Impact of uncertainties in atmospheric boundary conditions on ocean model solutions

Ayan H. Chaudhuri^{1*}, Rui M. Ponte¹, Gael Forget²

1. Atmospheric and Environmental Research (AER), 131 Hartwell Ave, Lexington, MA 02421.
2. Massachusetts Institute of Technology (MIT), 77 Massachusetts Ave, Cambridge, MA 02139.

*Corresponding Author: achaudhu@aer.com (P) 7817612382 (F) 7817612299

ABSTRACT

We quantify differences in ocean model simulations derived solely from atmospheric uncertainties and investigate how they relate to overall model errors as inferred from comparisons with data. For this purpose, we use a global configuration of the MITgcm to simulate 4 ocean solutions for 2000-2009 using 4 reanalysis products (JRA-25, MERRA, CFSR and ERA-Interim) as atmospheric forcing. The simulations are compared against observations and against each other for selected variables (temperature, sea-level, sea-ice, streamfunctions, meridional heat and freshwater transports). Forcing-induced differences are comparable in magnitude to model-observation misfits for most near-surface variables in the tropics and subtropics, but typically smaller at higher latitudes and polar regions. Forcing-derived differences are expectedly largest near the surface and mostly limited to the upper 1000m but can also be seen as deep as 4000m, especially in regions of deep water formation. Errors are not necessarily local in nature and can be advected to different basins. Results indicate that while forcing adjustments might suffice in optimization procedures of near-surface fields and at low-to-mid latitudes, other control parameters are likely needed elsewhere. Forcing-induced differences can be dominated by large spatial scales and specific time scales (e.g. annual), and thus appropriate error covariances in space and time need to be considered in optimization methodologies.

23

24 1. INTRODUCTION

25

26 Ocean general circulation models (OGCMs) aid in interpreting observations, assessing dynamics
27 and simulating past and future conditions of the ocean on a variety of space and time scales.
28 Uncertainties in the information derived from OGCMs can arise from deficiencies in the model
29 numerics and physical parameterizations, as well as from inaccurate initial and boundary
30 conditions (e.g., surface forcing fields). Quantifying and understanding these different sources of
31 ocean state uncertainty is important to improve data analysis and climate forecasting capabilities.

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33 The availability of globally gridded multi-decadal atmospheric reanalyses (*Kistler et al.*, 2001;
34 *Uppala et al.*, 2005; *Onogi et al.* 2005; *Saha et al.*, 2010; *Rienecker et al.*, 2011; *Dee et al.*,
35 2011) has stimulated many studies of the global ocean circulation on climate time scales.
36 Atmospheric fields from these reanalyses are commonly applied as a surface forcing of OGCMs.
37 Uncertainties in these forcing fields (e.g. *Chaudhuri et al.*, 2013) arise from many factors such as
38 differences in model set-up, assimilation schemes, data streams and cloud parameterization
39 schemes amongst others (*Milliff et al.*, 1999; *Sun et al.*, 2003; *Drobot et al.*, 2006; *Brunke et al.*,
40 2003; *Smith et al.*, 2001; *Nicolas and Bromwich*, 2011; *Wang and McPhaden*, 2001; *Zhang et*
41 *al.*, 1995). Comparisons of surface reanalysis fields against each other and against observations
42 have been conducted on global (*Wang et al.* 2011; *Xue et al.* 2012) to regional scales (*Bromwich*
43 *et al.* 2010; *Jakobson et al.* 2012; *Naud and Booth*, 2014). For example, zonal mean of 4
44 reanalyses for zonal wind stress (Fig. 1) suggest that they are quite similar. However the standard
45 deviations show that there is large variability at higher latitudes, with CFSR wind stress being

the largest. The CFSR wind stress product is in better agreement in terms of mean biases with the QuickSCAT climatology as reported by *Xue et al.* (2011). The spreads become larger for both mean and standard deviation in the case of precipitation flux (Fig. 1), especially in the equatorial regions. The general conclusion from all the above studies is that no single reanalysis does better than others for all the different variables.

Our approach is complementary to studies that have compared and contrasted the responses of different ocean models to a given set of forcing fields. For example, *Griffies et al.* (2014) forced 13 ocean models with the same forcing fields and reported that the models responded in a consistent manner. *Danabasoglu et al.* (2014) however reported significant differences in the Atlantic Meridional Overturning Circulation (AMOC) amongst 18 ocean models forced by Common Reference Ocean Experiment (CORE2) fields from *Large and Yeager* (2009). They tentatively attributed the large model spread to differences in internal ocean model parameterizations. In such uniform forcing experiments model-model differences can be due to internal model errors whereas model-observation differences can also include atmospheric state errors, experimental design limitations, and/or observational error or limitations (*Griffies et al.*, 2014).

In this study we attempt to quantify oceanic state uncertainties that derive solely from atmospheric forcing uncertainties and assess their relative importance as compared with overall model-observation errors. This work is in part motivated by the “Estimating the Circulation and Climate of the Ocean” (ECCO) effort (*Wunsch et al.*, 2009; *Forget et al.*, 2015a). ECCO aims at achieving a least squares fit of an OGCM to the great majority of meteorological and oceanic

observations by adjusting a control vector representing initial conditions for temperature and salinity, the surface atmospheric state, and internal model parameters (e.g., *Forget et al.*, 2015a, 2015b). In this context, the present study aims at addressing several important questions. How much of model-observation errors can be ascribed to errors in forcing fields? To what extent may forcing field adjustments suffice to reduce model-observation misfits to the expected level of data noise? Are differences in model simulations due to uncertainties in atmospheric forcing fields strongly dependent on location, depth, and time scale?

To this end, we employ a global OGCM to simulate the evolving ocean state over 2000-2009 under 4 different sets of atmospheric forcing fields. Comparing these simulations against each other and against observations permits an assessment of forcing-induced uncertainties in the context of overall model-observation errors. Details of the methodology are presented in section 2. Since the largest impact of atmospheric forcing errors is likely to occur at the surface, we evaluate model-observation differences in variables such as sea surface temperature (SST) and sea-ice cover (SIC) in section 3. We then analyze how errors propagate to the ocean interior by evaluating diagnostics of sea level anomalies (SLA), full depth temperature and oceanic transports in section 4. The spatio-temporal structure of the model errors is discussed in section 5. Our findings are summarized in section 6.

2. DATA and METHODS

For numerical computations we use the Massachusetts Institute of Technology general circulation model (MITgcm) (*Marshall et al.* 1997; *Adcroft et al.*, 2004), which is a general

purpose hydrodynamic model that solves the Boussinesq and hydrostatic or non-hydrostatic form of the Navier-Stokes equations for an incompressible fluid. The model uses a grid that reduces grid lines convergence and place grid poles on land. The fully global grid has a zonal spacing of 1° longitude. The meridional grid spacing is 0.3° of latitude within $\pm 10^\circ$ of the equator and increases to 1° latitude outside the Tropics. The model has 50 vertical levels ranging in thickness from 10 m near the surface to approximately 450 m at a maximum model depth of 6150 m. The partial-cell formulation of *Adcroft et al.* (1997), which permits accurate representation of the bathymetry, is used. The model is integrated with real fresh water surface fluxes and properly accounts for their effects on global mean sea-level.

The ocean model is coupled to a sea-ice model that computes ice thickness, ice concentration, and snow cover as per *Zhang et al.* (1998) and that simulates a viscous-plastic rheology based on *Hibler* (1979). The C-grid sea-ice code allows for no-slip and free-slip lateral boundary conditions. Ice mechanics follow a viscous-plastic rheology and the ice momentum equations are solved numerically using line-successive-over-relaxation (LSOR) (*Zhang and Hibler*, 1997). Ice thermodynamics are represented using a two-layer formulation that conserves enthalpy (*Losch et al.*, 2010). Ice concentration, volume, snow, ice enthalpy and salinity are advected using the multidimensional 2nd and 3rd-order advection scheme with flux limiters.

The ocean model setup uses third order advection with direct space time treatment, and vertical diffusive fluxes are treated implicitly. The model employs a turbulent kinetic-energy based scheme to parameterize vertical mixing (*Gaspar et al.*, 1990). Since eddies are not resolved at coarse horizontal resolution, eddy-induced mixing of properties along isentropes is

parameterized with the GM-Redi scheme. The Redi scheme applies a diffusion operator to mix properties along isentropes (*Redi*, 1982) and the GM scheme applies an advective flux to parameterize the transport effect of eddies (*Gent and McWilliams*, 1990). Laplacian diffusion and friction are used except for horizontal friction, which is biharmonic. No-slip bottom, free-slip lateral, and free surface boundary conditions are employed. Further details about the model configuration are presented in *Forget et al.* (2015a).

We use the MITgcm to simulate 4 ocean solutions for the past decade (2000-2009) using 4 reanalysis products: (1) Modern-Era Retrospective Analysis For Research and Applications (MERRA; *Rienecker et al.*, 2011), (2) Japanese Reanalysis (JRA25; *Onogi et al.* 2005), (3) European Reanalysis (ERA-Interim; *Dee et al.*, 2011) and (4) NCEP Climate Forecast System Reanalysis (CFSR; *Saha et al.*, 2010). Details of the reanalysis products are provided in Table 1. Surface boundary conditions for each simulation set (for each reanalysis product) are prescribed to the model using bulk aerodynamic formulae to calculate surface forcing fluxes. Each set of forcing fields is provided to the model in its native resolution and is then interpolated onto the model grid during runtime. Wind stresses are directly prescribed to the momentum equation. The surface boundary forcing fields (zonal and meridional wind stress, downwelling surface longwave and shortwave radiation, precipitation, air temperature and specific humidity at 2m) from each reanalysis product are prescribed every 6 hours. Each of the reanalysis products is based on its own boundary layer schemes, which bulk formulae can only approximate. Thus forcing errors assessed also include possible uncertainties in boundary layer physics, aside from those in the atmospheric state. A monthly mean climatology of freshwater discharge from land derived from *Fekete* (2002) is also prescribed for all 4 simulations. Initial conditions for all

simulations are the same and obtained from the ECCO state estimate (*Forget et al.*, 2015a) for December 31, 1997. Each simulation is run for 12 years (1998-2009). The first 2 years are not used and analysis is conducted on model diagnostics over 2000-2009. Internal model parameterizations remain the same while only the atmospheric forcing differs between simulations. The results provide a basis for evaluating model errors solely due to uncertainties in surface forcing.

Monthly-mean differences in model solutions for diagnostics of volume, heat and freshwater fluxes, meridional overturning circulation, sea-surface temperature (SST), sea-level anomaly (SLA) are presented to characterize the range of solutions associated with the choice of atmospheric forcing. Errors are partitioned into time-mean and time-variable components; respective uncertainties in surface atmospheric fields can have quite different magnitudes (*Chaudhuri et al.*, 2013). At each grid point, for a given pair of simulations, the time -mean differences (T_m) are computed by averaging the 120 monthly difference fields. The time-variable differences (T_v) are then computed by subtracting T_m and linear trends. These values are calculated for each of the 6 pairs that can be formed from the 4 simulations. The root-mean-square (rms) of the set of T_m estimates (respectively T_v) divided by $\sqrt{2}$ is denoted as M_m (respectively M_v). The $1/\sqrt{2}$ factor follows from assuming an equipartition of errors within pairs of simulations. This simplifying assumption is further discussed below. The main goal of this paper is to gauge M_m and M_v against estimates of overall model-observation error. The latter (denoted as D_m and D_v for the time-mean and time-variable components) is estimated by computing the rms of the 4 sets of model-observation misfits associated with the 4 model solutions. The details of the observational datasets used to compute D_m and D_v are provided in

section 3. D_m and D_v represent overall model-observation errors that may include model errors due to uncertain initial conditions, parameterized physics, and forcing fields as well as observational and representation errors. Cases of $M_m \ll D_m$ would thus indicate that forcing field uncertainties are of little relevance to explain model-observation misfits. Conversely M_m/D_m ratios of order 1 would indicate that reducing forcing field uncertainties is essential to reducing model-observation misfits. Forcing field uncertainties can induce differences in ocean simulations on a variety of time scales including long term model drifts (e.g., see *Griffies et al.*, 2009). For the relatively short 10-yr period of our analysis, the effect of linear trends is accounted for in M_m and D_m . Conversely linear trends (computed by a least-squares fit) are subtracted before computing M_v and D_v in order to focus on interannual and shorter time scales.

Treating model runs or retrospective data analyses as realizations of a random process is a commonly accepted method for assessing uncertainties in practice (e.g., see *Danabasoglu et al.* 2014, *Balmaseda et al.* 2015). This approach is the basis of the present paper as well. The $1/\sqrt{2}$ factor involved in the definition of M_m and M_v is simple to understand. Let p and q denote samples taken from a normal distribution with standard deviation σ and zero mean. Then the rms of $(p-q)$ will converge to $\sqrt{2}$. Simply scaling the rms of $(p-q)$ by $1/\sqrt{2}$ in our case is an ad hoc method though. Indeed the sample size (four atmospheric re-analyses) is rather small and, as a result, one cannot affirm that the distribution is normal or otherwise. The estimation of D_m and D_v is prone to similar issues. Due to the limited sample size the estimates presented here, while readily insightful, should therefore be taken with caution for the time-mean as well as the time-variable component. Furthermore errors that are common to all 4 reanalyses may not be clearly distinguished with our methodology.

A brief survey of global model-observation differences in SST and SLA between individual models (Table 2) suggests that their values are quite similar to each other thus further justifying our methodology of equipartition of the error. Furthermore, no one model seems to do better than other models.

3. ANALYSES of SURFACE VARIABLES

Uncertainties in surface atmospheric forcing are expected to affect strongly the upper ocean fields, which can be directly driven by the boundary fluxes of momentum, heat and freshwater. In this section we present diagnostics for selected surface variables (temperature and sea-ice), including comparisons of M_m and M_v values with D_m and D_v values where possible.

3.1. Sea Surface Temperature (SST)

The role of SST is essential for determining coupling and heat exchange between the atmosphere and ocean. Examining first possible bias errors in SST, values of M_m are largest ($> 1^\circ\text{C}$) in the Gulf Stream and Kuroshio boundary current systems (Fig. 2). Values of $M_m > 0.6^\circ\text{C}$ are also seen near the Agulhas Current, the Malvinas–Brazil Current, the East Australian Current and the Leeuwin Current. Values of M_m are relatively small over part of the eastern Pacific cold tongue, and at high latitudes. In comparison to M_m , M_v magnitudes are generally weaker (Fig. 2). The largest M_v values ($> 0.4^\circ\text{C}$) are again in the Gulf Stream and Kuroshio boundary current regions.

To compute corresponding D_m and D_v values, we use a SST product developed at the National Climatic Data Center (NCDC) (*Reynolds et al.*, 2007), which is directly comparable to our model

solutions since it covers 2000-2009 and is of 1° spatial resolution. This dataset is constructed by using optimum interpolation of SST retrievals. Values of D_m and D_v are largest ($> 1^\circ\text{C}$) in western boundary current regions. Moreover D_m and D_v also show large values along the ice edge, especially in the Greenland and Norwegian seas. Other regions such as the Peruvian upwelling region, the tropical Pacific cold tongue and portions of the Southern Ocean show high D_m and D_v values.

The ratios M_m/D_m and M_v/D_v , also shown in Fig. 2, are not negligible (> 0.5 or larger) over most of the global oceans. Forcing-induced errors in SST are thus likely important in the context of errors induced in the model by non-forcing related factors such as internal model parameterizations and missing physics. In regions with small ratios M_m/D_m and M_v/D_v (northern North Atlantic, the Arctic basin, the Antarctic circumpolar regions and the equatorial Pacific), factors other than atmospheric-related uncertainties must be relevant. Under the Arctic ice cover, SST seems very insensitive to surface forcing errors as expected. In the Southern Ocean, factors controlling the sea-ice edge and also the position of the Antarctic Circumpolar Current are expected to be important. The northern North Atlantic is a region of deep winter convection and thus model errors related to poor resolution of the convective processes are likely involved. The processes controlling the extent and intensity of the eastern equatorial Pacific cold tongue are not well understood (*Li and Xie, 2014*).

3.2. Sea-Ice

Sea-ice is a key component of the polar ocean physical system. It controls the exchange of heat, water, momentum, and gases at the sea surface. Changes in the albedo of the surface brought on

by changes in the ice cover over very large areas are a major factor in global climate change. The sea-ice cover has a seasonal nature in both hemispheres with maximum sea-ice extent witnessed in March in the Arctic (co-incident with minimum sea-ice extent in the Southern Ocean) and minimum sea-ice extent during September (co-incident with maximum sea-ice extent in the Southern Ocean).

Sea-ice fraction provides an estimate of the fraction of grid cells that are covered with sea-ice and range between 0 (no sea-ice) and 1 (totally covered with sea-ice). Values of M_m for Arctic sea-ice fraction (Fig. 3) are mostly < 0.1 with larger differences in the marginal ice zones. Estimates of M_v for sea-ice fraction are larger than M_m values in most regions of the Arctic especially in the coastal and marginal ice regions. The largest values (> 0.1) are seen around Svalbard.

Comparisons with data are done using the Bootstrap Sea Ice Concentrations from Nimbus-7 SMMR and DMSP SSM/I-SSMIS (Comiso, 2002) archived by the National Snow and Ice Data Center (NSIDC; we will use NSIDC sea-ice to refer to this dataset). Values of D_m are largest (> 0.3) around Svalbard and along the ice edge in the Greenland and Norwegian seas. Seasonal/coastal ice regions, such as the East Siberian, Beaufort, Laptev, Kara and Barents seas, show largest D_v values (> 0.2). Smaller M_m/D_m ratios (< 0.5) occur in the central Arctic, which has perennial sea-ice cover, whereas coastal and marginal ice regions show larger ratios (> 0.5). Similar spatial patterns are also seen in M_v/D_v values. Thus regions of marginal ice cover are

more sensitive to forcing differences than perennial ice-cover regions in the central Arctic (Fig. 3).

In the Southern Ocean (Fig. 4), values of M_m and M_v are relatively small (<0.2) and estimates of D_m and D_v show values > 0.2 along the continental margin. Ratios of M_m/D_m and M_v/D_v are low (< 0.5) in the coastal Indian Ocean sector and along 60°S in the Pacific Ocean sector of the Southern Ocean. The marginal-ice zones again display high ratios, which indicate that errors related to atmospheric forcing are particularly important in these regions.

Sea-ice thickness uncertainties in the Arctic are examined in Fig. 5. For comparison with data, we use sea-ice thickness estimates derived from ten campaigns of the Ice, Cloud, and land Elevation Satellite (ICESat) mission that span the period 2003–2008 (*Kwok et al.*, 2009). These campaigns represent estimates of the fall and winter Arctic sea ice cover. Each fall and winter ICESat campaign covers a 33-day period from roughly mid-October to mid-November (2003–2007), and from late February to late March (2004–2008) (*Kwok et al.*, 2009). We average these data into spring and fall bins for each year and compare them with the temporally consistent averages of sea-ice thickness from the model. Since we lack Southern Ocean ice thickness data, only Arctic results are presented in Fig. 5.

Values of M_m for fall sea-ice thickness (Fig. 5) are on the order of 0.7m and mostly greater than M_v values. The largest M_m values in the fall occur in the Canadian Archipelago. Large values of D_m ($> 1.5m$) are seen all along the coast starting from the Chukchi Sea and eastward towards Svalbard. This pattern is likely because the respective coasts usually become ice-free during the fall and model solutions do not resolve thin-ice conditions well. Both M_m/D_m and M_v/D_v ratios indicate that sensitivities to atmospheric forcing can amount to a substantial part of the model-observation differences in the central basin but not along the coastal regions.

4. ANALYSES OF DEPTH-DEPENDENT AND INTEGRATED VARIABLES

Apart from the impact on surface fields, atmospheric forcing errors can also propagate into the interior. In this section we quantify implications of uncertainties in surface atmospheric forcing to errors in sea-level, subsurface temperature, and heat, freshwater, and volume transports.

4.1. Sea-Level

For the time-mean sea-level, values of M_m (Fig. 6) are largest ($> 0.03m$) in the North Atlantic and Pacific subtropical gyres, as well as in the western tropical Pacific and the South Indian Ocean. For model-observation differences D_m , we use a mean dynamic topography derived from the Mean Sea Surface constructed by *Andersen and Knudsen* (2009) and the geoid constructed by *Pavlis et al.* (2012). The values of D_m are $> 0.04m$ in most regions and largest near boundary current regions. The ratio of M_m/D_m display extensive regions with values > 0.5 , indicating that

errors in time mean sea level associated with forcing uncertainties can be substantial, particularly away from regions of strong currents.

In terms of sea-level anomaly (SLA) where the time mean sea-level is removed from all products, values of M_v are > 0.02 m in the western basins of the northern hemisphere, i.e., the Gulf Stream-North Atlantic and Kuroshio-Oyashio current systems. However the largest M_v values (> 0.04 m) in the northern hemisphere are found along the coast of the Chukchi, East Siberian, Laptev and Beaufort Seas. To compute D_v values we use the Archiving, Validation and Interpretation of Satellite Oceanographic (AVISO) SLA dataset that combines sea surface height data from altimetry missions such as TOPEX/Poseidon, Envisat, Jason-1, and OSTM/Jason-2. We note that the $1/3^\circ$ binned product contains eddy and fine-scale features, which are not directly comparable to 1° model solutions. To partly overcome this issue, we apply a Gaussian smoother at 100 km length scale to the AVISO sea-level dataset. Values of D_v show spatial patterns similar to M_v , with high values in the Arctic shelves and western ocean basins. The ratio M_v/D_v is > 0.4 for most tropical regions. The sub-tropics and high latitudes show regions of interspersed low ratios (< 0.4) such as in the Argentine Basin. Thus SLA errors associated with uncertainties in atmospheric forcing can be a substantial part of the total SLA error in most regions.

4.2. Vertically Integrated Streamfunction (Ψ_b)

The M_m and M_v values for vertically integrated streamfunction (Ψ_b) are used to investigate first order changes in volume transports due to uncertainties in atmospheric forcing (Fig. 7). The largest M_m values (> 4 Sv) are seen in the Southern Ocean, which is likely related to the large

range in the time-mean zonal wind stresses at these latitudes (*Chaudhuri et al.*, 2013). The northern hemisphere western boundary current regions as well as the equatorial Pacific show M_m values on the order of 2 Sv, which likely relate to uncertainties in wind-stress given its relevance in determining Ψ_b . Values of M_v for Ψ_b are comparable to M_m values and are also largest (> 2 Sv) in the Southern Ocean and in the tropical Pacific. Since there are no observations of Ψ_b , we compared the M_v values to the signal (i.e. estimates of variability in Ψ_b from the 4 simulations). The resulting ratios (not shown) suggest that estimated uncertainties associated with boundary forcing are approximately 10% of the signal in most regions.

4.3. Meridional Overturning Streamfunction (Ψ_o)

Estimates of forcing-related uncertainties in Ψ_o (including bolus terms) are presented in Fig. 8. In the Pacific-Indian basin large values of M_m are seen in the near surface tropical overturning cells. The tropical cells, which are particularly intense in the Pacific, consist of subtropical water subducted in eastern areas of the ocean, flowing westward and equatorward in the upper pycnocline layers through both western boundary and interior pathways. The cell is closed by upwelling at the equator, with return flow to the subtropics in the surface Ekman layer (*McCreary and Lu* 1994; *Lu et al.* 1998). At deeper levels as seen in M_v the differences in Fig. 8 are likely due to the excitation of the equatorial waveguide by differences in the boundary current variability at annual timescales (*Jayne and Marotzke*, 2001; *Reznik and Zeitlin*, 2009) or surface winds.

The northern North Atlantic is a region of large winter convection and thus large values of M_m in this area may be related to differences either in wind stresses (CFSR winds are much stronger than other products in this region (not shown)) or in heat fluxes (JRA ocean heat loss is much stronger than other products (not shown)) imparted by the reanalyses. The equatorial Atlantic show weaker M_v compared to the Pacific-Indian basin. For the global meridional overturning streamfunction, other than features discussed above values of M_m are large (> 1.5 Sv) in the so-called ‘Deacon cell’ in the Southern Ocean (Fig. 8). The ‘Deacon Cell’ is a wind driven meridional cell in which deep water upwells, is blown north by wind stress in an Ekman layer, and sinks at the Antarctic convergence zone (Toggweiler and Samuels 1993; Tomczak and Godfrey 1994). The large difference is likely due to uncertainties in Southern Ocean winds. Zika *et al.* (2012) have suggested that the Deacon Cell is compensated by a combined effect of transient eddy fluxes and standing meanders. Estimates of M_v suggest high variability (> 1.5 Sv) in the Southern Ocean similar to M_m values.

4.4. Meridional Fluxes of Heat and Freshwater

Zonally integrated meridional heat and freshwater fluxes are depicted in Fig. 9. Mean heat transport shows good agreement amongst the 4 solutions with poleward transfer of heat from the equator on the order of 1.5 PW. Values of M_m for mean heat transport are generally below 0.1 PW (Fig. 9) and are thus on the order of 5% of the signal. The variability (standard deviation) in heat transport estimates are largest (> 3 PW) in the equatorial regions and taper out towards the higher latitudes (e.g., see Jayne and Marotzke 2001). Values of M_v are under 0.25 PW or on the order of 5% of the signal.

Mean freshwater fluxes show large differences between simulations, with M_m values greater than 1 Sv in southern tropical and sub-tropical latitudes. These differences are mostly comparable to the signal in the equatorial and sub-tropical regions and thus indicate a large forcing-induced spread. Variability in freshwater fluxes is also the largest in the equatorial regions and M_v values are approximately 10% of the individual variability displayed by the solutions (Fig. 9). The JRA forcing yields the largest interannual variability in freshwater transports. *Saito et al.* (2011) suggest that JRA can both overestimate and underestimate areas of precipitation in the 30°N–30°S sub-tropical domain. On the other hand *Wang et al.* (2011) report that CFSR has better time-mean precipitation distribution over various regions compared to previous generation reanalyses, leading to an improved representation of freshwater fluxes. *Trenberth and Fasullo* (2010) suggest that most reanalyses overestimate the hydrological cycle with the exception of MERRA. The larger spread amongst the solutions for freshwater transport in the sub-tropical regions can be attributed to large range in estimates of precipitation and evaporation in these reanalyses (*Chaudhuri et al.* 2013).

4.5. Vertical Profiles

Values of M_m and M_v for selected zonal and meridional sections of temperature are estimated to investigate the influence of uncertainties in atmospheric forcing at depth. Results indicate that for temperature, the largest differences are expectedly at the surface but substantial effects can also be felt at depth. For brevity, we show representative meridional temperature sections along 160°W in the Pacific and Arctic (Fig. 9). In this case, the estimates of M_m and M_v are computed based on monthly mean climatologies created from 2004-2006 from the simulation. This methodology is implemented to enable comparison with an observational climatology, which is a

blend dataset of OCCA (OCean Comprehensive Atlas; *Forget et al.*, 2010) complemented by World Ocean Atlas (*Locarnini et al.*, 2006) in the marginal seas and the Polar science center Hydrographic Climatology (PHC) in the Arctic.

The largest M_m values (> 0.8 °C) in Fig. 10 are seen in sub-tropical latitudes within the top 300m. In the Southern Ocean, however, errors on the order of 0.05 °C can be seen to propagate to levels as deep as 4000m. Values of D_m are mostly larger than M_m throughout the water column, and also show large discrepancies near the ocean bottom. Ratio of M_m/D_m suggests forcing-related errors dominate the near surface and are not totally negligible at depth. Values of M_v and D_v are > 0.1 °C in the upper 300m and mostly small in the deeper ocean. However the ratio of M_v/D_v (Fig. 10) display high values not only at the surface but also in the ocean interior in regions such as the Southern Ocean, and equatorial and sub-tropical Pacific.

In contrast with the Pacific, the Arctic shows larger M_m and M_m/D_m values at intermediate depths (below 200m) than at the surface (Fig. 10). This is related to the general hydrography of the region where the upper layer (about 50 m) is of lower salinity and lower temperature than the rest and remains relatively stable, because the salinity effect on density is bigger than the temperature effect (*Aagaard and Carmack*, 1989). The upper layer is also modulated by freshwater fluxes from the Siberian and Canadian rivers, which are the same in all our experiments (*Rudels et al.*, 1996). Water in the intermediate layer (200- 800 m), where the largest values of M_m and M_m/D_m are seen, is referred to as Atlantic Water. This water mass originates from inflow of the warmer North Atlantic Current through the northern North Atlantic,

which subsequently cools and sinks to form the intermediate Arctic layer, where it circles the Arctic Basin counter-clockwise (*Rudels et al.*, 1994; *Dickson et al.*, 2000). The Atlantic Water mass is mostly warmer than the surface water, and remains submerged only due to the role of salinity in density.

Values of M_m and M_v for salinity (Fig. 11) show similar structure to the temperature profiles with largest values at the surface, but also propagating into the ocean interior in the Southern Ocean. Values of D_m and D_v are expectedly larger at intermediate depths compared to M_m and M_v values. Ratios of M_m/D_m and M_v/D_v show a sub-surface maximum (albeit at different depths) in some regions of the Arctic similar to what was seen in the temperature profiles (Fig. 10).

5. SPATIO-TEMPORAL CHARACTERISTICS OF MODEL ERRORS

Knowing the spatial and temporal structure of model errors related to atmospheric forcing can be helpful in designing appropriate objective functions and optimization procedures. For example, large-scale spatially correlated errors could indicate similar characteristics in the forcing errors and suggest the need to design appropriate covariance matrices with both diagonal and non-diagonal weights. Similarly, in the case errors are dominated by specific time scales (e.g. annual cycle), it might be advantageous to optimize such time scales separately. The analysis of M_m values in previous sections clearly suggests that time mean biases show considerable large scale structures and should be considered as a separate term in objective functions. Here we explore further the nature of the time-variable errors, with examples focusing on SST and SLA fields.

To investigate the structure of the SST errors due to uncertainties in atmospheric forcing, we present an empirical orthogonal function (EOF) analysis on SST differences between the ERA- and JRA-forced simulations that provide a representative example. The first two modes of SST differences between ERA- and JRA-forced simulations (Fig. 12) together explain ~35% of the variance. Both modes show significant large-scale patterns dominated by features in subtropical and subpolar gyres and associated boundary currents, as well as with zonal circulations in the tropics and high latitudes. In addition, their respective time series indicate the presence of a dominant annual cycle.

Similar EOF calculations based on differences in SLA between ERA and JRA-forced simulations (Fig. 13) show similar behavior. The leading mode explains 21% of the variance and shows dominant patterns primarily at the tropics and high latitudes, modulated by the annual cycle. The second mode, which explains 13% of the variance, also has a near annual frequency and shows a dichotomy in the Arctic between coastal regions and the open ocean, suggestive of the dynamical regimes discussed recently by *Fukumori et al.* (2015) in relation to wind stress forcing. *Fukumori et al.* (2015) suggest that a near-barotropic fluctuation is present in the deep basins of the Arctic, which is driven by winds along the continental margins. The winds set up a divergence between the shallow coastal areas and the deep ocean basins, which is most discernable in ocean bottom pressure anomalies that display opposite signs in the two regions.

The SST and SLA examples highlighted in Figs. 12 and 13 suggest that model errors due to atmospheric errors can be correlated spatially on large scales, which may reflect broadly

correlated errors in forcing fields. In this case, one should try to specify both diagonal and nondiagonal terms in the covariance matrix used to weight the atmospheric forcing adjustments in optimizing procedures. Furthermore the strong contributions of specific time scales to atmospheric-related uncertainties suggest that separating the atmospheric forcing control vector into several different terms (e.g., mean, annual cycle and long term trend) would be beneficial, provided appropriate weights can be derived for each term.

6. SUMMARY AND FINAL REMARKS

In this work, we quantify differences in oceanic variables of interest derived solely from atmospheric uncertainties and investigate how they compare to total model errors (such as due to internal model parameterizations and initial conditions) and observational noise. We present atmospheric uncertainty effects at both the ocean surface level and at deeper depths. At the surface, uncertainties in atmospheric forcing tend to substantially affect SST and SLA estimates especially in tropical and mid latitude regions. However, at high latitudes non-atmospheric factors become more relevant. Cryospheric surface variables such as sea-ice extent are also affected by atmospheric forcing uncertainties especially in the marginal ice zones; however regions under perennial ice-cover are less sensitive to forcing related issues.

While errors due to uncertainties in atmospheric forcing tend to be largest near the surface and limited to the upper 1000m in most regions, there are exceptions where forcing related errors can reach depths of up to 4000m (e.g. Southern Ocean). Furthermore, these errors are not necessarily

local in nature and can be advected to different basins, such as in the case of temperature errors in Atlantic Water being advected into the Arctic basin (Fig. 10).

Forcing-induced errors show coherent spatial structure, as in the EOFs estimated from differences in SLA and SST (Figs. 12 and 13), likely reflecting similar behavior in the forcing errors. Thus, in optimization procedures such as described in *Wunsch et al.* (2009) and *Forget et al.* (2015a), adjustments in atmospheric forcing are best weighted by covariance matrices that include both diagonal and nondiagonal elements to account for respective error correlation structures. In addition, forcing-induced model errors can display strong modulation at the annual cycle (Figs. 12 and 13), pointing to possible advantages in partitioning the atmospheric forcing control vector into several different terms (e.g., mean, annual cycle and long term trend).

While our discussion mainly distinguishes errors due to atmospheric forcing against the bulk of non-atmospheric factors, quantifying the other contributions would provide a better framework to understand the various sources of model-observation differences and their significance. As a preliminary exercise, we conducted a 10-year simulation (2000-2009) that is initialized and set-up exactly as the ERA simulation but with different internal parameters for GM intensity, isopycnal diffusivity and diapycnal diffusivity. The internal model parameter values were obtained from a 20-year optimized solution (*Forget et al.*, 2015a). Preliminary assessment of model diagnostics to the original ERA run suggest that errors due to internal parameterizations are expectedly smaller than forcing-induced errors for surface variables such as SST and SLA. However parameterization based errors take precedence for variables such as sea-ice thickness (Fig. 5), Ψ_0 (Fig. 8) and temperature at depth (Fig. 10). Future investigations could incorporate

model sensitivities to initial conditions, bulk flux algorithms and other model parameterizations. Quantitative comparisons of results from this study with results from CORE2 experiments where atmospheric forcing is held constant but different ocean models are used to test model sensitivity (*Griffies et al. 2009; Danabasoglu et al. 2014*) would provide an indication of sensitivity to model parameterizations. Lastly results discussed here are derived from short 10-year integrations; longer integrations would be preferred in order to assess sensitivities to forcing errors at longer decadal climate scales.

ACKNOWLEDGMENTS

This study was supported by NSF grant ARC-1022733 and NASA grants NNX11AQ12G and NNX14AP33G. The authors would also like to acknowledge support from the NASA Physical Oceanography program for the ECCO project. Simulations were performed on the NASA Advanced Supercomputing (NAS) division's Pleiades supercomputer at NASA/ARC.

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TABLES

Reanalysis	Resolution	Range	Assimilation Scheme
MERRA	0.5°x 0.667° x 72 levels	1979-Present	GEOS -DAS
ERA	0.75°x 0.75° x 60 levels	1979-Present	4DVAR
JRA	1.125°x 1.125° x 40 levels	1979-Present	3DVAR
CFSR	0.5°x 0.5° x 64 levels	1979-Present	3DVAR

Table 1. Information about the reanalysis products used in this study.

	ERA	JRA	CFSR	MERRA
RMSD SST (°C)	0.48	0.47	0.46	0.47
RMSD SLA (m)	0.083	0.087	0.079	0.080

Table 2: Global root-mean-square differences between model derived SST and SLA against Reynolds SST and altimeter derived SLA.

FIGURE CAPTIONS

Figure 1: Top: Mean (left panel) and standard deviation (right panel) of zonally averaged zonal wind stress for the 4 reanalysis models. Bottom: as in top panels but for zonally averaged global precipitation flux.

Figure 2: Values of M_m (root-mean-square time-mean model-model difference), M_v (standard deviation of model-model difference), D_m (time-mean model-observation difference), D_v (standard deviation of model-observation difference), and respective ratios M_m/D_m and M_v/D_v (dimensionless) in sea-surface temperature (SST) [$^{\circ}\text{C}$]. Reference SST data is from *Reynolds et al.* (2007). Latitudinal extent from 90°S to 90°N .

Figure 3: Values of M_m , M_v , D_m , D_v and respective ratios in sea-ice fraction estimated in the Arctic Ocean. Reference data is NSIDC sea-ice concentrations. Latitudinal extent from 60°N to 90°N . Main Arctic seas are denoted as follows: BF-Beaufort Sea, CK-Chukchi Sea, ES-East Siberian Sea, LP-Laptev Sea, KA-Kara Sea, BA-Barents Sea, NR-Norwegian Sea, GR-Greenland Sea. Ice-free regions are masked in the ratios.

Figure 4: Same as Fig. 3 for the Southern Ocean. Latitudinal extent from 60°N to 90°N . Main Antarctic sectors are denoted as follows: IS-Indian Sector, PS-Pacific Sector, RS-Ross Sea Sector, BAS-Bellingshausen and Amundsen Sea Sector, WS-Weddell Sea Sector. Ice-free regions are masked in the ratios.

Figure 5: As in Fig. 3 but for values of M_m , M_v , D_m and D_v and their ratios in fall sea-ice thickness (m). Reference data is ICESat-derived sea-ice thickness.

Figure 6: Values of M_m , D_m , M_v and D_v and respective ratio in sea-level anomaly (m). Reference data for time-mean errors is mean dynamic topography derived from the Mean Sea Surface constructed by *Andersen and Knudsen* (2009) and the geoid constructed by *Pavlis et al.* (2012), whereas AVISO sea-level anomalies are used as reference for time-variable errors.

Figure 7: Values of M_m , M_v in vertically integrated streamfunction (S_v).

Figure 8: Log_{10} values of M_m and M_v for zonally averaged overturning circulation (S_v) including bolus terms for Pacific+Indian basin (PAC+IND), Atlantic Basin (ATL) and Global (GLO).

Figure 9: Estimates of M_m (upper panels) and M_v (lower panels) for zonally averaged meridional heat transport (left column), freshwater transport, and surface heat and freshwater fluxes (right column), for the 4 reanalysis models.

Figure 10: Log_{10} values of M_m and M_v and their ratio in a temperature section [160°W , 90°S : 90°N] across the Pacific. The surface landmass between 50 - 80°N is the Alaskan mainland. Reference data is from OCCA (*Forget.* 2010).

Figure 11: Same as Fig 10 for salinity.

Figure 12: The first two modes and principal components, along with the variances explained, of SST [$^{\circ}\text{C}$] differences between ERA and JRA reanalysis from 2000 to 2009.

Figure 13: The first two modes and principal components, along with the variances explained, of sea-level [m] differences between ERA and JRA reanalysis from 2000 to 2009.

Figure 1

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