94, 28.0, 28.06, 28.09 and 28.12.

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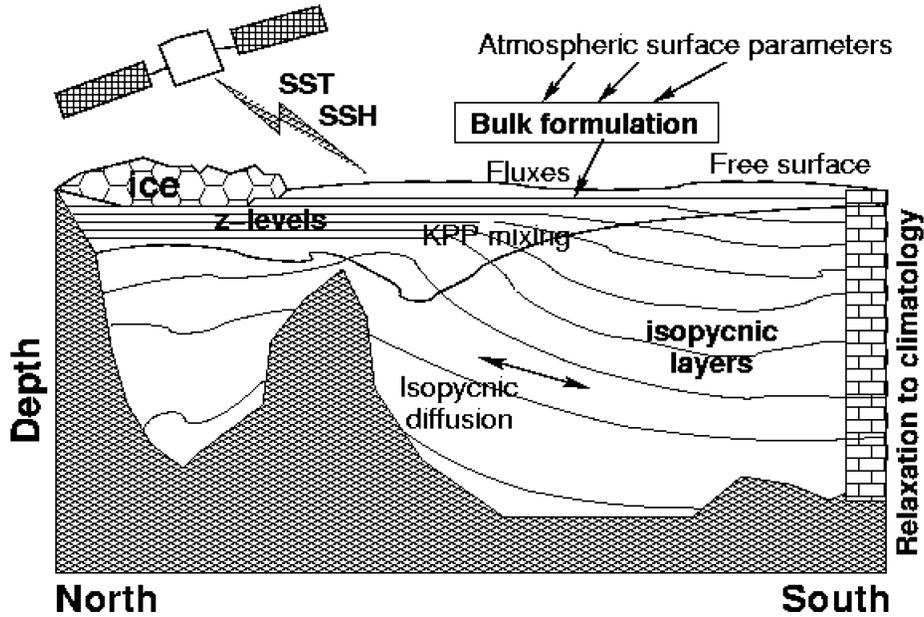
FIGURE 1 Sketch of the HYCOM model configuration.

The model is initialized using the Levitus climatology and is then spun up from rest during a 9-year time period, before running a 1985–1996 interannual experiment (without assimilation). Bulk formulation is used and the model is driven by wind stress, wind speed, air temperature and humidity, precipitation, and longwave and shortwave surface radiation. **135** During the spinup of the model, the forcing fields are derived from the Comprehensive Ocean-Atmosphere Data Set (COADS) monthly climatology. The sea surface salinity and sea surface temperature are relaxed towards the Levitus climatological fields with a 30-day timescale. During the 1985–1996 free model simulation and during the assimilation experiments, high frequency (every six hours) European Center for Medium-Range Weather **140** Forecasts (ECMWF) atmospheric data sets are used.

Assimilation Method

The assimilation method is derived from the SEEK filter, a reduced-order Kalman filter introduced by Pham et al. (1998). This sequential method has already been described and used in a number of studies (Brasseur et al. 1999; Penduff et al. 2002; Durand et al. 2003). In **145** the present implementation, we use the procedure described in detail by Brankart et al. (2003) and updated to take into account the hybrid nature of the vertical coordinate of HYCOM. The estimation vector is composed of the interface pressure, temperature and salinity for each of the 26 layers, in addition to the sea surface elevation. At every assimilation step, the misfit between observations and their model counterparts is evaluated and projected onto **150** the dominant error modes. The background error is initialized using the system variability **Q8** diagnosed from a free model run. In practice, we use the 10 dominant (EOFs) of a 3-year sample of model snapshots. A suboptimality is introduced in the scheme by keeping the background error covariance matrix unchanged from one assimilation cycle to the next. However, the use of spatially-dependant, three-dimensional, multivariate statistics derived **155** from a free model integration relates the present method to a simplified, reduced-order Kalman filter. The gain is local, meaning that the control of a specified water column will depend only on the observations within a specified influence bubble. In practice, the size

160 of each influence bubble is parameterized inside a box of 14×14 grid points. Observation errors are assumed to be independent; the corresponding prescribed values are given in section on data set which follows.

In order to reconcile statistical optimality with the physical constraints on the model state vector, an adjustment operator is introduced before model restart to restore the model constraints. A detailed justification of this operation can be found in Brankart et al. (2003),
165 where a similar approach was successfully applied to the MICOM model. The adjustment step consists of three main operations. First, we consider that the correlation between surface observations and the state variables at depth is not sufficiently reliable. In practice, only the state variables located inside the mixed layer are updated if TS profiles are not assimilated. Second, we restore the hydrostatic stability of the water column. The objective is to generate
170 a corrected analyzed state where layer density increases with depth and layer thicknesses are greater than their minimum value (HYCOM imposes a minimum thickness on each layer). In particular, if a layer has been reduced by too much during the statistical analysis, the excess reduction is transferred to the closest layer. Third, we apply a change along the Cooper and Haines mode (i.e., a vertical shift of the water column) to adjust the vertical structure of the
175 water column to the analyzed dynamic topography (Cooper and Haines 1996).   **Q9**

Data Sets

Sea Surface Topography

The altimetric observations consist of along-track sea surface topography, obtained as the sum of TOPEX/Poseidon or ERS altimeter SLA, and the mean dynamic topography
180 (see next section). The assimilation window is three days. Each analysis is computed using all the available data gathered within a 3-day interval (1.5 days before and after the analysis time). The accuracy of the SLA can be estimated at around 3 cm, except in the coastal zones, where tidal corrections are less accurate. We therefore decided to mask SLA data where the bottom depth is less than 150 m. In order to eliminate spurious effects due to
185 the transition between zones where the model state is corrected by the assimilation of SLA and zones where it is not (corresponding to significant portions of the American and European continental shelves), we replace the sea surface topography by the mean dynamic topography part in such coastal areas. A bulk error of 3 cm RMS was prescribed on the total sea surface topography data in the assimilation system.
190 For the present study, three different MDTs for the North and Tropical Ocean, computed with three different methods, were collected and used to perform three different assimilation experiments. The first, referred to as MDT1, is the MDT from Rio and Hernandez (2002),   **Q10** the second, referred to as MDT2, is from Le Grand et al. (1998), and the third, referred to   **Q11** as MDT3, is derived from a $1/12°$ Atlantic MICOM experiment (Chassignet and Garraffo
195 2001).

MDT1 is a $1° \times 1°$ product, estimated from a multivariate statistical analysis using hydrographic data, surface drifter velocities and altimetry over the 1993–1999 period. It is based on the SMO CLS01 mean sea surface (Hernandez et al. 2001) and the EIGEN2 geoid computed from CHAMP data. This MDT is produced to match the altimetric average re-
200 moval period (1993–1999); the associated drawback is that in situ and drifter measurements available during this period provide insufficient sampling in some areas.

MDT2 is a $1° \times 1°$ product obtained by merging in situ measurements and altimetric observations using a nonlinear inverse model (Mercier 1986). Hydrographic data come from a compilation of 70 years and altimetric observations cover the 1993–1996 period.
205 This approach is based on dynamical considerations but still strongly limited by resolution and accuracy of available data.
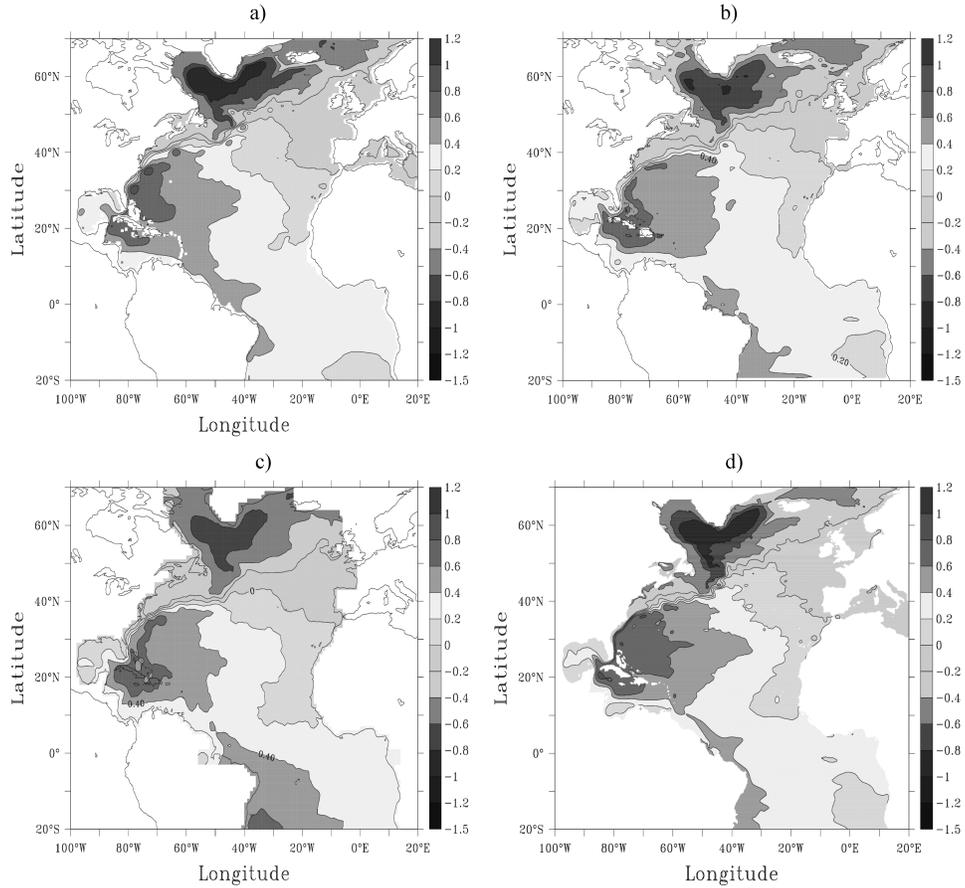
FIGURE 2 Mean Dynamic Topography (in meters): (a) MDT0, (b) MDT1, (c) MDT2, and (d) MDT3.

MDT3 is a 2-year average of a sea surface elevation time series from a $1/12°$ Atlantic MICOM free run experiment performed with ECMWF forcing. This product has a high spatial resolution, but the associated mean field includes systematic model errors (mainly due to unresolved physics, numerical errors, errors in boundary conditions, or forcings).   **210**

Figure 2 shows these three different MDT fields, which roughly correspond to three available types of MDT estimates. The 4-year average MDT computed from a $1/3°$ Atlantic HYCOM free run experiment over the period 1993–1996 (named MDT0) is also shown for comparison. With regard to large-scale features, the four MDTs are fairly similar in terms of structure and amplitude, showing the classical picture of the mean ocean circulation in the   **215** North and Tropical Atlantic Oceans. The signature of the main currents is present. All MDTs represent a realistic Gulf Stream separation at Cape Hatteras, except MDT0 that is affected by a significant model bias in the Gulf Stream region. The Gulf Stream (GS) overshoots in HYCOM $1/3°$; this is a classical problem of ocean numerical models at eddy-permitting resolutions. The three MSSH fields differ mainly at the smaller scales, and also in their   **220** representation of several regional circulation features. In broad terms, MDT2 is smoother than the other two. This is due to the longer period of observations used to compute this field. In MDT3, the Florida Current, the Gulf Stream and North Atlantic Current (NAC) front are sharper than in the other MDTs. This is due to a better representation of the

225 strength of ocean currents in a 1/12 degree grid mean than in a lower resolution mean. In MDT1, MDT2, and MDT3, the NAC front turns north around 45°W, and then toward the northeast around 40°W, forming the North Atlantic Drift Current along the eastern flank of the Reykjanes Ridge. Many small-scale discrepancies between the different MDTs can be observed all along the resulting fronts. Another quantitative difference can be found **230** in the subpolar gyre, with fronts associated with the Labrador Current and the East/West Greenland Current system that are significantly weaker in MDT2, illustrating a poorly defined subpolar gyre circulation.

An illustration of the differences of the Gulf Stream front representation by the different MDTs is given by Figure 3, which represents cross-sections of the three MDT products along
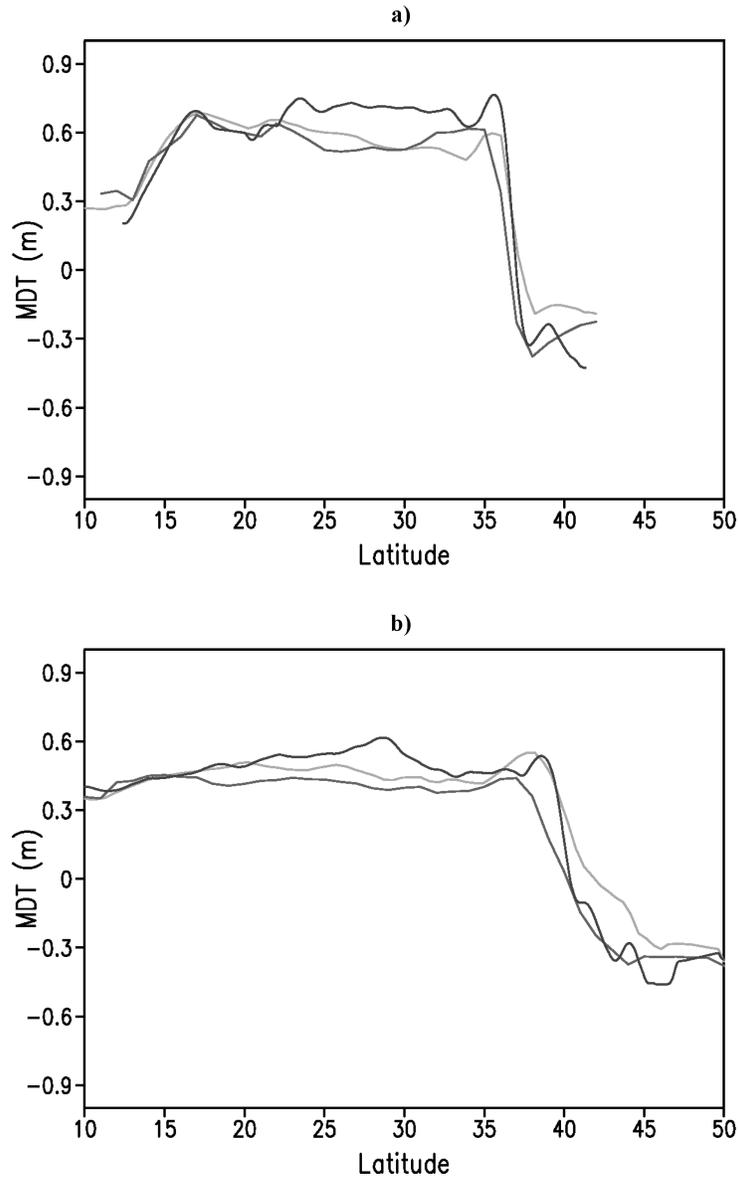
FIGURE 3 Section of the different mean dynamic topographies (in meters) (a) along 72°W and (b) along 55°W. MDT1 is in green, MDT2 is in red, and MDT3 is in blue.

72°W (Figure 3a) and along 55°W (Figure 3b). Both sections cross the Gulf Stream front, **235**
represented by a sharp MDT gradient. The difference across the Gulf Stream front is 10 cm
larger in MDT3 than in MDT1 and MDT2. MDT1 has a weak front amplitude just off Cape
Hatteras (20 cm lower than MDT2). MDT3 and, to a lesser extent MDT1, have secondary
fronts south and north, associated to the inertial Gulf Stream recirculation, which are not
present in MDT2. The Gulf Stream mean path is the same in MDT2 and MDT3, but is **240**
different in MDT1, with a too Northern position (Figure 3b).

Sea Surface Temperature and Salinity

The Sea Surface Temperature (SST) observations assimilated in HYCOM consist of
high-resolution composite AVHRR data (10 km resolution at the equator). The available
SST data are gathered within the 3-day assimilation time window. An observation error **245**
of 0.4°C is prescribed. The SST data coverage can be strongly limited by the presence of
clouds, particularly at high latitudes and during the winter season.

For multivariate data assimilation, inaccuracies in the specification of forecast error
covariances can lead to inappropriate corrections of unobserved fields like SSS. In order
to avoid such problems, we assimilated Levitus 98 monthly climatological Sea Surface **250**
Salinity (SSS) in addition to SST and SSH. Unlike SST or SLA, SSS measurements are
sparse both in time and space. The European Space Agency SMOS mission that will be
launched in the near future will provide SSS data every 10 days at 200 km resolution, but at
the present time there is no high resolution SSS data set available. Because climatological
fields are very smooth, we introduce a smoothing operator into the observation operator **255**
to compute the model equivalent to the data. In this way, only large-scale features in the
model can be influenced by SSS data assimilation. For this reason, the error associated to
the large-scale averaged SSS is set to 0.05 psu.

In Situ Profiles

We could have chosen WOCE temperature and salinity profiles as data sources for **260**
our in situ data assimilation experiment. However, their spatial and temporal distribution
remains sparse, with many sites having only temperature profiles and most of them having a
poor vertical extension compared to local ocean depth. These features would have induced
specific problems.

In the multivariate data assimilation experiment, we first want to illustrate important **265**
aspects of the complementarity between surface and subsurface data. Thus, we have chosen
to extract the assimilated profiles from the Levitus 98 monthly climatology. In order to
enable assimilation into HYCOM, they are transformed into hybrid coordinates, so that
observations of temperature, salinity, and interface pressure are available for each layer
and are effectively assimilated. Note that correlations between interface pressure errors **270**
and layer TS errors have been dropped in the background error covariance. Observation
errors have been set at 0.2°, 0.1 psu, and 20 m, respectively, for temperature, salinity, and
pressure measurements. This information is used with an adequate observation operator
(same treatment as for climatological SSS), so that it only influences the large-scale features
of the 3-D TS structure. **275**

Problem Definition and Experimental Framework

Problem Definition

Since information about the geoid is not sufficiently precise, the mean (MSSH) of the
altimetric signal SSH has to be removed, so that only the altimetric residuals can be accessed.

280   To obtain the quantity assimilated into the model, that is, the absolute ocean dynamic topography (ADT), a mean dynamic topography (MDT) must be added to the SLA. Ideally, this MDT should correspond exactly to the average state of the ocean over the time period corresponding to the altimeter measurements. One simple solution would be to use the model mean ocean state over this time period. Unfortunately, this variable is affected by significant

285   systematic errors, due to various model deficiencies. For this reason, it is necessary to use a more realistic mean ocean state. Our assimilation experiments will be performed by adding one of the MDT described in the section on data sets to the SLA.

Figure 4 illustrates the definition of the height measurements used. In summary:

$$ADT = SSH - MSSH + MDT.$$

MDT is a synthetic product and can then be altered by (1) errors associated with the data

290   and the method used to compute this field, (2) bias related to nonuniform time and space coverage of the observation data sets used, and (3) bias introduced by the fact that the average state of the ocean is calculated over a time period which does not correspond to the altimeter measurements. These differences can lead to spurious effects in the assimilation experiments. Thus, let us define:

$$MDT(x, y) = S0 + S1(x, y),$$

295   where $\int \int_D S1(x, y)dxdy = 0$, D being the model ocean domain.

$S1$ is due to permanent ocean circulation patterns that vary geographically and determines a specific shape of the MDT. $S0$ is the horizontal mean difference between the geoid and the *MDT*. $S0$ is then purely conventional and defines a reference surface parallel to the geoid. In a system covering the global ocean domain, $S0$ is conventionally equal to 0.

300   Problems arise because (1) the conventional reference may not be the same for the model and the data (for instance, our model is regional and was initialized with $S0$ equal to 0 over the North and tropical Atlantic), and because (2) during the spinup, the model may have drifted from the initial climatological state. For these reasons, a significant horizontal mean difference exists between the MDT and the corresponding current model state. In the

305   absence of vertical information (i.e., vertical profiles acting as an additional constraint in the assimilation system) that could track these misfits back to identifiable causes, no doubt we will face difficulties in interpreting this important part of the dynamic topography. This misfit can be of the same order of magnitude as the part due to sea-level variations. To account
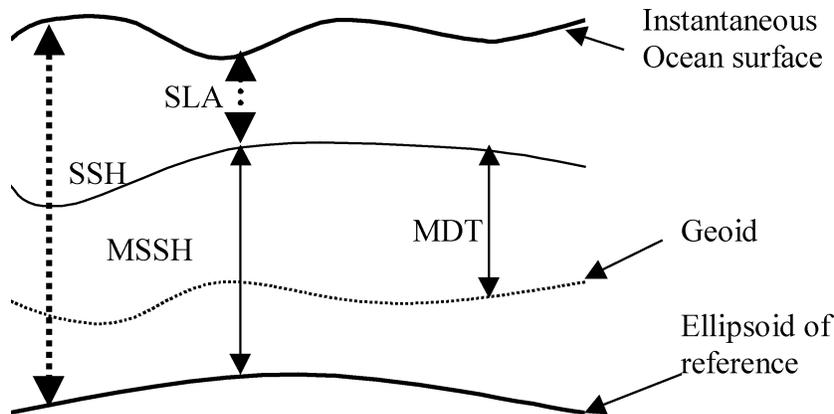
FIGURE 4 Definition of height measurements.

for this problem, an averaged misfit value $S0$ is introduced, characteristic of the current model state. The $S0$ component is removed from the corresponding MDT field, meaning **310** that we do notcorrect the error corresponding to the model drift in the initial model state. Otherwise, layer interfaces could be shifted along the Cooper and Haines mode, resulting in unrealistic vertical displacement.

Description of the Experiments

In a first set of experiments, four simulations are performed between October 1992 and **315** December 1993. The first (called EXPT0) is a free-run simulation, without assimilation. It is defined as the control run. The other three differ from the first in terms of the assimilation of sea surface temperature, sea surface salinity, and sea surface height observations. The same initial condition and the same forcing fields were used for these experiments. They differ only in the MDT used to compute the absolute dynamic topography assimilated into **320** the system: MDT1 for EXPT1, MDT2 for EXPT2, and MDT3 for EXPT3

In a second stage, a fifth simulation, called EXPT4, is performed over the same period (October 1992–December 1993). In EXPT4, TS profiles are assimilated in addition to the surface data and MDT2 is used. The methodology is described in the section on the assimilation of TS profiles. **325**

In this study, we present different diagnostics of these simulations. Focusing on the North Atlantic region, these diagnostics are all calculated from the model forecasts, before statistical correction, and over the year 1993 (from January to December).

Sensitivity Tests: Results

The system response is analyzed for the first set of simulations. The ability of the model, **330** with assimilation, to reproduce the essential elements of the North Atlantic circulation is discussed. The sensitivity of the solution to the different MDTs is analyzed.

For this study, we assume that the assimilation results have already been validated. The diagnostics presented here are (1) the surface currents which are directly related to the changes in MDT through geostrophy and are crucial for climate studies and many **335** operational applications; and (2) the mean barotropic streamfunction as an expression of the vertically integrated circulation pattern. A temperature section along 55°W is also examined to consider the sensitivity of the solution in terms of thermohaline characteristics.

Surface Currents

Figure 5 shows the difference between the simulations in terms of the mean currents for **340** 1993 at 50 m depths. The classical surface currents of the North Atlantic are clearly visible in all solutions but, as expected, there are also important regional differences.

The mean flow of the Florida Current is particularly strong in the control run and in EXPT3, where it exceeds 90 cm/s, while it is about 55 cm/s in EXPT1 and 65 cm/s in EXTP2. This result is coherent with the local gradients of the different MDTs, which are **345**

Q12 more realistic in EXPT0 and EXPT3 (Fratantoni 2001). Although the corresponding current velocities are greater in EXPT3, the GS extension and the NAC pathway are fairly similar in all assimilation solutions. The NAC turns north around 45°W, and then northeast around 40°W. While it is around 30 cm/s in EXPT1 and exceeds 40 cm/s in EXPT3 and the control run, it is significantly weaker in EXPT2, with a mean speed of around 20 cm/s. The latter **350**

Q13 value is too low (Frantatoni 2001). In all assimilation experiments, the Azores Current is shifted to the south and its flow slightly increased. From less than 10 cm/s in the control run, its speed reaches around 12 cm/s, with a maximum of 15 cm/s, in EXPT1, EXPT2, and EXPT3 near 25°W.
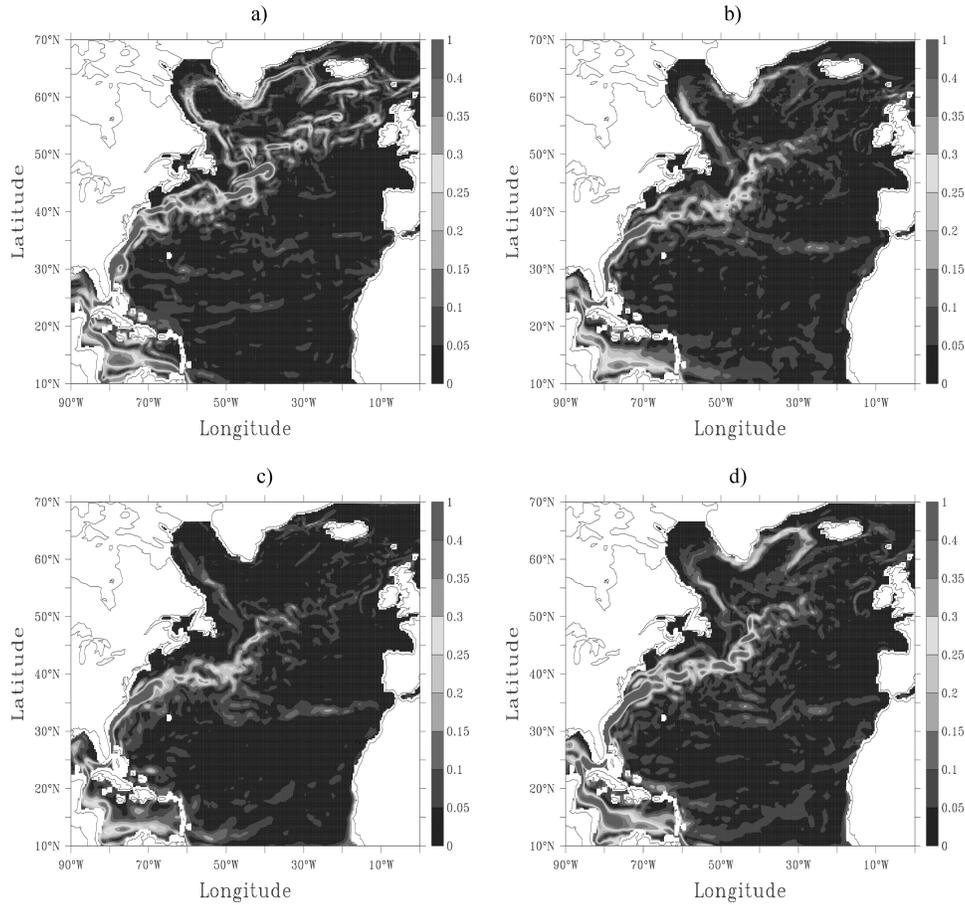
FIGURE 5 Mean surface current velocity for (a) control run EXPT0, (b) EXPT1, (c) EXPT2, and (d) EXPT3 assimilation experiment.

355   North of 50°N, the representation of the main surface current system is more variable from one experiment to another. In both the control run and EXPT3, west of 30°W, one branch of the NAC turns northward to join the Irminger basin, before crossing the Mid-Atlantic Ridge. This branch is much weaker in EXPT1 and EXPT2. The North Atlantic Drift Current is particularly strong in the control run, with the mean surface current exceeding 20

360   cm/s in some areas, which appears unrealistic. Observations highlight various branches of weak currents flowing to the northeast, rather than an intense and continuous flow (Reverdin et al. 2003). By contrast, in EXPT2, the surface currents are weak in the whole northeast region.

The East-West Greenland/Labrador Current system appears fairly different from one

365   solution to another. The corresponding mean currents are reduced in all assimilation experiments, and particularly in EXPT2. With a mean velocity of about 35 cm/s in the control run, the Labrador current reaches 25 cm/s in EXPT1 and around 15 cm/s in EXPT2 and EXPT3. By comparison to Reynaud et al. (1995), the latter value is too weak With a mean **Q14** speed of about 35 cm/s in the control run, the West Greenland Current flows at around

370   5 cm/s in EXPT2, around 15 cm/s in EXPT1, and 20 cm/s in EXPT3. In EXPT2, all the West Greenland Current turns west near 61°N whereas it turns west near 63°N in the control

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run; in EXPT1, and EXPT3, different branches exist between 60°N and 64°N, which is in better agreement with observations (Cuny et al. 2002).

For each of the experiments, the regional differences in the simulated surface currents can be largely attributed to differences in the MDTs, moderated by the effect of the as- **375** similation. In EXPT2, the currents are generally weaker. In contrast, EXPT3 exhibits the highest current values. This can be related to the MDT resolution; whereas, MDT2 is the smoothest MDT solution and MDT3, which was deduced from a 1/12° free-run simulation, has the highest resolution.

Barotropic Streamfunction   **380**

Figure 6 shows the 1993 mean barotropic streamfunction for the four simulations. Note first that changes introduced by assimilation are significant. Barotropic transport is increased by assimilation, and the increase is much larger in EXPT1 and EXPT3 than in EXPT2. At the global scale, all simulations show well-developed gyres of similar structure, though they all differ in amplitude. At the regional scale, however, differences in structure may also be **385** observed.

The Northwestern subtropical gyre limit has drifted southward with the assimilation, thus correcting the GS pathway. Both the transport associated with the Florida Current/Gulf
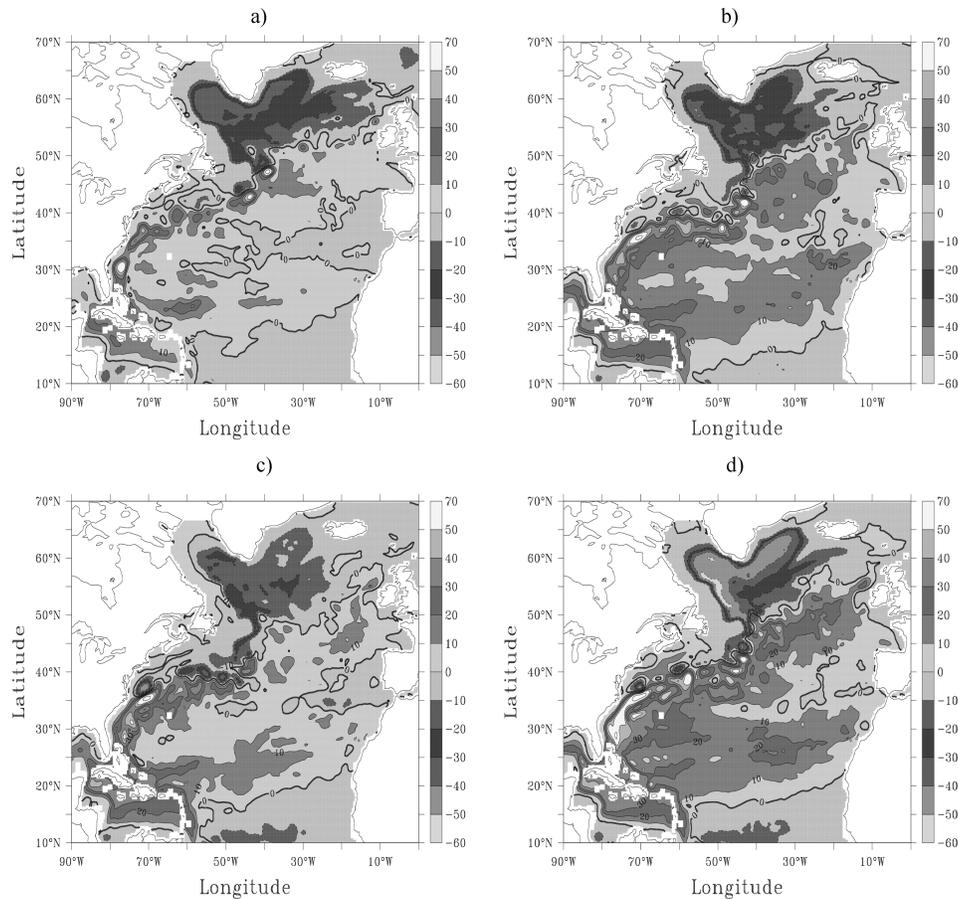
FIGURE 6 Mean barotropic streamfunction for (a) control run EXPT0, (b) EXPT1, (c) EXPT2, and (d) EXPT3 assimilation experiment.

Stream system and the associated recirculation are more pronounced in EXPT1 and EXPT3
than in the other solution. Numerous small cells are found along the GS and NAC paths,
which have been shifted by the assimilation along the line of maximum gradient of mean
dynamic topography. These small-scale structures are very different from one solution to
the next but are much larger and stronger in EXPT1 and EXPT3. This feature can be related
to the presence of small-scale structures along the subtropical front in MDT1 and MDT3
that are almost absent in MDT2.

The amplitude of the subpolar gyre also differs significantly. With a magnitude of approximately 30 Sv in the control run, it maintains approximately the same magnitude in
EXPT1, increases to 50 Sv in EXPT3, and decreases to a value of 20 Sv in EXPT2. Significant differences are also observed in its structure. The location of maximum transport varies.
Both subpolar gyre structure and maximum transport location observed in the solution are
clearly correlated with the subpolar gyre structure and minimum mean sea level location of
the different MDTs.

Vertical Temperature Section at 55°W

As already observed in the previous section, the assimilation of surface data modifies the
representation of not only the sea surface fields, but also the entire ocean state, through a
vertical extrapolation of sea surface information. In this section, we analyze the sensitivity
of the interior thermohaline properties to the use of different MDTs in the assimilation
system.

Figure 7a represents the thermal structure along the 55°W section from Levitus 98 climatology, while Figures 7b, 7c, 7d, and 7e show the differences along this section between
the mean thermal structure of, respectively, the control run, EXPT1, EXPT2, EXPT3, and
the climatological thermal structure of Figure 7a. Assimilation strongly modifies the subsurface temperature structure and significant differences are observed from one experiment
to another.

In the Labrador Sea, temperatures in the control run are more than 2°C too warm over
the Northern and Southern limits of the Labrador Sea, where the West Greenland/Labrador
Current system flows. In the assimilation runs, these anomalies have been reduced and even
replaced by a cold anomaly in the Northern part of around 0.5°C in EXPT2. In the interior
basin, the temperature is colder from the surface to the bottom in all assimilation runs, but
the amplitude of cooling varies from one experiment to another.

The most marked anomaly with respect to the climatology is observed in the control experiment north of 40°N, just south of Newfoundland. Subsurface temperatures are as much
as 5°C too warm in the upper 500 meters, associated with the large-scale "misplacement"
of warm water from the Gulf Stream extension (located around 41°N) and the associated
absence of colder Labrador Current water near the continental shelf. This anomaly has been
considerably reduced in all assimilation experiments, especially in EXPT2 and EXPT3.
The differences observed in the amplitude of this anomaly reduction are due to the presence
of a small northern branch of the GS present near 45°N (its signature is also observed on
Figure 5), preventing colder water from the Labrador Sea turning southwest around Newfoundland. This current becomes weaker in EXPT2 than in EXPT1 and EXPT3, which
explains the differences in the thermal structure observed in this area in the different assimilation simulations. Parallel to the observed decrease in the warm anomaly, a negative
feature of 2°C amplitude appears north of 40°N, between depths of 500 m and 1000 m
in EXPT2. This negative feature can be associated with the change in Labrador Sea water
mass characteristics noted in the previous paragraph.

In the midlatitudes, between 20°N and 40°N, the control run is too warm between approximately 100 m and 600 m compared with the climatology. This feature is significantly

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amplified in both location and amplitude by the assimilation in EXPT3, an anomaly that
corresponds to the presence of 18°C Mode Water formed in the Sargasso Sea. In the clima-
tology, this water mass has a temperature of 17°C–18°C (Figure 7a) and extends between **440**
depths of 200 m and 450 m. In all model runs, the water mass is too warm with a temperature
of around 19°C. The misrepresentation of water mass structure results in a stronger positive
anomaly in EXPT3, where it is too deep.
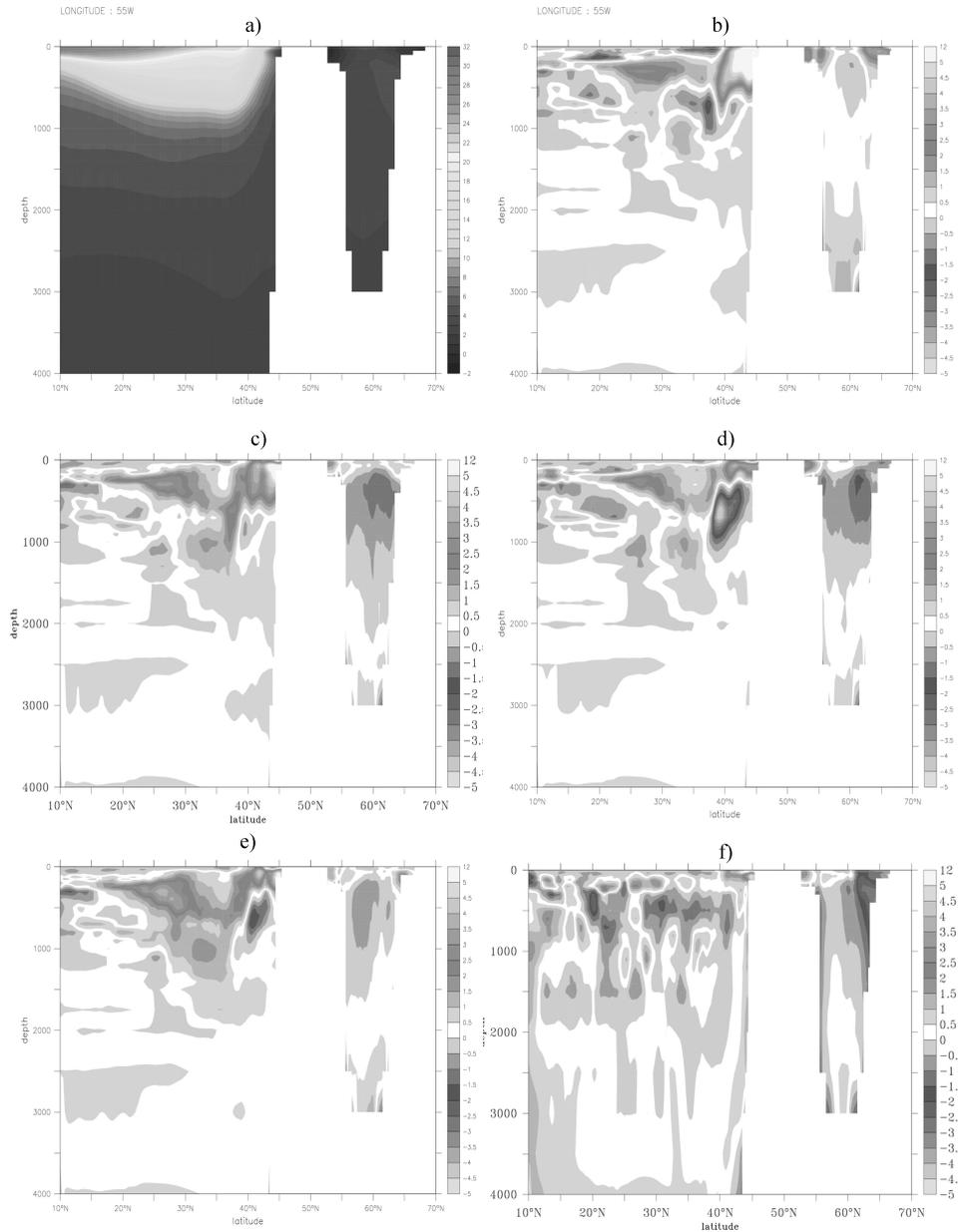
FIGURE 7 (a) Temperature section along 55°W from Levitus climatology. Anomaly be-
tween the mean temperature from (b) control run, (c) EXPT1, (d) EXPT2, (e) EXPT3, (f)
EXPT4, and the climatological temperature field along 55°W.

The results of this section illustrate the impact of the choice of the MDT product on different ocean estimates. Although the impact of assimilation is generally positive, the different experiments do not perform equally well everywhere. The surface currents and the deep properties are very sensitive to the change in mean dynamic topography, in particular where gradient differences are observed. In addition, the MDT, through different mechanisms, also has an impact on a wide range of model oceanic processes, such as volume transport and thermohaline circulation which are shown to be very sensitive quantities.

Comparison with Independent TS Profiles

Another way to compare results is to validate each experiment using in situ data. For this purpose, the model thermal and salinity fields were compared to (a) the climatology and (b) an ensemble of XBT profiles collected in the North Atlantic. Figure 8 shows the distribution of available XBT in 1993.

RMS misfits between the XBTs or climatology and the equivalent model state were computed for both the Gulf Stream region and the Labrador basin and are shown in Figure 9. For each assimilation experiment, the continuous-line curve represents the forecast and the dashed-line curve the analyses (after dynamical adjustment). There is little difference between analyses and forecasts, indicating a stable assimilation system. Here, the discussion is limited to the results of the experiments where no TS profiles are assimilated.

In the Gulf Stream region, where the results are particularly sensitive, EXPT1 and EXPT3 are closer than the free run to both climatology and XBTs down to a depth of at least 400 m. However, EXPT2 shows better results than EXPT1 and EXPT3 down to a depth of 700 m, with misfits remaining better than the free run. Note that the vertical structure of the temperature misfits exhibits a local maximum around 100 m depth, which is related to the difficulty in correctly simulating the mixed layer depth. This maximum is greatly reduced by the assimilation, particularly in EXPT2. Note also that the error in salinity is also reduced, due to the assimilation of surface salinity observations.
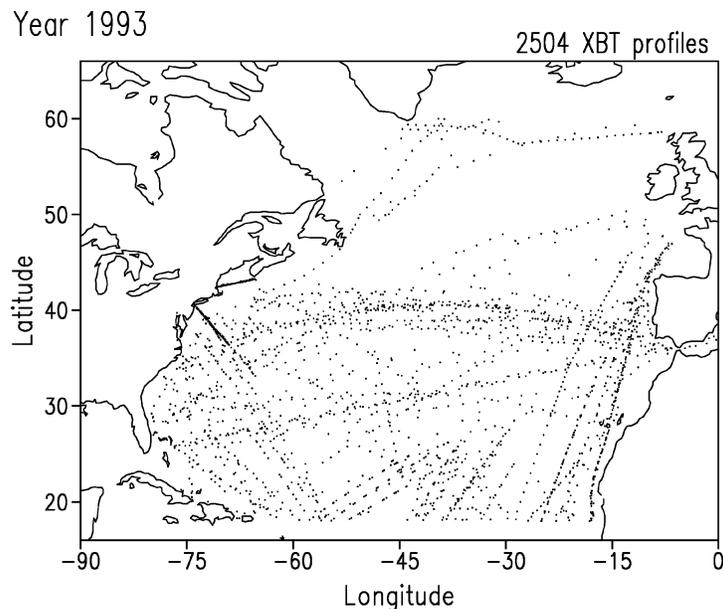
FIGURE 8 Horizontal distributions of XBT profiles available during 1993 (SISMER).

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a)   b)
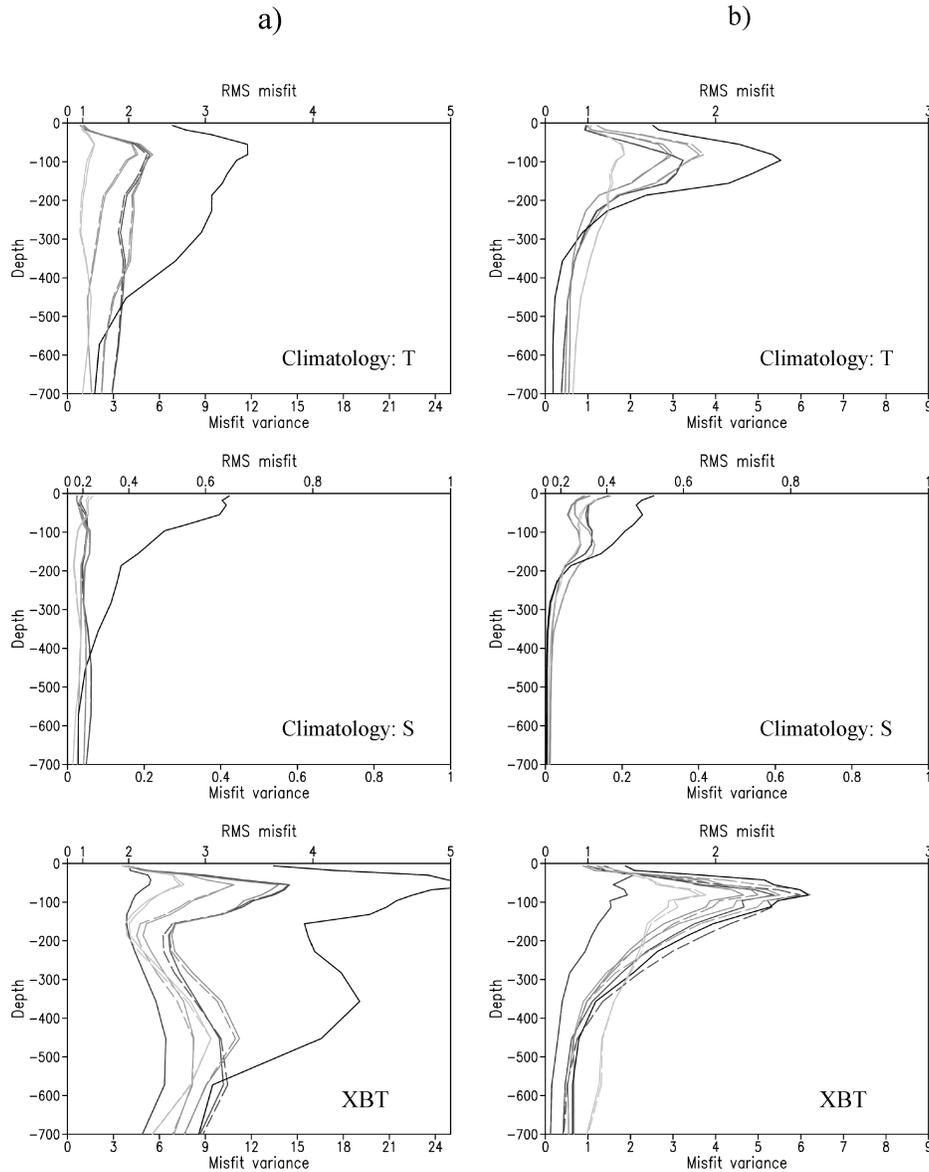
FIGURE 9 Temperature (°C) and salinity (PSU) RMS misfit with respect to TS climatology
and to XBT profiles in (a) the Gulf Stream region and (b) the Northeast region, down to
700 m and averaged for 1993. The figure shows plots of the free run (black curve), the
EXPT1 (turquoise), EXPT2 (green), EXPT3 (red), and EXPT4 (yellow) solutions. For each
assimilation experiment, the 3-day forecast is shown by the continuous-line curve and the
analyses by the dashed-line curves.

In the Labrador region, the results are different. Here again, the impact of assimilation   470
is positive, but the differences between the temperature misfits of EXPT1, EXPT2, and
EXPT3 are less marked, due to a much weaker vertical temperature gradient. In this region,
EXPT1 shows slightly better results.

Specific Case of the Assimilation of TS Profiles

475   In a multivariate case, the Kalman gain generates corrections based on specified cross-field error covariances. Errors in surface fields are related to errors in subsurface. Consequently, when the system is constrained only by surface observations, inaccuracy in cross-field covariances will lead to inaccurate corrections in water mass properties at depth. Given our limited knowledge of model errors and the associated correlations, this is a significant

480   limitation on the quality of ocean predictions. The assimilation of new observations, such as TS fields, should moderate this problem.

However, this also means the introduction of new information types in the system, that is, of heterogeneous variables. The question we wish to address here is the problem of consistency between the sea level elevation corrections provided, on one hand, by the

485   MDT information and, on the other, by the TS profiles. Indeed, there is no reason why these different data sources should contain exactly the same information about the $S0$ and $S1$ components defined previously in the section on problem definition. Note that this difficulty would not arise (1) if the model had not drifted from its initial climatological state, (2) if both fields were not associated with large errors, and (3) if the conventional sea

490   level reference was the same for the model and the data.

In order to overcome potential difficulties, we adopted the following strategy. Based on the definition in the problem definition section, a new $S0$ horizontal mean level was determined using in situ profile information, whereas the $S1$ component was prescribed using the MDT. This approach amounts to an estimation of the sea level topography in two

495   steps. First, we assume no possible connection between errors in sea level elevation and errors in the subsurface fields. In practice, it means that we have dropped the cross-field correlation to zero in the forecast error covariance. Thus, the sea level elevation correction will be constrained only by sea surface observations. Second, modifications are introduced into the adjustment operator in order to take advantage of the TS data content for the estimate

500   of corrected sea level elevation. The density field is deduced from the analyzed vertical fields (after restoration of the hydrostatic assumption). The corresponding sea surface topography is calculated and compared to the analyzed sea surface topography prescribed by the filter. The horizontal mean difference between these two fields is then computed and added to the analyzed sea surface topography. In this way, we also expect to correct progressively

505   the bias introduced in the initial state by model drift. This approach was used to perform EXPT4.

In order to test this method, we show in Figure 10 the temporal evolution of mean sea surface heights averaged over the whole model domain corresponding to EXPT0, EXPT2, and EXPT4. The free run simulation drifts slowly away from its initial state with time

510   because the model thermal content is increasing during the numerical experiment Using surface observations only, the SEEK filter eliminates this model drift by controlling the mean sea level trajectory from the initial assimilation date (EXPT2 curve). The behavior of the EXPT4 curve is different. After a first stage in which mean sea surface height decreases, it stabilizes and starts reproducing the same seasonal variability as the EXPT2 curve: a new

515   S0 value based on the TS information has been determined. It means that, with the proposed method, we are able to use climatological subsurface observations correctly to reduce the error in the initial mean state and then to improve significantly the model trajectory.

An analysis is also conducted on the representation of water mass properties in EXPT4. A comparison of Figure 7f with the other figures (particularly Figure 7d, where the same

520   MDT was used) shows that in EXPT4 temperature has generally been brought into better agreement with the climatology in the ocean interior. In particular, the 18°C mode water properties have been successfully modified. In mid-latitudes, the estimated temperature
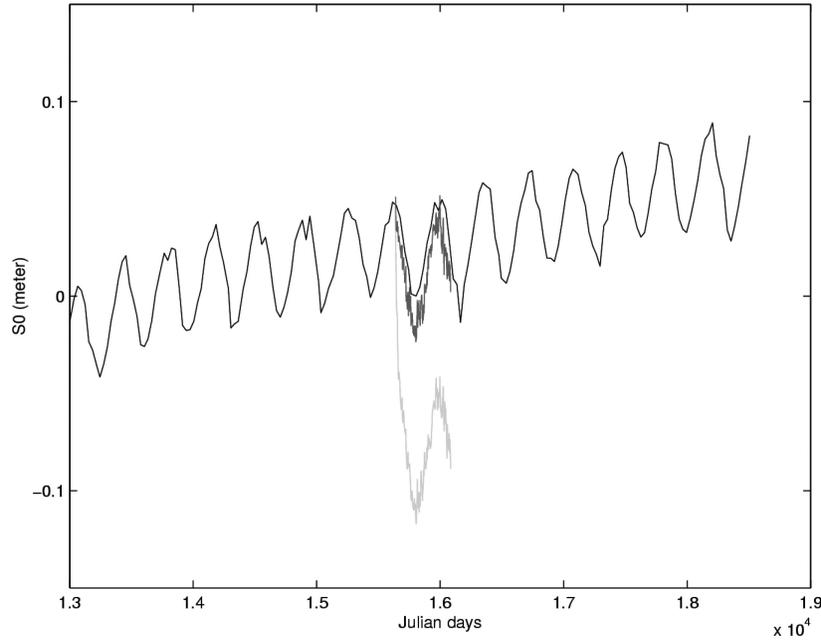
FIGURE 10 Time evolution of the sea surface elevation averaged over the model domain for EXPT0 (black), EXPT2 (red), and EXPT4 (green).

exhibits some small-scale colder features of less than $1°C$ around 500 m depth. This might be due to a limitation of the system in producing accurate water mass property corrections using a limited number of error modes (10), when the number of degrees of freedom for **525** the problem is much greater. Finally, Figure 9 shows the resulting RMS misfit between the XBTs or climatology and the equivalent model state, also illustrating the positive impact of the assimilation of TS profiles.

These results illustrate that the corrections to the depth of the model isopycnic layers provided by the assimilation of TS profiles are accurately estimated. A specific TS profile **530** assimilation experiment was performed without using this protocol (not shown), and the results suffered from significant errors in the estimation of depths for water mass correction.

Summary and Conclusions

The main objective of this work was to assess the sensitivity of simulated ocean state estimates to different MDT products in a multivariate assimilation context. Ongoing and **535** future satellite missions should provide accurate measurements of the gravity field over the ocean, but it will still take several years before precise estimates of the absolute dynamic topography are produced. In the near future, to compute this quantity we will still have to rely on estimates obtained using the different existing approaches.

In this study, we first presented three different experiments, assimilating SSH, SST, and **540** SSS observations into a hybrid vertical coordinate model of the North and Tropical Atlantic Ocean. In each experiment, a different MDT product was used to assimilate satellite sea level anomalies into the model. Important differences between the simulations were observed in terms of ocean circulation, transport, and thermohaline fields, which can be related to the differences between the three MDT estimates. The diagnostics demonstrate that the **545** assimilation system is sensitive to the choice of the MDT, not only in terms of surface

currents, but also in terms of deep properties. It remains difficult to determine which MDT estimate is the most appropriate for assimilation. With regard to the different diagnostics presented, and particularly to the systematic comparison with in situ TS profiles, the results vary from one oceanic region to another. In conclusion, errors in MDT estimates are shown to be among the most important factors affecting simulation quality.

An additional experiment was performed to evaluate the joint assimilation of dynamic topography and TS profiles. According to the protocol proposed, the MDT information was used to control the horizontal mean sea level pattern, while TS profiles were used to correct progressively the horizontal averaged sea level elevation. Assimilation of TS profiles proved efficient in correcting the significant error in the initial model state (largely due to model drift) and in making the vertical structure of the ocean more realistic. Finally, the study also underlines the necessity of ensuring consistency between MDT and in situ profiles.

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