Influence of assimilating transports and in situ data from the Rapid-MOCHA array into the GECCO2 ocean synthesis

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ABSTRACT

By assimilating information required for the estimation of the Atlantic meridional overturning circulation (AMOC) by the Rapid-MOCHA array, we investigate how transports should be constrained. For the period 2004–2011, we find that even the large adjustments in Florida Strait transport (FST) imposed by assimilating FST data do not impact the AMOC strength at 26.5° N while the AMOC away from this section changes due to the baroclinic response. Moreover, the high correlation between the FST and AMOC previously reported cannot be confirmed for this longer period. When assimilating FST and AMOC transports in conjunction, simulated transports can both be brought easily into consistency with the Rapid estimates while the representation of the hydrographic data at the mooring locations improves mainly at the eastern boundary. The dynamical constraint through the equations of motion conditions that the errors of the components are correlated and the total AMOC error is a much smaller than the sum of its components. Although Ekman and mid-ocean transports improve when AMOC is assimilated, the excellent AMOC representation relies on error compensation through adjustments of mainly the Ekman component. Assimilating the mooring data together with FST does not improve the representation of the AMOC. Density information is difficult to extract via assimilating temperature and salinity because of the strong density compensation in the subtropical gyre. Alternatively assimilating density from the mooring data directly was of limited success.

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1. Introduction

In mid-latitudes the oceanic heat transport of the Atlantic is carried by the overturning circulation conditioning its variability to play an important role in climate variability. Information from the Rapid-MOCHA project (Rapid in the following, Cunningham et al., 2007), which is now continued as RAPID-WATCH project, has widely been used to describe characteristics and the variability of the Atlantic meridional overturning circulation (AMOC, Rayner et al., 2011). For instance, based on the Rapid data Johns et al. (2011) showed that the overturning carries 88% of the total heat transport, and Kanzow et al. (2007) found indications for the traditionally assumed depth-independent compensation of the wind-driven surface flow. On seasonal time scales, the peak-to-peak amplitude of the AMOC was found to be 6.7 Sv with important contributions from the geostrophic mid-ocean transports (Kanzow et al., 2010). They also suggested that the seasonal cycle is dominated by wind stress at the eastern boundary while Chidichimo et al. (2010) showed that the changes of the eastern-boundary densities give rise to 5.2 Sv peak-to-peak transport variability. At the western boundary, the variability of the deep western
boundary currents (DWBC) was studied by Johns et al. (2008), who showed that the DWBC is – 26.5 Sv and is divided nearly equally between upper and lower North Atlantic Deep Water. The unusually low AMOC in the winter 2009/2010 leading to a 30% decline in the annual mean AMOC was attributed by McCarthy et al. (2012) to not only anomalous Ekman transports from December 2009 to March 2010 but also to the intensification of the geostrophic mid-ocean transports. The wealth of information from the Rapid array calls for a comparison with results from ocean models and the usage of this information in ocean syntheses.

In comparison studies, the AMOC time series of ocean syntheses and unconstrained runs compare well with the monthly mean AMOC from the Rapid array (e.g. Balmaseda et al., 2013; Köhl, 2015), partly due to the seasonal cycle. The seasonal cycle among other statistical properties is also found as one of the robust features of AMOC across different reanalyses while the year-to-year variability is found to be inconsistent (Munoz et al., 2011). Despite relative good agreement in their outflow from the Labrador Sea, different syntheses were found to perform differently at 26° N (Tett et al., 2014). The question appears whether ocean syntheses actually provide skill or at least an advantage over unconstrained model simulations. On interannual time scales, different synthesis products show only modest correlations with each other and correlations are actually found to be reduced in comparison to their unconstrained counterparts (Karspeck et al., 2015). The necessary criterion for a skillful estimate is thus better fulfilled by the unconstrained runs. The only clear advantage of the synthesis is their better agreement in mean and standard deviation with the Rapid estimates. A detailed comparison of a constrained and unconstrained run with respect to the performance in simulating the AMOC is provided by Smith et al. (2010). They find that assimilating in situ data improves the density structure and transports in the upper layer. However, since the density information projects mainly on the gyre circulation AMOC changes at 26° N remain small.

In a simplified ideal model experiment where the heatflux was the only control and driver of AMOC variability Brüdgam et al. (2013) demonstrated the general possibility to recover the AMOC by assimilating the appropriate observations with the adjoint method. Due to the lack of imposed wind stress variability, the AMOC excludes any wind driven component and, in particular, the Ekman component. Their study also revealed however that the existence of model errors leads to an error of 1.5 Sv and low correlation of less than 0.5 in the reconstructed AMOC variability.

Only few studies investigated the impact of assimilating data from the Rapid array. Although assimilating complete fields of salinity and temperature with a simple relaxation method showed that the AMOC time series can be reproduced reasonably well (e.g. Pohlmann et al., 2009; Dunstone and Smith, 2010), studies that use only data as sampled by the Rapid arrays proved constraining the AMOC with limited data to be difficult.

For instance, Stepanov et al. (2012) used a sequential method to assimilate temperature and salinity observations from the Rapid array at the eastern and western boundary. Various assimilation configurations were tested. Simply assimilating the data with their standard covariances lead to larger unrealistic changes away from the 26° N. Of all methods they tried out, the best result was obtained when only data below 900 m are assimilated but the information is spread via regression coefficients. Alternatively, Hermanson et al. (2014) used temperature and salinity covariances of overturning transport anomalies and demonstrated that the AMOC estimations can be reproduced when assimilating AMOC time series this way. However, the addition of in situ data was found to actually degrade the results.

A first attempt to assimilate Rapid data with the adjoint method was presented by Baehr (2010). She found that assimilating Florida Strait transports (FST) adjusts the time-mean value, while the short-term variability was barely affected because, in this particular implementation, the adjustments to the initial conditions were responsible for the impact on the solution. Using temperature and salinity from the moorings exclusively or jointly with FST data was shown to either lead to a reduced or a larger AMOC, respectively, while, again, the changes in variability remained small. It was also noted there that changes in the AMOC are more prominent away from 26.5° N but limited to a region of adjacent latitudes.

The results from the adjoint assimilation experiment were based on a 1-yr experiment. Baehr (2010) argued that longer time scales are needed to link open ocean and the FST variability. The longer time scales may open a prospect for adjusting, besides the mean, also the variability, particularly because the forcing will gain importance over adjusting the initial conditions. Now, with almost a decade of data available from the Rapid array, the time has come to test these presumptions and repeat some of her experiments and also to test additionally if the AMOC time series can directly be constrained. The seasonal variability of the FST is sensitive to the wind stress (Anderson and Corry, 1985) and at least two of the mechanisms by which the AMOC variability is affected are directly related to changes in wind stress (Köhl, 2005). It is expected, different from the results presented by Baehr (2010), that the shorter term variability of both the FST and the AMOC can in fact be adjusted by assimilating data from the Rapid array.

To this end, a series of experiments will be presented in the following to answer the questions: To what extent the observed FST and AMOC time series can be reproduced, and to what extent assimilating data from the Rapid moorings conjointly with the FST time series is able to constrain the simulated AMOC variability. Since different processes contribute to the AMOC variability, realizations of the same variability can be related to different underlying mechanisms. For those experiments that successfully reproduce the observed AMOC variability, it will be investigated if the associated characteristics of the AMOC are in agreement with the assimilated AMOC time series.

The structure of the paper is as follows. After presenting the assimilation framework and the configurations of the performed experiments in Section 2, the relation between the changes in FST and AMOC of the various experiments is discussed in Section 3. Sections 4–6 present the results of assimilating FST data solely or jointly with either the AMOC time series or in situ data from the moorings. After discussing the results in Section 7, Section 8 summarizes the results with conclusions.
2. Methods

The assimilation experiments performed use the framework of the GECCO2 ocean synthesis from the German contribution of the Estimating the Circulation and Climate of the Ocean project (GECCO). The synthesis uses the adjoint method to adjust the initial temperature and salinity in 1948 together with the air temperature, humidity, precipitation, and zonal and meridional wind every 10 days to bring the model into consistency with the data, which derive from the EN3v2 data base (Ingleby and Huddleston, 2007) and various sources of satellite data. The global model is based on the MITgcm (Adcroft et al., 2004), has 50 levels, and uses zonally 1° and meridionally varying higher resolution.

The GECCO2 synthesis covers the period 1948–2011. For the last iterations, the synthesis was partitioned into overlapping 5 year windows and results from the iteration 23 are described by Köhl (2015). Subsequently, the synthesis was further iterated until iteration 28. Since only the initial condition in 1948 is allowed to change in GECCO2, the atmospheric state is the only control for all windows except for the first one. The iteration 28 serves as reference and is denoted here as GECCO2. This iteration was also analyzed by Köhl et al. (2014) to investigate the impact of assimilating surface salinity data from the Soil Moisture and Ocean Salinity mission SMOS. Similar to their approach, we start the sensitivity experiments from iteration 28 and do not apply a fixed number of iterations but stop the iteration once the RMS difference to the Rapid data becomes smaller than the prior errors or if the progress stalled (typically but not strictly when the change in cost function of two consecutive iterations becomes smaller than 1%). Below, we count the additional iterations separately identifying iteration 1 with the iteration 28.

Data from the Rapid array available at the time of performing the experiments cover the period April 2004 to September 2012 and we performed experiments for the 8 year period 2004–2011 covered by a single assimilation window. The data constraint is as in Köhl (2015) plus additional constrains from Rapid as described now. Three different experiments were performed as listed in Table 1: In the first two we test how far and on what time scales the Rapid transport estimates can be reproduced. To this end, daily Florida Strait Transport (FST) from the Rapid array is assimilated additionally in the first experiment FLORIDA. Different from other assimilation methods such as that employed by Stepanov et al. (2012), adding further constraints is relatively simple with the adjoint method by adding further cost function terms that can contain all functions of the state variable of the model. With \( T_{\text{FST}}(\nu) \) as the FST depending of the meridional velocity \( \nu \), the additional term reads

\[
\sum_i (T_{\text{FST} i}^{\text{model}} - T_{\text{FST} i}^{\text{Rapid}})^2 / \kappa_{T_{\text{FST}}}^2
\]

where \( i \) denotes the day and \( \kappa_{T_{\text{FST}}}^2 \) the prior error of the FST transport. The adjoint method then seeks to minimize the sum of all terms with an iterative optimization procedure. After iteration 9, the optimization was stopped because no further progress was achieved during the last 3 iterations, which is suspected due to insufficient controls (10 daily corrections for daily data). For the subsequent experiments the period of the wind corrections was therefore decreased to daily.

FLORIDA, MOC assimilated additionally to the FST the daily AMOC at 1000 m from Rapid despite full depth AMOC data is available in order to have comparable weight and because the error profile of the AMOC data is not known.

With the last experiment we try to find out if assimilating the data used for the Rapid based AMOC estimate constrains the simulated AMOC. FLORIDA, TS assimilates instead of the AMOC time series daily temperature and salinity profiles from the Rapid moorings. The data contain the temperature and salinity profiles at 76.74° (at 26.51 and 26.52° N), 76.50° W, 71.97° W, 50.57° W, 41.21° W, and 16.23° W. The time series of the three nearby western moorings were averaged such that five different T and S profiles constrain the model at 26.5° N. Since substantial biases of GECCO2 with respect to the data remain and only a small correction of the model towards the Rapid data was observed in a first attempt, the time mean of the temperature and salinity data was replaced by the mean values over the same period from iteration 28. The mean AMOC of the model is nearly identical to Rapid 18.1 Sv vs. 17.6 Sv and we will focus in this experiment on changes of the variability.

As error weights for the FST and AMOC data, 1.0 Sv and 1.5 Sv were adopted from Larsen (1992) and Cunningham et al. (2007), respectively. The errors for the mooring data are the same as for the EN3 data and derive solely from the representation error of the model due to lack of mesoscale eddies. Similar to Baehr (2010), we enhance the weight of all Rapid data cost contributions by a factor 100, since otherwise their contributions relative to the total cost from global 3D data are negligibly small. For example, after the enhancement the FST contribution changes from about 10% to about 0.1% from first to final iteration. For the experiments that assimilate additionally only transports, the cost function contributions of the other components kept decreasing (only by a few percent). Differently, when in situ data from Rapid is assimilated the cost of the other components increased by a few percent.

| Experiment | FST | AMOC | T&S |
|------------|-----|------|-----|
| FLORIDA    | x   |      |     |
| FLORIDA, MOC | x | x   |     |
| FLORIDA, TS | x |     | x   |
Fig. 1. Evolution of the normalized RMS difference between Rapid and the optimizations for FST and AMOC depending on the iteration number for (a) FLORIDA, (b) FLORIDA_MOC and (c) FLORIDA_TS. (c) Shows additionally the RMS differences to the Rapid mooring data and for the cyan curve the time mean of the model-data AMOC differences is removed. The RMS differences are normalized by the prior errors which are 1.0 Sv and 1.5 Sv for FST and AMOC, respectively. The RMS term has the form \(1/n \sum_{k=1}^{n} (T_k - T_{obs})^2 W_k (T_{obs} - T_k)\) with \(T_{obs}\) the corresponding value from the model, \(W_k\) the inverse of the error and \(n\) the number of data. Note that the FST is assimilated in all optimizations but the AMOC only in FLORIDA_MOC.

Fig. 2. (a) Daily FST from Rapid, GECCO2 and FLORIDA. (b) Monthly mean AMOC at 1000 m from Rapid, GECCO2, and FLORIDA. All transports are in Sv.

3. Relation between FST, AMOC and the Rapid mooring data

The evolution of the RMS differences of the model to the Rapid FST and AMOC is shown in Fig. 1 for the three experiments as mean RMS normalized by the respective prior errors. Reaching a value of one is therefore necessary for a state consistent with the data and the prior errors. The large initial RMS of more than 12 reflects the huge model FST, which is more than 10 Sv larger in the model than from the cable measurements.

After assimilating FST data, the cost is reduced by nearly two order of magnitude (for comparison the cost function reduction achieved during the GECCO2 optimization was in the range 2–3) yielding a RMS value of about 1.5. The corresponding comparison of the model FST to the time series from Rapid shown in Fig. 2a shows the good match, in which only part of the high frequency variability differs. Note the rapid adjustment of the transport strength within only a few weeks, suggesting that fast wind driven processes are relevant for the transition. The available data after April 2004 show also practically no impact on the period where no data exist even though the adjoint could transport the information into this period. The effect on the correspondence of the simulated AMOC with the data remains remarkably low, given the strong adjustment in the FST (its normalized RMS increases slightly from 2.7 to 3.1). The corresponding AMOC time series shown in Fig. 2b reveals no adjustment of the mean and only minor changes of the variability, suggesting a weak relation between FST and AMOC. The strong FST of GECCO2 is thus probably not related to the larger and more realistic mean AMOC in comparison to the unconstrained run. The effect of the FST assimilation on the other components of the AMOC summarized in Tables 1 and 2 shows an improvement of the mid-ocean transport and the decrease in agreement with the total AMOC transport.

Adding Rapid AMOC data (FLORIDA_MOC) as constraint and increasing the frequency of the wind updates to daily allows for a reduction of both RMS values below one within only 5 iterations. To avoid fitting noise, the optimization was stopped despite further progress could be anticipated from the slope of the cost function evolution. This result is very different from the optimization performed by Baehr (2010), who found no sensitivity to the high frequency part of the FST and the experiments presented there showed mainly an adjustment of the mean values. The FST and AMOC time series from FLORIDA_MOC shown in Fig. 3a,b show a near perfect match to the Rapid estimate. However, analyzing the residuals reveals that they are not Gaussian noise but contain signatures of the seasonal cycle. Despite the excellent agreement in AMOC,
the mid-ocean transport improve only slightly over FLORIDA and mainly due to the better agreement between the means (Table 2). However, a noticeable improvement is seen for the Ekman component.

Adding temperature and salinity from the Rapid moorings (FLORIDA. TS) instead of the AMOC transports decreases the convergence to the FST data. Now, even after 15 iterations the RMS value stays slightly above one. Note that the FST cost contribution represents now only 20% of the Rapid data at initial time and decreases to less than 1% over the course of the iterations. At this stage, the model temperature and salinity have reached consistency with the Rapid data as the corresponding RMS value fell below one (red line). The bias removal is essential to this, e.g. for the EN3 data the RMS values remain between 2 and 3 (not shown). Although the cost function contribution related to the mooring data decreased by a factor of two, different from anticipated, assimilating data from the moorings together with the FST does not improve the correspondence to the AMOC data (the RMS actually increases by a few percent). That the mean of the temperature and the salinity is not adjusted due to the bias correction of the in situ data poses no limitation since the differences in mean show a negligible contribution to the cost (cyan curve). Visually, the agreement with the Rapid time series decreases noticeably (Fig. 4). Assimilating Rapid temperature and salinity is expected to improve mainly the mid-ocean transports. Although the total AMOC does not improve, the level of agreement of the mid-ocean transport increases and is comparable to FLORIDA. MOC while the correspondence between the Ekman components does not change (Table 2).

4. Assimilating FST

The FST is considered to be an essential part of the Rapid array and it was used for the first direct estimation of heat transport (Hall and Bryden, 1982), because it provides a very good estimate of much of the narrow western boundary current transport that is difficult to resolve by moorings. Moreover, FST fluctuations were brought in context of AMOC changes by Cunningham et al. (2007). Although constraining this transport alone seemed to be invaluable for the AMOC estimation, the relation between the FST and the 1000 m AMOC at 26.5° N shown above suggests no effect of FST assimilation on the overturning at all. Nevertheless, the RMS difference of the Rapid AMOC time series provides only information on a very specific aspect of the AMOC. More details of the basin wide changes in AMOC are shown as the mean and the variability in Fig. 5a,b. Particularly at 26.5° N, changes of the mean remain almost exactly zero at 1000 m while the upper cell of the AMOC, excluding the region near 26.5° N, decreases by up to 1 Sv except for the upper 500 m between 26° N and 40° N which shows an increase.

In comparison to Baehr (2010), patterns differ and the size of the response in FLORIDA is much smaller given that the FST changes are about 4 times larger than hers. These differences are mainly due to the different length of the runs. During the first year after Rapid data became available the structure of the signal is dominated in both experiments by a cell between 30° N and 40° N. Yet, the sign is opposite, which can be rationalized by the different means of adjustment; either changes in wind stress in our case or changes in initial conditions in hers. The peak of the variability of the changes of more than 3 Sv lies in a latitudinal band north of 26.5° N. This variability basically corresponds to a change from a positive cell in the early phase.

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**Table 2**

RMS (STD) differences to the monthly mean estimates of the Rapid transport components (in Sv).

| Experiment | FST     | Mid-ocean | Ekman   | Sum     | AMOC   |
|------------|---------|-----------|---------|---------|--------|
| GECCO2     | 12.0 (3.8) | 6.7 (4.6) | 1.5 (1.4) | 20.4 (9.6) | 3.3 (2.7) |
| FLORIDA    | 0.6 (0.6)  | 5.9 (3.7) | 1.5 (1.5) | 8.0 (5.8)  | 3.9 (3.2) |
| FLORIDA. MOC | 0.5 (0.5)  | 3.8 (3.3) | 1.1 (1.0) | 5.3 (4.8)  | 0.9 (0.8) |
| FLORIDA. TS | 0.8 (0.7)  | 3.9 (3.9) | 1.5 (1.5) | 6.2 (6.0)  | 3.4 (3.4) |

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**Fig. 3.** Daily (a) FST and (b) 1000 m AMOC from Rapid, GECCO2 and FLORIDA_MOC. The curves for FLORIDA_MOC and a second version of Rapid are shown with an offset of −15 Sv. All transports are in Sv.
Fig. 4. Daily 1000 m AMOC from Rapid, GECCO2 and FLORIDA. TS. The curves for FLORIDA. TS and a second version of Rapid are shown with an offset of –15 Sv. All transports are in Sv.

Fig. 5. Mean (left column) and STD (right column) of the AMOC changes in Sv. (Top) FLORIDA relative to GECCO2, (middle) FLORIDA MOC relative to FLORIDA, FLORIDA . TS relative to FLORIDA.
to a negative cell in the later stage. A second cell of highly variable changes is focused on the equator. The time-dependent response in AMOC is related to the baroclinic response to the wind stress curl changes.

In order to further investigate how the model is adjusted to change the mean FST by more than 10 Sv, the mean zonal and meridional wind stress is shown in Fig. 6. Classically, the mean FST is related to the wind stress curl at the latitude of the Florida Strait via the non-topographic Sverdrup balance due to the adjustment process by baroclinic Rossby waves (e.g. Anderson and Corry, 1985). Positive wind stress curl changes are in fact visible at that latitude in the eastern part of the North Atlantic while wind stress curl changes near the western boundary remain smaller. However, additional changes of equal strength exist in the entire basin north of the Florida Strait. Since the constraints from data other than Rapid are still active during the optimization and contribute 90% of the cost, most of the complex basin wide wind stress changes are unrelated to the FST changes.

In order to match the Gulf Stream separation in GECCO2, a strong enhancement of zonal wind stress was observed roughly over the Gulf Stream (Köhl, 2015). The coastal region near the Gulf Stream area was identified by Anderson and Corry (1985) as being most relevant for the seasonal variability of the FST. On longer time scales of several years, this region of high sensitivity of the FST to wind stress moves eastwards, as baroclinic Rossby waves carry the signal westward, but remains west of the Mid–Atlantic Ridge (Czeschel et al., 2012). Since for the mean transport all processes affecting the transports on short and long time scales are important and convoluted to yield the effect on the mean, it is difficult to estimate from the time evolving sensitivities shown by Czeschel et al. (2012) how and where the mean wind affects the mean transport. Therefore, the sensitivity analysis is repeated here with the current setup to show the sensitivity of the mean FST to the mean zonal and meridional wind. Due to stability constraints, the sensitivities used in the application of the adjoint method for the GECCO2 synthesis are only approximations since submodels, such as the sea ice and the mixed layer model or the eddy parameterization, are removed from the adjoint. Moreover, the necessary additional higher viscosity in the adjoint leads to a systematic underestimation of the amplitudes.

The time period of the adjoint run is the same 8 years as analyzed here and the background state is from GECCO2. The FST mean is calculated over the same period as data is available from Rapid. Fig. 7 shows that the largest sensitivities of the mean FST to the wind are located directly in the vicinity of the Florida Strait with a center of the curl slightly northeast of the Strait at 77°W, 27°N, where the sign changes rapidly. Bands of positive and negative sensitivity to zonal wind extend toward the middle of the basin. Bands of alternating sign also exist over the Gulf Stream. Although the signatures of the patterns of high sensitivity can be identified in the wind stress corrections shown in Fig. 6 (note the opposite sign since the FST is reduced in FLORIDA), they are among the smallest signals there. In the eastern basin, only the wind stress parallel to the coast is relevant for changing the mean FST. Nevertheless, in the regions of large sensitivity of FST to wind changes the changes reach the size of the mean wind, and the multiplication of the sensitivities with the changes yields close to 7 Sv FST change.

Fig. 6. Temporal mean of the (a) zonal and (b) meridional wind stress difference FLORIDA – GECCO2 for the period April 2004 to April 2011 in N/m².

Fig. 7. Temporal mean of the sensitivity of the April 2004–December 2011 mean FST to (a) zonal and (b) meridional wind for the period April 2004–April 2011. The units are those of the gradient of the FST with respect to the wind, thus (m³/s)/(m/s).
The wind stress corrections entail a displacement of the Gulf Stream position which can be seen in Fig. 8. The figure also shows the changes in the barotropic stream function and that the few degrees shift of the position is connected with a large (40 Sv) change in stream function. In comparison to the changes in barotropic stream function presented by Baehr (2010), we find a similar pattern with a larger amplitude and a reversed sign consistent with the different changes in FST in our compared to her experiments. Besides the shift in position, the Gulf Stream transport strength is reduced by 10 Sv, which is in agreement with the perception of the FST being part of the Gulf Stream flow.

By analyzing one year of Rapid data, Bryden et al. (2009) found that 80% of the Gulf Stream transport variations carry through to AMOC variations. Although we can confirm a high correlation of nearly 0.8 for the 10-day filtered time series of the same period 1 April 2004 to 1 April 2005, for the complete time series assimilated here, we see only correlations of 0.42 in agreement with similar correlations for the experiments presented here. Hence, the FST is much less related to the AMOC than previously argued for.

The correct Gulf Stream separation is connected with the strong recirculation south of the Gulf Stream, which in reality is related to an inertial recirculation. (Marshall and Nurser, 1988) argued that the winter time strengthening of the inertial recirculation leads to a southward shift of the Gulf Stream position. In the GECCO2 coarse resolution model, the correct position and strength is directly forced by enhanced wind stress but the strength of the FST also determines the strength of the Gulf Stream and thus its position as shown above. Although the FST sensitivities are weak over the Gulf Stream, the corrections in GECCO2 are very strong there, such that the wind forced recirculation entails changes in the Florida Strait and contributes to the much too large a FST. In FLORIDA, the FST is corrected at the expense of a less realistic Gulf Stream separation. Note that the connection between the FST and the wind stress over the Gulf Stream is not robust and may depend on details of the topography of the Bahama Archipelago or the exact position of the wind stress corrections over the Gulf Stream, since the first ECCO solution (Köhl et al., 2007) does not show a FST intensification despite similar wind stress corrections as GECCO2.

5. Assimilating FST and AMOC

Different from the FST, the mean and STD of the AMOC are not much different between GECCO2 and Rapid (18.1 Sv and 3.9 Sv versus 17.6 Sv and 3.9 Sv, respectively). Note that GECCO2 here includes additional iterations and is therefore not identical to Köhl (2015). Assimilating additionally the Rapid AMOC time series leads therefore only to small additional changes in the mean. Fig. 5c shows that the small reduction of 0.5 Sv is mainly restricted to 26.5°. Although the AMOC changes are basin wide, changes with STD larger than 1 Sv are located in a narrow band at 26.5°. Such very localized AMOC changes are related to directly wind driven mechanisms (Köhl, 2005) which include the Ekman transport and coastal jets though vertical motion of the isopycnals. The monthly GECCO2 AMOC time series compares less well with Rapid than the original after 23 iteration (r = 0.70 instead of 0.75), while assimilating Rapid AMOC increases the correlation to r = 0.93 (Fig. 9a). Although the FST transport is matched within its prior error, the monthly time series show some differences, e.g. that the minima in the first half of the period are less deep.

For the Rapid estimate, the Ekman transport is calculated from QuikSCAT winds, which are also assimilated in the NCEP RA1 product that provides the background information on the wind for all experiments presented here. The correlation
between QuikSCAT and NCEP RA1 is in fact very high while assimilating data reduced the correlation to QuikSCAT in GECCO2. The Ekman transport of the unconstrained run therefore agrees better with Rapid than GECCO2 despite the better agreement of the GECCO2 AMOC with the Rapid AMOC (Köhl, 2015). However, assimilating Rapid AMOC improves the correlation of the Ekman transport to Rapid from $r = 0.87$ to $r = 0.91$ (Fig. 9c) and reduces the RMS error (Table 2). The zonal wind stress changes relative to FLORIDA show contributions along 26.5° related to the Ekman transport but are also dominated by larger changes north of 26.5° N (not show), which according to the sensitivities of the AMOC to wind stress shown by Köhl (2005) should not have a connection to the AMOC changes and are relate to the no-Rapid data that still dominate the costfunction.

Due to the adjustment of the FST, the mid-ocean transport also strongly adjusts to lower values. However, the time scale of the adjustment is slower than for the FST and it takes more than 6 months to reach the values of Rapid (Fig. 9d). As in GECCO2, the variability is still dominated by a seasonal cycle and higher frequency changes visible in Rapid are not matched, the correlation increased slightly from $r = 0.53$ to $r = 0.61$ and the agreement with Rapid remains therefore surprisingly low given the excellent AMOC match (Fig. 9a). Despite the too large FST, Köhl (2015) argued that the western boundary current transport of GECCO2 is more realistic because the transport of the Antilles Current, which is absent in the unconstrained run, appeared as part of the FST in GECCO2. In FLORIDA MOC the FST is reduced and the Antilles Current remains missing. This indicates that the high FST of GECCO2 is more likely a result of a lack of constraints on this transport than a necessary condition for representing the assimilated data.

AMOC variability can be related to different mechanisms. Köhl (2005) identified with and an adjoint model technique four main processes, which are, apart from the already mentioned wind driven Ekman transport and near coastal heaving of isopycnals, boundary waves and westward propagating Rossby waves. The relevance of these mechanisms have been identified in GECCO (Köhl et al., 2007), and Hermanson et al. (2014) find similar mechanisms based on a correlation analysis. Although assimilating a single AMOC time series cannot guarantee the correct representation of all four processes, Ekman transport time series match well and all other processes affect the density near the boundaries, such that an evaluation with the Rapid temperature and salinity data is useful. Fig. 10 shows the RMS difference between GECCO2 and Rapid for temperature, salinity and the resulting density at the mooring locations. Note, the differences in mean were removed. The largest differences exist at the eastern and western boundary reaching 0.8 °C and 0.12–0.18 PSU mostly in the top 1000 m. At the eastern boundary larger differences reach down to 1500 m. For the density differences, GECCO2 shows the best agreement with the mid-ocean data. Since the AMOC is mainly determined by the east–west density difference, there is an ongoing debate whether the mid-ocean data are important. The good agreement there suggests the least impact when assimilated. Due to large density compensation in the subtropical gyres, the resulting density differences are much smaller than what the sum of the effects of temperature and salinity RMS values would suggest. However, they are larger than what results from transforming the temperature and salinity errors into vertical displacement errors (not shown).
that kg/m³ at locations adjustment of cell related of its differences.

**Fig. 10.** Vertical profiles of RMS differences between GECCO2 and Rapid data for (a) temperature in °C, (b) salinity in PSU and (c) the resulting densities in kg/m³ at the locations of the moorings as indicated in panel (a). The bold purple curve is the RMS difference of the density difference 71.5–16.5° N. Note that the time mean residuals were removed prior to analysis. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

**Fig. 11** shows the reduction in RMS differences for FLORIDA MOC and FLORIDA in the upper and middle panels. Both experiments show for temperature and salinity mainly an effect at the eastern boundary. Much of the reduction in RMS of FLORIDA MOC is already visible in FLORIDA and thus probably related to the assimilation of FST; the small improvement of temperature on the western boundary even disappears. Small density improvements are restricted to the top 750 m and show up on the eastern and, albeit smaller, on the western boundary. Improvements for FLORIDA includes the 71.5° W location while for FLORIDA MOC only near surface values change substantially at that location. Although the costfunction related to AMOC is reduced by more than a factor of three, the RMS reductions in temperature and salinity at the Rapid moorings remain below 25% and reductions of the density error are with up to 30% only slightly larger.

There can be compensating errors of density between east and west, and the AMOC assimilation is expected to improve mainly the east-west density difference. The cyan curve added to **Fig. 11** shows that the RMS differences in the east-west density gradient are in fact considerably reduced. Although the match in density difference is dynamically expected, the lack of improvement of the individual profiles suggests that the simulated improvement of the AMOC variability is apart from the Ekman component not entirely consistent with the observations.

### 6. Assimilating FST and Rapid temperature and salinity

Assimilating the Rapid AMOC time series improved the correspondence of the density only slightly over FLORIDA despite its large improvement of the AMOC. The above described relation between AMOC and FST from the complementary experiment FLORIDA TS, in which Rapid temperature and salinity are assimilated, showed that an improvement in representation of the hydrography does not imply improvement of AMOC (**Fig. 1c**). Yet, the changes in the basin wide AMOC are considerably larger than those introduced by assimilating the AMOC time series and consist overall in a reduction of the upper cell by 0.5 Sv while locally between 26.5° N and 40° N changes can be up to −2.5 Sv. In the near surface area at about 20° N and southward of this latitude, an enhancement is visible. The pattern of the variability of the changes shares some similarity with that of FLORIDA, but the associated time variability is different and reflects oscillatory modes due to baroclinic adjustment processes with faster oscillations closer to the equator.

**Fig. 11a,b** shows that the improvement at the mooring locations is mainly at the eastern boundary where most of the RMS differences are removed below 300 m. However, in the layers above 300 m the differences increase. Changes at the other locations are with little more than 10% minor. At the eastern boundary, density compensation seems to play a major role since the impact on the density remains absent while the size of changes in temperature and salinity RMS would suggest a dramatic improvement.

The degree of density compensation can be studied by the Turner angle, which is defined after Ruddick (1983) from the linear density equation as

$$T = \arctan \left( \frac{\alpha \Delta T - \beta \Delta S}{\alpha \Delta T + \beta \Delta S} \right)$$

where α and β are the thermal expansion and haline contraction coefficients. Turner angles between 45° and −45° indicate no density compensation while the angles beyond this range indicate partial compensation with complete compensation at
$\pm 90^\circ$. At positive (negative) angles temperature (salinity) perturbations are governing the density perturbations. To focus only on the degree of compensation, we calculate the averages of absolute values of Turner angles.

The most efficient means of changing the local density, particularly on short time scales, is by wind stress curl changes via Ekman pumping. In the subtropics, where warm and salty water in the upper layers are paired with fresh and cold deeper layers, turner angles are large. For the Rapid mooring locations, Fig. 12 shows as thin lines angles for vertical displacements of around $60^\circ$ with maxima around $70^\circ$ at 500 m, 1500 m, and 3000 m and minima around $50^\circ$ in between. In the western basin, averaged Taylor angles based on changes introduced by assimilating Rapid temperature and salinity, closely follow this typical shape. However, the eastern basin shows larger compensations that are larger than $80^\circ$ between 800 and 1800 m. It seems that in the region of the largest temperature and salinity adjustments mainly density compensated changes are
admitted by the assimilation procedure that do not directly change the dynamics of the model. Nevertheless, the east–west density difference improves in the upper 700 m by typically 0.06 kg/m³ (purple curve), which is about 30% of the RMS difference.

7. Discussion

In comparison to the large adjustment of the FST, the adjustments of the other components of the AMOC are more subtle. Table 2 summarized the RMS and (to disregard differences in mean) the STD of the monthly mean transport differences to the Rapid estimates. While assimilating FST leads to the same good match between model and Rapid FST in all experiments, the impact on the mid-ocean transports is different. Adding just FST data changes the mean from a 5 Sv larger to a 4 Sv lower transport than the Rapid estimate and improves the STD by nearly 1 Sv. Although the east–west density differences provide information on the strength of the mid-ocean transports, the STD of FLORIDA. TS is with 3.9 Sv at the same level as FLORIDA, in comparison to the 3.3 Sv of FLORIDA. MOC. However, the RMS values of FLORIDA. TS and FLORIDA. MOC are at the same level because the mean value of FLORIDA. TS is almost identical to Rapid while FLORIDA. MOC is nearly 2 Sv too low.

Although only for FLORIDA. MOC the Ekman component improved by 0.4 Sv, the reason for FLORIDA. MOC matching the AMOC time series, while FLORIDA. TS failed to improve the AMOC time series, can not be rationalized from the slightly better representation of the components in FLORIDA. MOC. In FLORIDA. MOC the error of the upper mid-ocean transport is still much larger than even the prior AMOC error of 1.5 Sv. Thus, the components are adjusted in a way that errors compensate and the error of the sum of the components is much smaller than the sum of the errors (Table 2). When FST is assimilated, the sum of the errors (without the bias of means) is typically two times the AMOC error, while it is six times for FLORIDA. MOC. However, errors are expected not to add since the components are linked through the dynamical constraint of the equations of motion. This reduction in error actually demonstrates the key benefit of combining models with observations by data assimilation. This advantage also holds for purely atmospherically forced models and while GECCO2 as described by Köhl (2015) showed a slightly better correspondence to the Rapid AMOC than its unconstrained counterpart, for the later iteration used here the advantage has almost completely disappeared.

The match with the Rapid AMOC of FLORIDA. MOC has to be considered artificial because the individual components do not improve in the same way as the total does. Kanzow et al. (2010) and Zhao and Johns (2014) have shown that much of the seasonal cycle is driven by the Ekman forcing and wind stress curl near the boundary and the importance of the coastal–parallel wind was also described by Köhl (2005). The fast reaction of the AMOC changes is thus likely be related to wind stress changes while the Ekman transport provides a means to change an AMOC component without much impact on the other components. It is thus suspected that the changes in the Ekman component are the main agent to bring the model AMOC into consistency with the Rapid estimate. This hypothesis can be tested by replacing the Ekman transport contribution to the AMOC in FLORIDA with that of FLORIDA. MOC. As a result, the difference of the resulting AMOC to Rapid is largely reduced and STD of the difference is only 0.5 Sv larger than for FLORIDA. MOC. It is therefore mainly the Ekman transport that is used as a control, which also explains why the adjustment to the mid-ocean transport remains small.

On the other hand, wind stress changes near the boundary lead to vertical motion of the isopycnals and provide means to change the geostrophic component. On seasonal time scales, the importance of the geostrophic component was emphasized by Kanzow et al. (2010) and much of the adjustments to bring temperature and salinity into consistency with the mooring data can be explained by vertical motion (Fig. 12). In fact, density changes introduced by assimilating the mooring data are highly correlated with changes of the vertical velocity ($r \approx 0.6$). However, a simple relation to the Ekman pumping or coastal...
parallel wind does not not hold for the changes at any of the mooring locations (typically $r < 0.2$ for both cases, except for 71.5° N where $r \approx 0.3$). A simple local relation may not be appropriate to explain changes introduced by the adjoint method. For instance, for the Pacific mooring ALOHA Stammer et al. (2008) have shown with the adjoint modeling framework that over periods of several years large regions affect the temperature and salinity changes at the station. At large, the impact of the desity changes on the realism of the AMOC, in particular its mid-ocean transport component (Table 2), remained small and appeared not to be an important means to change the AMOC.

Different from the adjoint estimate, the Rapid error estimate should result from the sum of the errors of the components since the estimates are independent and their errors should be uncorrelated. With this aspect in mind, the estimate of prior error for the Rapid AMOC will be reconsidered. It is reasonable to assume that the accuracy of the mid-ocean transport estimate is in the same range as that of the FST (1 Sv). Since the NCEP reanalysis actually assimilates QuickSCAT wind, the error of the Ekman transport is probably not smaller than the difference between the estimates from Rapid and the NCEP wind (0.9 Sv). For uncorrelated errors the AMOC error follows as the sum and is likely to be close to 3 Sv and therefore in the same range as the STD of AMOC difference Rapid – GECCO2. For an error of the Rapid AMOC in the range of 3 Sv, it is understandable that assimilating AMOC improves its components only slightly. However, since the density information at the mooring locations is considered sufficient for the estimation of the mid-ocean transport, the lack of improvement of the mid-ocean transport in FLORIDA. TS in comparison to FLORIDA remains surprising and may be related to the fact that density improves only in the top 700 m and by not more than 30%. Reasons include density compensation in combination with large representation errors of the model. The difficulty to constrain the mid-ocean transport and the resulting remaining large error in comparison to Rapid is consistent with the finding of Bridgman et al. (2013) that the presence of model errors leads to large errors in the reconstruction of the thermohaline AMOC component.

8. Conclusions

In the present paper we address the question: how the data from the Rapid array should be used in the context of a variational data assimilation method to constrain the AMOC most efficiently. Different from previous experience with assimilating Rapid data by the adjoint method (Baehr, 2010), we find that the simulated transport time series of FST and AMOC can easily be brought into consistency with the data on all time scales. The importance of the adjustment of the forcing fields, which did not play a role in the previous attempt to assimilate FST, and in case of the AMOC, the direct assimilation of the AMOC transports are the main reasons for this difference. For the adjustment of the AMOC and FST fast wind driven processes are most relevant. Since the variability is hardly affected by the initial condition as shown by Baehr (2010), changing wave processes such as baroclinic Rossby or boundary waves are not important means to change transports on time scales of less than a year. However, in agreement with her finding it was found that assimilating in situ data from the Rapid array together with FST is not suitable for constraining the AMOC time series.

Given the errors of the individual components of the Rapid AMOC, it appears that the STD of the difference GECCO2 – Rapid AMOC may be close to the total error of the Rapid AMOC. Yet, the mid-ocean transport did not improve when assimilating mooring data. Large representation errors paired with density compensation are seen as the main reason for this difficulty.

The question remains if it is useful to implement the constraints on the transports and/or the Rapid data. Since all experiments shown here use an enhanced weight of the Rapid data relative to the other data products, the effect of the Rapid data might be smaller with a realistic weight. An implementation of the FST constraint right from the start of the iteration is, however, likely to prevent the drift towards the state with very large FST even if the weight is much smaller, because the FST appears not to be well constrained in GECCO2. Moreover, assimilating FST is shown to improve the temperature and salinity representation at the eastern boundary.

For the AMOC time series the answer is less clear. Mean value and STD of the GECCO2 AMOC are quite realistic and the AMOC may already be consistent with the Rapid estimate. An additional positive effect on the Rapid hydrography is small. It is argued that the obtained high agreement with the Rapid AMOC is artificial and mostly achieved by modifying the Ekman component since the errors of the mid-ocean transports remain inconsistent with the prior error. Including the AMOC constraint is in agreement with the recommendation by Tett et al. (2014) not suggested, particularly because the most important means to evaluate simulated AMOC will become unusable.

Assimilating densities from the moorings directly could avoid the problem of density compensation. Moreover, considering the effect of density compensation, when calculating the representation error for density, yields substantially smaller errors for densities than directly applying the linear density equation to temperature and salinity errors. In a first attempt to directly impose density changes via the assimilation procedure, density from the moorings was either additionally or alternatively to the temperature and salinity assimilated. However, the reduction of the Rapid data related cost reduced only by 10–20% possibly due to the complexity of the associated costfunction. Assimilating density differences between the east and west could be another option. However, the model seems to be fairly rigid against changing it’s mid-ocean transport. Before performing such experiments it is useful to first test whether the model’s mid-ocean transport can be brought into consistency with the Rapid estimate by directly assimilating the respective Rapid time series.

The difficulty to constrain the AMOC by the mooring data also holds for other methods (Stepanov et al., 2012). Although consistent with the low agreement among syntheses, the lack of impact of the Rapid data does not explain the large differences found between assimilation products and forced models (Munoz et al., 2011; Karspeck et al., 2015) because all other assimilated data could be more useful than Rapid data to constrain the AMOC. For instance, Stepanov et al. (2012) claims an
advantage of assimilating EN3 prior to their experiment. In this respect Köhl (2005) has constructed the most useful data distribution for constraining the AMOC with the adjoint method. Areas of data with large impact on the AMOC are located over wider areas along the boundaries, which is consistent with the findings of Stepanov et al. (2012) who sees an advantage of spreading the information along the boundary current. It is thus not granted that the data distribution most useful for a direct calculation of the AMOC is also most useful for being assimilated.

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