

This discussion paper is/has been under review for the journal Ocean Science (OS).
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Transports and budgets in a 1/4° global ocean reanalysis 1989–2010

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Received: 27 December 2011 – Accepted: 9 January 2012 – Published: 23 January 2012

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Published by Copernicus Publications on behalf of the European Geosciences Union.

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9, 261–290, 2012

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Large scale ocean transports of heat and freshwater have not been well monitored, and yet the regional budgets of these quantities are vital to understanding the role of the oceans in climate and climate change. In contrast atmospheric heat and freshwater transports are commonly assessed from atmospheric reanalysis products, despite the presence of non-conserving data assimilation based on the wealth of distributed atmospheric observations as constraints. The ability to carry out ocean reanalyses globally at eddy permitting resolutions of $1/4^\circ$ or better, along with new global ocean observation programs, now make a similar approach viable for the ocean. In this paper we examine the budgets and transports within a global high resolution ocean model constrained by ocean data assimilation, and compare them with independent ocean and atmospheric estimates.

1 Introduction

Ocean reanalyses or ocean syntheses have the potential to provide useful complete time-evolving descriptions of the ocean state, and particularly the ocean circulation, which cannot be obtained in any other way given the disparate and asynoptic nature of the ocean observational data sets. In the atmosphere reanalysis activities have become of great importance to research and it has become commonplace to use atmospheric wind patterns from reanalysis products to investigate regional phenomena and processes e.g. heat, water vapour or chemical transports. In the oceans the scarcity of historical observational data coupled with a plethora of research mode numerical models has meant that there are many more ocean circulation products available. A recent summary of the state of the art in developing ocean synthesis products can be found in Stammer et al. (2010). The vast majority of global synthesis products have been based on using lower resolution numerical models, often to permit the application of more sophisticated data assimilation approaches. The Simple Ocean Data

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Assimilation (SODA) system of Carton and Geise (2008) and Zheng and Geise (2009) is the most similar in design to the work being presented here.

In this paper we report on a global $1/4^\circ$ by 46 level ocean reanalysis product developed as part of the EU GMES Marine Core Services program MyOcean. The entire period of the reanalysis covers 1989–2010 but we report primarily on the period 1993–2009 as the MyOcean program is developing an intercomparison of ocean products for this period. The standard evaluation of this reanalysis product against the observational data being assimilated or against independent but related products to the observations, are available through the MyOcean product reports, Valdivieso et al. (2011a, b). In this paper we focus on using the reanalysis product to assess transports for heat and freshwater, both in terms of exchanges at the ocean surface with the atmosphere, and horizontal transports within the oceans. We develop closed budgets for heat and freshwater exchanges both globally and within ocean basins, where the role of ocean data assimilation is explicitly resolved and interpreted.

In Sect. 2 we describe the model, the surface forcing, and the data assimilation process. In Sect. 3 we look at the global and regional heat budgets within the reanalysis product. In Sect. 4 we do the same for the freshwater budgets. In Sect. 5 we make a particular focus on the meridional overturning circulation (MOC) in the Atlantic ocean and the meridional heat and freshwater transports in that basin. Section 6 is a summary and conclusions.

2 Model and data assimilation description

2.1 Ocean model

The numerical model used is the NEMO coupled ice-ocean model (Madec, 2008) version 2.3, based on the OPA9 ocean model (Madec et al., 1998) and the LIM2.0 sea ice model (Louvain sea Ice Model: Fichefet and Maqueda, 1997; Goosse and Fichefet, 1999). It is a primitive equation z-level model with hydrostatic and Boussinesq

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approximations, and a free surface (Roulet and Madec, 2000) with partial cell topography (Adcroft et al., 1997). It has the tri-polar “ORCA” 1/4° grid and 46 levels with thicknesses from 6–250 m, as developed through the DRAKKAR Consortium (Barnier et al., 2007) with parameter settings as in (Barnier et al., 2006) and (Penduff et al., 2010).

2.2 Surface forcing

Surface atmospheric forcing for the period of 1989–2009 is obtained from ECMWF ERAInterim 6 h reanalysis (Simmons et al., 2007; Dee and Uppala, 2009), which provides 10 m wind, 2 m air humidity and temperature for the bulk fluxes calculated from Large and Yeager (2004). Downwelling short and long wave radiative fluxes and precipitation over the oceans are unmodified from ERAInterim. Monthly climatological runoff (Dai and Trenberth, 2002) is applied along the land mask. This reanalysis also used a surface layer salinity relaxation/damping to climatological values, with a timescale of 180 days (36 days under ice).

2.3 Initial conditions and data assimilation

Initial conditions for temperature, salinity and circulation are derived from the end of a previous 1/4° reanalysis run from the same model in December 2004 (run UR025.1 described in Smith et al., 2009) which was forced with DFS3 meteorology, Brodeau et al. (2009). Most importantly the initial conditions will have seen the start of the Argo profile assimilation and therefore will have more complete and consistent water properties from the beginning.

The data assimilation is restricted to temperature, T , and salinity, S , profile data, referred to as $S(T)$ (Haines et al., 2006; Smith and Haines, 2009), and is similar to what is currently employed in the ECMWF System 3 reanalysis (Balmaseda et al., 2008). It is a sequential scheme with Optimal Interpolation. Temperature profiles are assimilated along with a salinity balancing increment as advocated by Troccoli and Haines (1999),

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and salinity profiles are then assimilated along isotherms (i.e. $S(T)$), up to latitudes of around 50° N/S. At higher latitudes salinity is assimilated in a univariate way on z levels. Assimilation increments are calculated using a First Guess at Appropriate Time (FGAT) every 5 days (73 cycles per year), and introduced evenly over the following day (known as Incremental Analysis Updating, see Bloom et al., 1996). Observations are from the UK MetOffice quality controlled ENACT/ENSEMBLES (EN3.v2a) dataset (Ingleby and Huddleston, 2007) and they have not had any XBT bias corrections applied. This reanalysis run is referred to as UR025.3 and there is also a control run initialised and forced in the same way but without applying the data assimilation.

3 Heat fluxes, transports and budgets

The data assimilation method used is non-conservative and so heat, salt and freshwater content are changed by the assimilation increments and must be accounted for in any budget study, e.g. Fox and Haines (2003), Zuo et al. (2011).

Figure 1a shows the mean sea surface temperature, SST, for the period 1993–2009, and Fig. 1b shows the differences from HadISST over the same period. Figure 1c shows the equivalent control run difference with HadISST. It can be seen that errors in the reanalysis are quite small everywhere with much smaller biases throughout the tropics and in western boundary current regions than in the control run, or from previous runs of this model which used different meteorological forcings, e.g. Brodeau et al. (2009), Fig. 12. This demonstrates that the data assimilation, although not directly assimilating SST, is sufficient to hold the surface properties much closer to observations. A full report on this reanalysis run can be found in Valdivieso and Haines (2011). These SST corrections have a significant influence on air-sea fluxes as will be seen later.

Figure 2 shows the time evolving annual terms from the global ocean heat budget over the 17 yr period. It can be seen that the dominant balance is between surface heat fluxes warming the oceans at a mean rate of $Q_{\text{net}} \sim 5.4 \text{ Wm}^{-2}$, balanced by data

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assimilation cooling the oceans at a mean rate of $Q_{\text{assim}} \sim -4.9 \text{ Wm}^{-2}$. The imbalance represents an average warming rate of 0.5 Wm^{-2} , which is captured accurately in the rate of increase in heat storage, Q_t . This rate of increase in heat storage is quite consistent with estimates of the global radiation budget from the top of atmosphere measurements, e.g. Wong et al. (2009), showing that the total ocean state, including the deep ocean, is well aligned with observational data and is not drifting unrealistically.

The component of the excess surface heat flux balancing the data assimilation i.e. 4.9 Wm^{-2} , can be regarded as predominantly an estimate of the global excess in downwelling shortwave radiation in the ERA-Interim product due to missing cloud effects, e.g. cloud opacity and regional biases such as in stratocumulus, Kallberg (2011), Dee et al. (2011). The assimilation compensates for this effect by removing excess heat. The directly estimated surface heat fluxes from the ERAInterim atmospheric reanalysis, based on the specified surface SST, also have a net global downwelling radiation of $\sim 7 \text{ Wm}^{-2}$, Kallberg (2011). In the equivalent ocean control run without data assimilation, the model compensates by developing globally too high SSTs (Fig. 1c), allowing additional surface heat losses through sensible, and especially latent heat fluxes, to compensate for the additional downwelling shortwave. The largest temporal anomaly in Fig. 2 is the 1998 ENSO event when the ocean gain in heat storage peaks, followed by two years of net ocean heat loss. Assimilation clearly plays a significant role in this ENSO heat content cycle, although the net surface heat flux also shows a strong signal.

Table 1 gives the mean heat (and freshwater, FW) budgets globally and for each ocean basin. In most basins the residuals are small compared to the 17 yr storage changes and even in the Arctic and southern oceans the budget residuals are much smaller than interannual storage signals. The Atlantic is gaining freshwater from the Arctic in the North and from the southern ocean in the south, and is thus a net evaporation basin, although this is only weakly indicated from the surface fluxes, with assimilation removing most of the FW. The Atlantic trend is to get warmer $\sim 2 \text{ Wm}^{-2}$ and saltier (loosing 9.6 cm yr^{-1} of FW). The Pacific gets warmer, also 0.5 Wm^{-2} , but fresher

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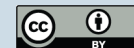
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(gaining 1.8 cm yr^{-1} of FW) with the freshening attributable to the resolved net precipitation because the assimilation terms would make the basin saltier. The Indian ocean also warms (0.75 W m^{-2}) but with a very small trend in freshwater. The basin is seen to be a net evaporation basin with FW advected in and lost at the surface, again with assimilation playing a smaller role.

To investigate regional heat balances we need to look at ocean heat transports. If we consider the 17 yr heat budget as a function of latitude, Fig. 3a shows the global meridional ocean heat transports integrated southward from 90° N . The red line shows the actual heat transport calculated from model velocity fields with the pink shading around it representing the interannual standard deviation in transport at each latitude. The solid blue line is the meridional heat transport based on the surface heat fluxes alone, assuming zero net storage of heat at all latitudes, and the dashed blue line shows the same calculation with surface heat fluxes and data assimilation increments added together. Clearly assimilation increments are needed to make the surface flux calculation consistent with the direct meridional transports (red curve), with the small discrepancies at the southern end resulting from the non-zero storage because the ocean is warming. Heat transports across a number of latitudes, based on the ocean section inverse models of Lumpkin and Speer (2007) and Ganachaud and Wunsch (2003), are also shown, with their estimated error bars. The global transport estimates are generally close to the ocean inverse estimates, except in the subtropics 24° N , which we see below, is entirely due to the Atlantic.

The same transport calculations for the Atlantic basin alone are shown in Fig. 3b. Although the peak northward transport in the Atlantic reaches 1.15 PW at 36° N , the transport is considerably reduced through most of the subtropics and generally below the inverse calculations from 20° S – 30° N , although the error bars overlap with the interannual standard deviations from our model transports. The main reason for this is due to the lack of a bias correction for the assimilation increments at the Equator where the sharp equatorial thermocline cannot be maintained properly and assimilation increments then induce large secondary circulations, e.g. Bell et al. (2004). In a later

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reanalysis product this effect has been greatly reduced. At higher latitudes in the Atlantic however, the transports in this reanalysis agree very well with inverse estimates, e.g. at 30° S or 47° N. In Sect. 5 we will look in more detail at the transports at 26° N in the Atlantic where the RAPID transport array, Cunningham et al. (2007), now provides independent evaluation measurements.

Figure 4 shows the mean southern ocean heat transports across several sections. The model transports are generally in very good agreement for all the sections, especially over the last period constrained well by Argo 2005–2009. Again the interannual standard deviation in transports is shown, along with the transport trends. There are no statistically significant trends seen in any of these transports. A further paper, Valdivieso et al. (2012), will look in more detail at the time variability of these transports and seek to assess whether such variability could be linked to the variability in ocean storage or whether there are associated changes in air-sea fluxes, which could have a feedback on atmospheric climate.

While on the global scale the dominant balance in the heat budget is between surface fluxes adding heat and data assimilation removing heat, at the local scale, particularly near strong currents, the balance is generally three way between surface fluxes, data assimilation and the mean heat transport convergence, with the last two usually showing larger spatial variability, as previously found in other data assimilation experiments, e.g. Haines (2003), Fig. 6. Here in Fig. 5 we show for example details of these three terms in the heat budget over the subpolar gyre of the N Atlantic, along with an independent estimate of the surface fluxes from OAFlux (Yu et al., 2008). The resolved surface heat fluxes reproduce the spatial distribution of the OAFlux product rather well with a very similar position for the reversal from warming to cooling across the Atlantic basin. The dipole in transport convergence along the Gulf Stream path, suggesting a tendency for the Gulf stream to drift north, is balanced by data assimilation tending to move the current, and heat content, back to a more correct southerly position. Around the subpolar gyre, transport convergence of heat around the boundary and along steep topography, e.g. the edge of the Labrador Sea and also the Reykjanes

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ridge, is removed by the data assimilation increments. The assimilation at these scales appears to be compensating for the lack of eddy heat transport because it is taking heat away from the boundary currents and depositing it in the basin interior, which is exactly the role that baroclinic eddies would be expected to play. Despite the relatively high model resolution, eddy transports are clearly insufficient at these latitudes in the subpolar gyre.

4 Freshwater fluxes, transports and budgets

We have also assessed the mean model freshwater fluxes and transports against independent observational estimates. The surface fluxes of freshwater are complicated by the existence of the surface salinity relaxation term which has often been used in model simulations because the surface salinity is not closely coupled to atmospheric conditions, and so is free to drift. In a new reanalysis for MyOcean v2, we have removed this term altogether, instead relying on data assimilation to counter any surface salinity drift tendency.

Figure 6 shows the time evolving annual terms from the global freshwater budget over the 17 yr period. As in the global heat budget, Fig. 2, the dominant balance in the freshwater budget is between a surface gain in freshwater, F_{net} , at a mean rate of 4.0 cm yr^{-1} , balanced by assimilation loss of freshwater, F_{assim} , at $\sim 5.5 \text{ cm yr}^{-1}$. The interannual variability of the assimilation term is very large, however, particularly prior to and during the period when the Argo array gets up and running between 1999–2004, as much of the ocean starts to have salinity observations down to 2000 m for the first time. After 2004 the freshwater budgets shows signs of being much more in balance again from year to year as the Argo array becomes established.

To interpret the role of the data assimilation in more detail we first note that the sea surface salinity damping contributes a surface loss of freshwater at a mean rate of $\sim 2 \text{ cm yr}^{-1}$. The damping, along with the salinity data assimilation, use virtual salt fluxes rather than freshwater fluxes and so the actual imbalance to the addition of water

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leads to a net sea level rise in the model at a rate of $\sim 6 \text{ cm yr}^{-1}$, but the total effect on the storage means the ocean gets saltier over the period.

If we regard the data assimilation and surface salinity damping terms together, minus the change in storage, as giving an estimate of the excess precipitation in the ERA-Interim product (assuming that evaporation is now calculated correctly from the bulk formulae based on the assimilation corrected SST fields in Fig. 1a), that is equivalent to a precipitation excess of $\sim 6 \text{ cm yr}^{-1}$ over the oceans. Kallberg (2011) and Dee et al. (2011) also comment on the excess in net precipitation in the ERA-Interim products, which vary with time. Alternatively the additional latent heat required to evaporate $\sim 6 \text{ cm yr}^{-1}$ of excess precipitation turns out to be $\sim 4.7 \text{ Wm}^{-2}$, which is very close to the excess surface heating removed by data assimilation in the global heat budget. Indeed in the control run without data assimilation, the model achieves both a balanced heat and freshwater budget just through having the higher SSTs in Fig. 1c, which lead to higher surface heat loss and freshwater loss through evaporation of an extra $6\text{--}7 \text{ cm yr}^{-1}$ of water from the ocean surface.

We can see how the freshwater budget is balanced as a function of latitude in Fig. 7a which shows zonal averaged results for the global ocean, and Fig. 7b for the Atlantic only. The red line shows the freshwater (FW) transport across each latitude circle south from 90° N in Fig. 7a, and from 66° N in Fig. 7b (the latitude of the Bering strait), calculated directly from the model velocity fields and using a salinity reference value equivalent to each section average salinity. The pink shaded area around this line shows the standard deviation of annual variability in transports about the 17 yr mean. The blue line shows the FW transports calculated from the surface fluxes (E-P-R) alone, integrated southward while assuming no FW storage, as in Wijffels (2001). The blue dashed line is the same as the blue line, but including the surface salinity damping and the data assimilation terms, which clearly bring the results into close agreement with the direct FW transport assessments. The values in the Atlantic at 66° N are made to agree to avoid some of the problems associated with Bering Strait transports and Arctic ice fluxes which occur further north.

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Meridional ocean freshwater transports as in Fig. 7a, b, can be evaluated in two ways. Against individual ocean hydrographic sections, or against inverse models, using specific latitudes, some of which are marked on Figs. 7a, b. Also the ocean FW transports can be matched against atmospheric FW transports calculated from atmospheric reanalysis products, or against independent surface FW flux estimates, many results of which can be found in Wijffels (2001), Figs. 6.2.6–6.2.9. The estimates from this ocean reanalysis are broadly consistent with these previous estimates, and the interannual standard deviation gives some idea of the error incurred by inverse models when using single section data taken in one particular year to estimate transports.

5 Role of the Atlantic MOC

The meridional overturning circulation in the Atlantic plays a key role in controlling both meridional heat and freshwater transports and influencing N Atlantic SSTs, (e.g. Knight et al., 2005; Delworth et al., 2007; Robson et al., 2012). At 26° N the timeseries of the meridional heat transport is now continuously monitored by the RAPID array program, Cunningham et al. (2007). Figure 8 shows the monthly and annual timeseries of Atlantic MOC at 26.5° N from our reanalysis from 1989–2010, along with the MOC derived from the RAPID array from 2004 onwards. Since much of the variability is the direct result of wind stress. The correspondence with the observational variability in the Rapid array period in both figures is clear. The mean MOC for the long period 1993–2009 is 17.0 Sv and during the RAPID array period 2004–2009 it is 17.3 Sv compared with the RAPID observational value of 17.5 Sv for this period. It is interesting to note that the annual mean values in 1992 and 1998, indicating a decline over the period, compare very well to the Bryden et al. (2005) section estimates also shown in the figure. Between 1990 and 2000 the MOC in fact shows a low frequency maximum around 1995, which is now also consistent with other reanalysis estimates e.g. Pohlmann et al. (unpublished data, 2011), which is probably driven by variations in the NAO and associated buoyancy forcing, e.g. Robson et al. (2012).

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the total meridional FW transport in the Atlantic shown in Fig. 7b into the component transported by the vertical overturning circulation, also called the Mov, e.g. Drijfhout et al. (2011). The vertical overturning circulation is calculated from the zonal averaged baroclinic currents and the zonal average salinity. The gyre transport is given by the difference between the two curves shown. It can be seen that at around 30° S the Mov component of Freshwater transport is negative, i.e. southwards out of the Atlantic, despite the fact that the total ocean freshwater transport is northward into the Atlantic basin. This southward transport of the Mov has been suggested as an indicator of instability in the Atlantic overturning circulation, e.g. Dijkstra (2007), Hawkins et al. (2011). The reanalysis confirms the negative sign of Mov found in other observationally based analyses at 30° S, and also shows that the Mov appears to be negative throughout the Atlantic basin, but with large changes in the Northern Hemisphere reflecting large and rapidly changing contributions to freshwater transport through the subtropical and subpolar gyres. The argument for the Mov at 30° S determining the stability of the Atlantic overturning assumes that other components of the meridional freshwater transport between 30° S and the deep water formation latitudes in the N Atlantic are not temporally correlated with the Mov component of transport. This aspect of the Atlantic freshwater transports will be investigated within this reanalysis in greater detail in a future study.

6 Summary and conclusions

In this paper we have looked at the reproduction of heat and freshwater budgets and transports in a 1/4° global ocean reanalysis product prepared as part of the EU GMES Marine Core Services program. The intention has been to show that the budgets of heat and freshwater can be consistently explained despite the presence of “unphysical” data assimilation terms. Second, the data assimilation terms can be interpreted as making corrections to the resolved physical processes acting in the model. And third that the horizontal transports in the model are mostly very well reproduced compared

with independent observation based estimates, and these transports do not show unphysical trends that are common in ocean model simulation runs.

On the global scale the data assimilation increments are shown to counteract global imbalances in surface forcing, in particular correcting for higher incidence of solar short-wave radiation (Kallberg, 2011; Dee et al., 2011), by removing heat, and also correcting for a higher than average global precipitation in ERAInterim, (Kallberg, 2011), by adding salt. For the zonally integrated meridional transport budgets the inclusion of the data assimilation terms effectively closes the budgets at all latitudes showing that the assimilating model only exhibits small trends that are consistent with real upper ocean changes. Furthermore, the modelled transports assessed against trans-oceanic section based estimates are very realistic, e.g. throughout the southern oceans and within the Pacific and Indian basins. Zuo et al. (2011) have previously shown that the transports into/out of Arctic latitudes in this reanalysis are also reasonably consistent with observations. There are some discrepancies in meridional heat transport across the Equator in the Atlantic and up as far as the subtropical gyre. The transports at 26° N are compared in detail with the RAPID array estimates of Johns et al. (2011) and it is shown that discrepancies mainly lie in the Midocean component of the heat transport, although the Ekman transport is also slightly deficient based on using weaker surface winds in ERAInterim compared with QuickScat.

The meridional freshwater transports in the model compare reasonably well with previous estimates although there is a much higher degree of uncertainty because of the greater scarcity of accurate salinity measurements in the ocean prior to Argo, and due to the independent difficulties of estimating precipitation over the oceans. The meridional transport of freshwater by the overturning circulation has been estimated at all latitudes in the Atlantic and shown to be highly variable, although the southward freshwater transport around 30° S in the Atlantic appears to be a robust result throughout the 17 yr reanalysis period.

A new reanalysis is currently underway using a more comprehensive data assimilation system, including altimeter, sea surface temperature and sea ice concentration

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increments, in addition to the ocean profile assimilation used here, based on the full FOAM data assimilation system currently in operational oceanography use at the UK Met Office. Other reanalyses from the EU MyOcean program will also be available in a uniform format that will allow similar diagnostic comparisons. In a future study we will use a similar suite of diagnostics to assess the robustness of these results between models and assimilation systems. It is only through such robust comparisons can we begin to learn to interpret data assimilation results more quantitatively and begin an effective investigation of model error processes.

Acknowledgements. This work was supported by grants from NERC through the National Centre for Earth Observation, the Rapid-Watch Valor project, and from the EU GMES MyOcean project.

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Table 1. Vertically integrated heat and freshwater budget averaged over the period 1993–2009 for the global ocean and for individual oceanic basins. Upper numbers in each cell indicate heat fluxes in Wm^{-2} ; lower numbers indicate freshwater fluxes in cm yr^{-1} and errors represent annual standard deviations. The residual is computed as the sum of advection, surface forcing and assimilation minus the contribution from the storage. Positive values indicate fluxes into the ocean.

Basin Areas m^2	Advection	Surface Forcing	Assimilation	Storage	Residual
Global (3.61×10^{14})	0.00±0.00 −0.01±0.00	5.39±1.23 4.00±1.99	−4.90±3.01 −5.46±9.61	0.48±1.62 −1.62± 8.81	0.01±0.89 0.15±1.91
Atlantic (6.87×10^{13})	3.06±1.01 22.20±4.03	7.53± 1.96 −10.49±2.40	−9.34±7.14 −23.79±36.92	2.18 ±6.46 −9.65±35.97	−0.93±1.87 −2.42±4.39
Pacific (1.35×10^{14})	−7.28±0.97 0.88±1.12	14.36± 2.50 3.38±3.91	−6.75±3.80 −3.05±8.39	0.50±3.66 1.79±8.30	−0.17±1.07 −0.58±1.64
Indian (4.31×10^{13})	−5.01±2.55 30.15±4.27	10.83±2.82 −20.48±3.67	−5.22±4.28 −11.86±17.24	0.75±5.92 −0.58±16.83	−0.16 ±1.30 −1.61±3.06
South. Ocean (1.02×10^{14})	7.71±1.23 −25.30±2.69	−7.03±1.31 18.34±1.85	−0.50±6.90 7.73±14.40	−0.88±5.92 −1.74±13.04	1.06±2.25 2.50±6.24
Arctic (1.23×10^{13})	16.40±1.60 −29.37±7.37	−21.05±1.91 57.30±12.12	4.72±4.23 −16.16±31.48	1.00±3.51 2.87±28.25	−0.94±0.80 8.90±7.37

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Table 2. Decomposition of the Atlantic 26° N meridional heat transport in UR025.3 over the 3.5 yr period April 2004–October 2007, compared with Johns et al. (2011) estimates for the same period, along with the meridional overturning components. Corresponding standard deviations are given in round brackets. For the full 2004–2010 period the RAPID MOC is 18.6 (3.7) Sv and UR025.3 MOC is 17.3 (3.3) Sv.

Meridional heat transport component	RAPID mean(std), PW	UR025.3 mean(std), PW	Transport in RAPID mean (std) Sv (MOC=16.6 (2.6))	Transport in UR025.3 mean (std) Sv (MOC=17.6 (3.5))
Florida Current	2.53 (0.24)	2.55 (0.23)	31.7 (2.8)	32.6 (2.4)
Ekman	0.35 (0.34)	0.28 (0.18)	3.5 (2.3)	3.2 (1.7)
Midocean	−1.77 (0.25)	−1.82 (0.23)	−36.5 (4.4)	
WB-Abaco	0.13 (0.16)	–	1.3 (–)	−37.4 (2.9)
Eddy	0.11 (0.04)	0.08 (0.03)	–	
Total	1.35 (0.40)	1.09 (0.27)	0.0	−1.6

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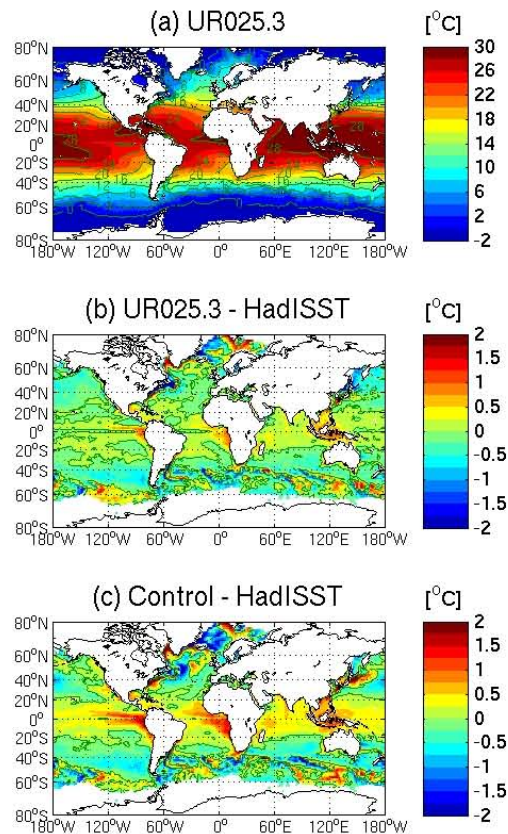


Fig. 1. (a) The mean sea surface temperature (SST) from the UR025.3 reanalysis for 1993–2009, (b) the difference between the mean SST and that from HadISST for the same period, (c) the difference between the equivalent model control run and HadISST through the same period.

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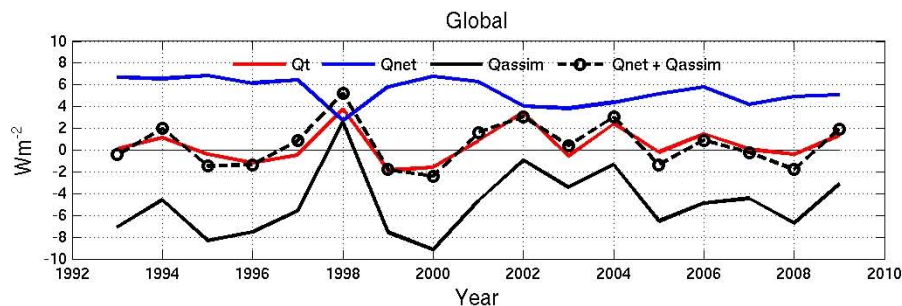


Fig. 2. Annual timeseries of the global ocean heat budget from 1993–2009. The terms are Q_{net} the surface heat flux, Q_{assim} the equivalent heat flux coming from data assimilation increments vertically integrated, and Q_t the heat storage associated with changes in global mean temperature.

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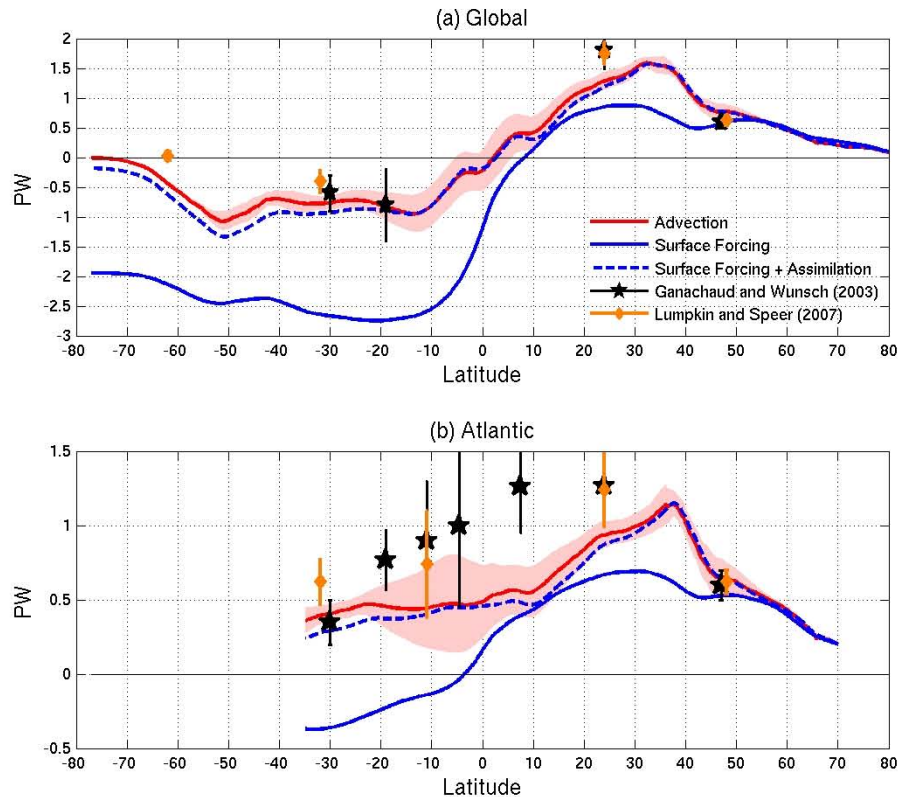


Fig. 3. Meridional heat transports from UR025.3 over 1993–2009 for **(a)** the global ocean and **(b)** the Atlantic basin. The Advection (red) is based on the resolved velocity and temperatures fields, including all eddy components down to the 5 day timescale. The shaded area shows the annual standard deviations in advective transport. The solid blue line uses surface heat fluxes and a steady state assumption to calculate meridional heat transport, and the dashed blue line uses surface heat fluxes combined with assimilation increments to do the same.

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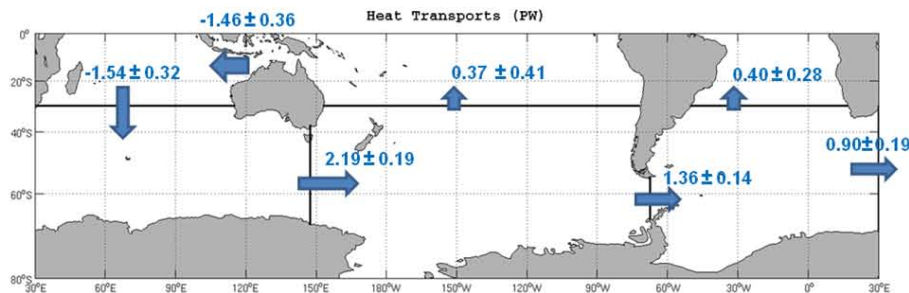
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Section	UR025.3	UR025.3	LS07 Inverse
	1993–2009	2005–2009	Analysis
	Mean + STD	Mean + STD	Mean + STD
Drake Passage	1.35 ± 0.13	1.36 ± 0.14	1.36 ± 0.09
South A. – Antarctica 30° E	0.95 ± 0.20	0.90 ± 0.19	0.88 ± 0.24
Australia – Antarctica 143° E	2.20 ± 0.19	2.19 ± 0.19	1.85 ± 0.18
Indonesian Throughflow	-1.35 ± 0.37	-1.46 ± 0.36	-1.46 ± 0.21
Indian 32° S	-1.50 ± 0.29	-1.54 ± 0.32	-1.55 ± 0.12
Pacific 32° S	0.34 ± 0.44	0.37 ± 0.41	0.55 ± 0.15
Atlantic 32° S	0.40 ± 0.22	0.40 ± 0.28	0.62 ± 0.15

Fig. 4. Heat transports and their annual standard deviations, in PW, through major sections in the southern hemisphere in UR025.3 over the periods 1993–2009 and through 2005–2009 in comparison to estimates from the inverse model results of Lumpkin and Speer (2007).

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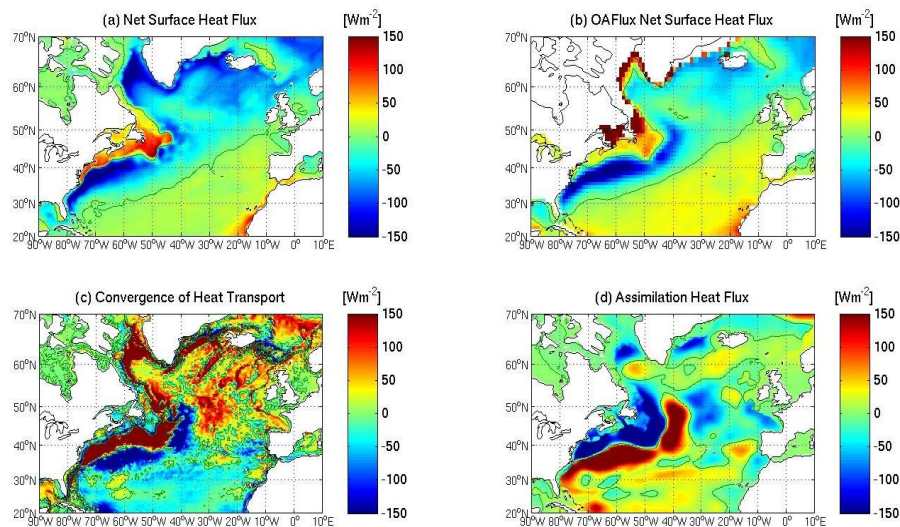


Fig. 5. Main components of the heat budget over the N Atlantic in UR025.3 for 1993–2009. **(a)** Net surface heat flux, **(c)** heat transport convergence, **(d)** assimilation increments equivalent heat flux. **(b)** Provides a comparison from an independent surface heat flux product OAFlex, from Wong et al. (2009).

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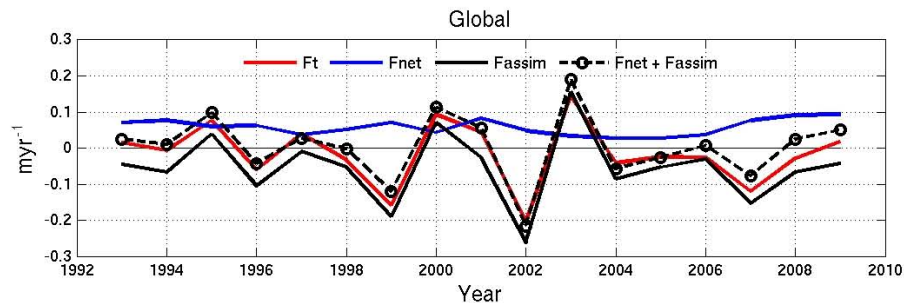


Fig. 6. Annual timeseries of the global ocean freshwater budget from 1993–2009. The terms are F_{net} the surface freshwater flux, F_{assim} the equivalent freshwater flux coming from data assimilation increments vertically integrated, and F_t the freshwater storage associated with changes in global mean salinity.

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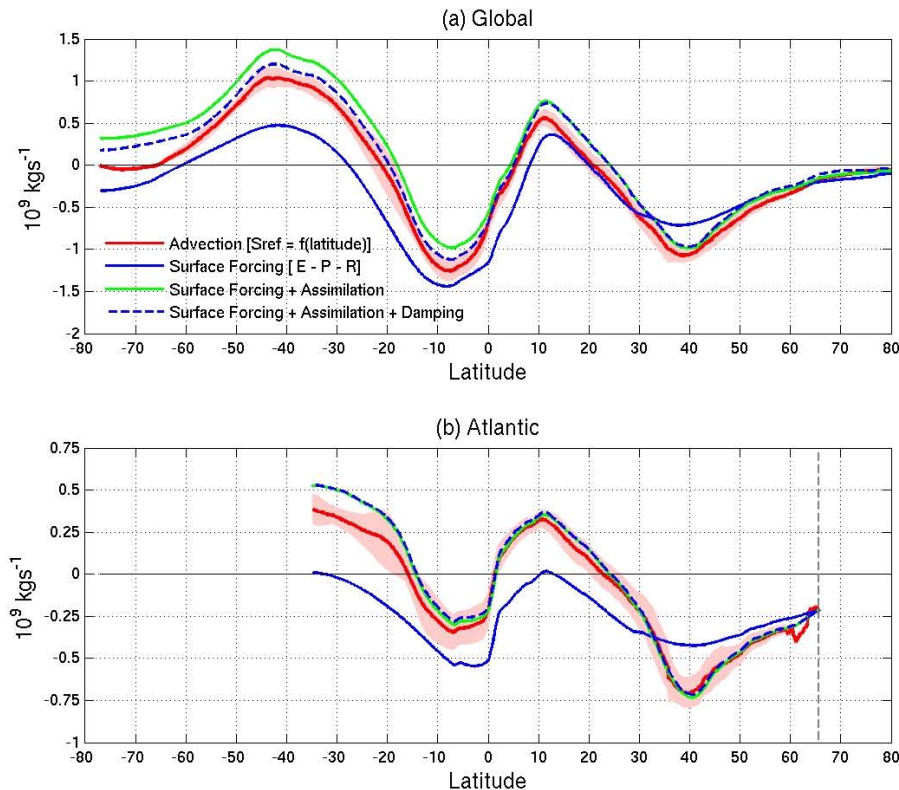


Fig. 7. Meridional freshwater transports from UR025.3 over 1993–2009 for **(a)** the global ocean and **(b)** the Atlantic basin. The Advection (red) is based on the resolved velocity and salinity fields, including all eddy components down to the 5 day timescale. The shaded area shows the annual standard deviations in advective transport. The solid blue line uses surface freshwater fluxes and a steady state assumption to calculate meridional freshwater transport, and the dashed blue line uses surface freshwater fluxes combined with assimilation increments to do the same.

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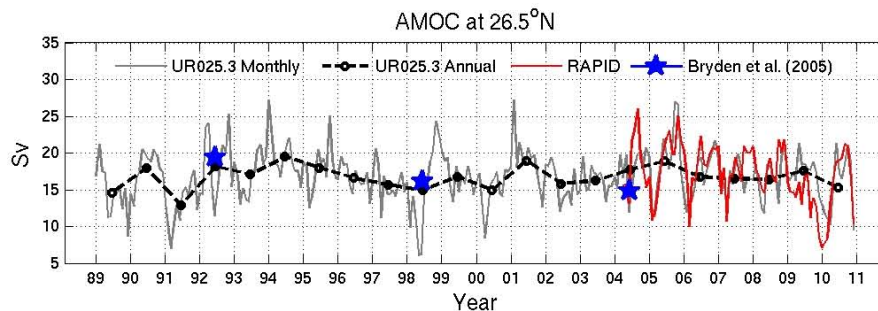


Fig. 8. Monthly and annual mean timeseries of the Atlantic meridional overturning circulation (AMOC), from UR025.3 in comparison with the RAPID array values from 2004–2010. Also shown at the annual mean section based estimates in 1992, 1998 and 2004, from Bryden et al. (2005).

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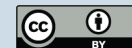
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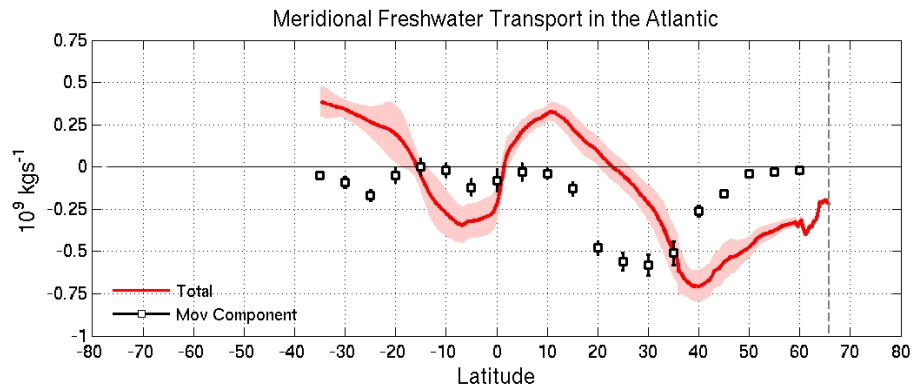


Fig. 9. Monthly and annual mean timeseries of the meridional freshwater transport in the Atlantic at 30°S from UR025.3. The total meridional transport (continuous line) and the “MOV” component due to the vertical overturning alone (every 5 degrees), are both shown.

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