The ECCO Report Series

Global Sea Surface Flux Estimates Obtained Through Ocean Data Assimilation

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Report Number 13


1The ECCO Project is a consortium of the Jet Propulsion Laboratory, The Massachusetts Institute of Technology and the Scripps Institution of Oceanography and is funded through a grant from the National Oceanographic Partnership Program (NOPP). Copies of this Report are available at www.ecco-group.org or from Detlef Stammer, Scripps Institution of Oceanography, La Jolla CA 92039-0290, ph.: (858) 534-4464, fax: (858) 534-4464; e-mail: dstammer@ucsd.edu.

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Abstract

Oceanic state estimation has advanced sufficiently that it is now possible to make estimates of the atmospheric forcing fields required to reproduce global ocean observations. Here we use the data assimilation results from nine years of observations during the World Ocean Circulation Experiment to assess the adjustments of surface flux fields of momentum, heat and freshwater, necessary to make the a priori ocean-forcing fields from the National Center for Environmental Prediction (NCEP) consistent with the observed ocean. Independent estimates of the adjustments from bulk formula and regional field observations are also employed to evaluate the results. Adjustments required of the NCEP freshwater and heat exchanges are all within the crude prior error bars on these fields, and on the large scale are consistent with known deficiencies in the NCEP products. Windstress adjustments are also everywhere within the prior error bars, but exhibit regional features that are due to ocean model failures to resolve intense boundary currents. On the largest scales, the inferred adjustments to NCEP winds are consistent with inferences made from satellite wind measurements. Rapid improvements in all of these procedures and results are anticipated.
1 Introduction

Ocean state estimation procedures are powerful tools for obtaining a synthesis of global data sets with a general circulation model so that the result is a dynamically consistent picture of the time-evolving circulation. These procedures have much in common with ongoing analysis and re-analysis activities in the atmospheric community. But there are substantial differences from the meteorological practice, which is driven by forecasting requirements. The oceanographic focus is primarily on estimating the oceanic state over the last 10-50 years in a dynamically consistent way so that the results can be used to diagnose climate variability in the ocean, its causes and effects. Recent accounts of ocean state estimation procedures as used here are provided by Wunsch (1996), Malanotte-Rizzoli (1996), and Fukumori (2000).

Combining a circulation model with ocean data requires control parameters that are adjusted during the estimation procedure (see below) so as to best fit (in a least-squares sense) the ocean observations, while at the same time being faithful to the physical principles implemented in the model code. Typically, these controls represent uncertainties in knowledge, e.g., of ocean mixing parameters or surface forcing fields. In particular, despite major advances in recent years, the ocean and climate communities still contend with substantial uncertainties in air-sea flux products, and the lack of formal error covariances for them. For example, Griffies et al. (2001) discuss the urgent need for improved flux fields to identify problems in ocean general circulation models. Because of their large uncertainty and the absence of any formal error quantification, the surface flux fields are important control parameters in the ocean state estimation procedure.

The present state of surface flux estimation is comprehensively reviewed by the Joint WCRP/SCOR Working Group on Air-Sea Fluxes (WGASF, 2000). Weller and Taylor (1999) describe recent intense observing campaigns that were intended to improve knowledge of surface flux fields. Another important activity has been the re-analysis efforts by operational Numerical Weather Prediction (NWP) centers such as the European Centre for Medium Range Weather Forecasts (ECMWF; Gibson et al., 1997) and the National Center for Environmental Forecasting (NCEP; Kalnay et al., 1996). Reanalysis products
include 6-hourly global air-sea flux fields on a regular grid, making them ideal as forcing fields in ocean models. However, all existing atmospheric re-analysis flux products have deficiencies (e.g. Milliff et al., 2000, Smith et al., 2001; Wang and McPhaden, 2001). Problems exist with the tropical wind stress, wind divergences, lower layer humidity, bulk flux algorithms, precipitation, clouds associated with a double ITCZ in the South Pacific, and the sea-surface albedo, among others.

The purpose of this note is to describe the adjusted surface flux fields as they emerge from the ocean state estimation procedures. These estimates are largely independent of other methods of determining the air-sea fluxes and can be thought of as representing the best-estimates of the air-sea fluxes required to make the ocean model consistent with the observations. They thus are built upon a foundation of the skill in the ocean models plus the information content of the ocean observations.

Stammer et al. (2002a) describe the first extensive attempt by a consortium, Estimating the Circulation and Climate of the Ocean (ECCO), to produce a full, nearly rigorous, state estimate of the ocean, using much of the data from the World Ocean Circulation Experiment (WOCE) and the NCEP re-analysis. The consortium consists of scientists at the Massachusetts Institute of Technology, Jet Propulsion Laboratory and Scripps Institution of Oceanography. A full description of the program is provided by ECCO Consortium (2002; see also Report Number 1 at http://ecco-group.org). Control variables in the calculation include the initial temperature and salinity distributions as well as daily surface net heat and freshwater and momentum fluxes. Overall, the constrained model displays considerable skill in reproducing the observations, and in particular, many of the qualitative and quantitative features of withheld data.

A focus of a subsequent analysis of those results was to use the estimates of the ocean state to compute transports of volume, heat and freshwater and their divergences that appear to be most important to understanding the climate system (Stammer et al., 2002b). The authors discussed in detail globally and basin-integrated meridional oceanic heat and freshwater transports evaluated at various nominal WOCE section position and compare them with horizontal transports as they would correspond to the estimated surface fluxes of heat and freshwater that are being discussed here. The transports do not agree, because
of an imbalance between the surface forcing and the model state. Over short integration times, such imbalances are expected and can be physically real, or be reflecting model adjustment from its initial state. One must therefore, be careful about inferring surface heat fluxes from divergences of meridional transports (and vice-versa).

ECCO meridional fluxes are mostly consistent with those estimates provided by Ganachaud and Wunsch (2000) but significantly underestimate northward transports in the North Atlantic - at 25°N by about 0.5PW. Given the problems the low-resolution model has in simulating boundary current processes it probably is not meaningful to discuss basin-integrated fluxes and respective horizontal flux divergences. Our emphasis here is therefore, somewhat different and will be on the air-sea flux (the ocean flux divergence) of heat, freshwater and wind stress over the interior ocean where previous bulk formula-based estimates revealed uncertainties of +−30W/m² or more (Josey et al., 1999). Our goal is to provide a quantitative description of the quality of the ECCO estimates by comparing them with independent flux estimates. Large and Nurser (2001, denoted hereafter as LN01) describe deficiencies in the NCEP re-analysis by using observationally based flux algorithms, satellite based radiation, a more realistic surface albedo, a reduced near surface humidity, and a blend of satellite, in situ and model derived precipitation. The comparison will reveal information about uncertainties, residing in the ECCO ocean model or the NCEP atmospheric model. In cases when the estimated adjustments to NCEP surface forcing fields are consistent with independent assessments of the prior NCEP flux products, such as differences between the LN01 fluxes and the NCEP re-analysis fluxes, the estimated adjustments are more likely to be improvements of the NCEP fields (clearly not a proof, however). In some regions, the results are ambiguous, and these may be the focus of future efforts. Because uncertainties in all available flux products increase significantly over the Southern Ocean, this region is omitted altogether.

2 Methodology

We use the MIT general ocean circulation model (Marshall et al., 1997a,b) and its adjoint, which is based on the primitive equations on a sphere under the Boussinesq approxima-
tion. The model consists of prognostic equations for horizontal velocity, heat, and salt that are integrated forward in time on a staggered “C”-grid (Arakawa and Lamb, 1977). A full surface mixed layer model is used (called KPP, Large et al., 1994) and convective adjustment is used to remove occasional gravitational instabilities underneath the planetary boundary layer.

The adjoint code was obtained from the forward code through automatic differentiation (AD; Giering and Kaminski, 1998; Marotzke et al., 1999). In practice, this (semi) automatic system has proven to be extremely flexible, permitting relatively easy regeneration of the adjoint code whenever a change is made to either the numerics of the forward code or the objective function (the latter is formally part of the forward model).

The model was configured globally with 2° horizontal resolution over ± 80° latitude with 22 levels in the vertical. (see Stammer et al., 2002a for details). It was run with free-slip bottom boundary conditions and non-slip boundary conditions at lateral walls. Laplacian viscosity and diffusivities are used, with \(\nu_h = 10^3 \text{ m}^2/\text{s}\) and \(\kappa_h = 10^3 \text{ m}^2/\text{s}\) and \(\nu_v = 10^{-3} \text{ m}^2/\text{s}\) and \(\kappa_v = 10^{-5} \text{ m}^2/\text{s}\), in the horizontal and vertical respectively. Near the surface, vertical coefficients are specified by the KPP mixed layer model and can therefore be higher by more than an order of magnitude in the planetary boundary layer. With this horizontal viscosity, western boundary layers would be resolved only for zonal-grid spacing less than about 100 km (Bryan et al., 1975). To allow a time step of one hour, an implicit scheme is implemented for the vertical mixing. Initial conditions were obtained from the Levitus et al. (1994) climatological January potential temperature and salinity fields, with the velocity field then adjusted over a 1 month period. The initial (a priori) model forcing consists of NCEP re-analysis daily surface heat and freshwater fluxes, and twice-daily wind stress.

In the following we will show results that are an extension of the earlier ones provided by Stammer et al. (2002a,b). They are obtained by extending the estimation period to a total of nine years (1992 through 2000) and by including extra constraints to prevent model drift. A schematic of the optimization is provided in Fig. 1. In the present calculation, the control parameters include adjustments to the initial-condition potential temperature, \(\theta\), Fig. 1
and salinity, $S$, fields, as well as the daily surface momentum, heat and freshwater fluxes over the full nine years; i.e., we assume that the model uncertainties reside entirely in the initial conditions and surface forcing fields. Observations, to which the model is fit, are shown in the top part of the figure, include the daily TOPEX/POSEIDON (T/P) and ERS altimetric sea surface height (SSH) anomalies, the mean SSH, monthly mean SST fields, Levitus et al. (1994) monthly mean temperature and salinity climatologies over the entire water column as well as surface flux fields of momentum, heat and freshwater.

3 Adjusted Net Heat Flux

Consider first the estimated time mean net surface heat flux field from the period 1992 through 2000 (top left panel in Fig. 2). There is a pronounced east-west asymmetry across ocean basins and a clear north-south symmetry relative to the equator. While maximum heat loss can be found over all western boundary currents, the heat uptake is enhanced along most of the eastern continental boundaries, where the off-shore Ekman transport brings up cold water from below.

Consistent with published surface flux climatologies, most of the ocean heat uptake is concentrated at low latitudes. All subtropical regions lose heat to the atmosphere roughly along 20° latitude. However, subpolar regions tend to gain heat, e.g., in the North and South Pacific. A heat loss can be found again farther poleward, with maximum in the North Atlantic. An exception is the Indian Ocean that shows heat uptake along all northern hemisphere western boundaries. Note also the strong warming (60 W/m²) over Flemish Cap in the North Atlantic.

Mean changes in net surface heat flux relative to the prior NCEP fields are shown in the middle-left panel of Fig.2 (see also discussion in Stammer et al., 2002b). Differences between LN01 and NCEP fluxes are also available during 1993 alone, and are shown for comparison. Adjustments made by ECCO, and the corrections made by LN01 to the NCEP re-analysis have strikingly similar large-scale patterns and amplitudes. This conclusion is especially true for the heat flux over the entire Indian and Pacific oceans. Large regions in the ECCO adjustments show changes of ±20 W/m², values that are
entirely consistent with accepted uncertainties (WGASF, 2000). Over much of these same areas, LN01 show similar (although somewhat larger) differences with NCEP, indicating the independent approaches agree in the tendencies of their required large-scale changes to NCEP.

Some of the largest ECCO adjustments correspond to known problems with the reanalysis, and are consistent with LN01. Most notably, the lack of stratus clouds in the NCEP model over the eastern tropical Pacific and Atlantic has been identified as leading to an excess of short wave radiation into the ocean over those regions (Trenberth et al., 2001). The feature appears in both ECCO and LN01 analyses as regions of large negative heat flux differences. The state estimation procedure reduces the net flux by 20 - 40 W/m². Note also that the optimization removes some of the small-scale Gibbs effects known to be present in those regions in the initial NCEP estimate net heat flux fields, and that originate from the influence on the spectral model of mountain ranges such as the Andes.

Large flux adjustments occur, generally, over western boundary current regions, particularly in the Brazil/Malvinas confluence, the Agulhas retroflection, the Gulf Stream and the Kuroshio and their extensions. Much of this change has to be attributed to ocean model error, because in the present implementation boundary currents are not properly resolved or represented. Deviations between the ECCO changes and those from LN01 are largest over the Gulf Stream and its extension. It is interesting however, that those large differences are limited primarily to the vicinity of the path of the Stream. North and south of it we again find agreement between the two estimates, even in the Labrador Sea where both fields indicate substantial reduction of mean cooling relative to NCEP estimates. Renfrew et al. (2001) found from direct air-sea flux measurements further evidence that the NCEP net surface heat fluxes overestimate the heat loss in the Labrador Sea.

A comparison of the seasonal cycle of the ECCO fluxes with the Southampton Oceanography Centre (SOC) flux data set (Josey et al., 1999) shows that over the Gulf Stream and Kuroshio fluxes are in reasonable agreement from January through April and again in August. However, from September through November both SOC fluxes are smaller by about 50W/m². This bias suggests that the estimation process reduces the
cooling over those region during fall to avoid too deep a mixing in the absence of sufficient horizontal heat supply through the sluggish circulation—another indication that more model resolution is required.

Zonal-averages of the net surface heat are shown in Fig. 3. Curves are shown for the global ocean and for individual basins, respectively and can be compared with results from ECMWF (e.g., Garnier et al., 2000) and NCEP reanalyses products (Beranger et al., 1999; WGASF, 2000). The heat gains in the tropical Pacific and Atlantic have roughly the same amplitude per unit area; the Indian Ocean however, shows markedly reduced values. A common feature of heat flux climatologies is the pronounced warming of the Southern Ocean between about 40° S and 55° S. This warming is mostly located over the Atlantic and especially over the Indian Ocean sectors of the ACC. Note a similar relative maximum heat flux over the northern hemisphere roughly at the same latitude range where cooling rates go to near-zero; a positive heat gain is never reached there, however.

The northern hemisphere outside the tropical regime loses heat to the atmosphere as does the southern hemisphere between 10° and 40° S. Note that heat loss by the ocean in the southern hemisphere occurs in the South Pacific, but is most intense in the Indian Ocean. Much of the latter heat loss is imported from the Pacific through the Indonesian Througflow or the Southern Ocean (compare Stammer et al., 2002b). The Atlantic shows almost zero heat gain or loss between 10° S and 40° S, contrary to the heating in most other estimates, except the NCEP ones (e.g., Trenberth and Caron, 2001; Bryden and Imawaki, 2001; LN01).

Note that measurements from the Subduction Experiment buoy array flux reference sites (available for the period January 1992 through June 1993; see Moyer and Weller, 1997) lead to a mean net heat flux of 24 W/m² into the ocean over that part of the Atlantic. The corresponding ECCO estimates are 15 W/m² as compared to an NCEP value of 3 W/m²; i.e., the estimation procedure reduced the apparent NCEP bias by more than 10 W/m² or about 50%.
4 Adjusted Fresh Water Flux

Estimates of the surface net fresh water fluxes as they emerge during 1992-2000 are shown in the upper right panel of Fig. 2. The expected large-scale features are evident, and again we can find a clear east-west basin asymmetry, and a pronounced symmetry in the meridional direction. However, the symmetry is now relative to the ITCZ and not (as is the case for heat flux) relative to the equator. Negative net freshwater flux into the ocean, i.e., positive E-P, are present over the eastern side of all subtropical gyres, and the Arabian Sea. Losses are quite similar in pattern between Atlantic, Pacific and Indian Ocean, although the Indian Ocean shows enhanced evaporation around 40° S. The salinity maximum in the North Atlantic and its origin in the strong freshwater loss over the eastern subtropical Atlantic is quite well known. Here a similar forcing effect also appears in each of the other sub-tropical gyres.

Large net precipitation can be found in the tropical convergence zones, including the western Pacific warm pool, over the mid-latitude storm tracks, and along the Antarctic Circumpolar Current (ACC). Largest freshwater input occurs near the boundaries, and can often be associated there with river discharge (e.g., in the Gulf of Bengal, or near the Amazon delta) or with ice import and melting (e.g., over parts of the Labrador Sea and around Antarctica). Neither process is properly represented in the present ECCO model, but future calculations will include river discharge explicitly. The estimation procedure attempts, necessarily, to enhance net freshwater fluxes from the atmosphere over those regions to correct for the missing river runoff—an unintended test that the estimation procedure does indeed produce physically sensible results. As can be seen from Fig. 4, river discharge can be large—over the Amazon region reaching the equivalent of about 20 m/yr net precipitation, i.e., an order of magnitude larger than the maximum net freshwater from the atmosphere (Fig. ??). The river inflow shown in Fig. 4 was obtained from estimates of the climatological difference between precipitation and evaporation over each continent, which were then partitioned between neighboring ocean basins (Fekete et al., 1999), then distributed along the coast according to observed river discharge (Perry et al., 1996)
Differences between ECCO-estimated freshwater fluxes minus NCEP and LN01 minus NCEP are shown in the middle and lower right panels of Fig. 2, respectively. As with the heat fluxes, there are clear similarities between the difference fields, especially over the Pacific Ocean and most of the southern hemisphere. In both, more evaporation i.e., net fresh water loss, is required over the western tropical Pacific, the tropical Indian Ocean, the subpolar North Pacific and over the entire ACC. By contrast, a net input of freshwater is clear over the eastern tropical Pacific, large parts of the subtropical gyres and western boundary currents, especially the Kuroshio. In LN01, some of those changes are more pronounced (e.g., in the tropical Pacific) and reach further across the basins than they do in the ECCO results, but they are clearly present in the latter estimates as well. It is noteworthy that most of the estimated changes in net freshwater input into the ocean relative to NCEP are qualitatively and quantitatively consistent with the changes between the recent, mostly satellite-based HOAPS (Hamburg Ocean Atmosphere Parameters and fluxes from Satellite data) estimates (Grassl et al., 2000) and the Southampton Oceanographic Centre mean freshwater input estimates (Josey et al., 1999). Just to list a few examples: relative to the SOC estimates, the HOAPS atlas shows substantially enhanced net freshwater input along the entire coast of the eastern US and Canada. Moreover, the HOAPS atlas shows substantially more evaporation along 40°S in the southern hemisphere than suggested by the SOC fields suggest. Both changes are consistent with our adjustment of NCEP net freshwater fluxes, although the missing river run-off in the ECCO approach makes the comparison along the coast of North America only qualitative.

Zonally averaged fresh water flux estimates are shown in Fig. 3. Global and basin integrals are positive at low latitudes. However, maxima are shifted in their geographic position between basins: the Pacific Ocean has the maximum at around 5° N, while the Indian Ocean gains most freshwater at about 5° S. A gain in freshwater occurs also over high latitudes. Over the Southern Ocean, the gain is equally distributed over all three basin sectors. In the northern hemisphere, maximum precipitation can be found in the Atlantic Ocean. Although all subtropical regions lose freshwater, the largest loss occurs over the southern Indian Ocean, suggesting that the enhanced heat loss there occurs through latent heat release to the atmosphere.
5 Wind Stress Changes

The adjusted wind stress fields are shown in the top row of Fig. 5. Time-mean adjustments relative to the NCEP first guess are presented in the middle row of the figure. A substantial fine structure is visible in the modifications of the time-mean NCEP zonal and meridional stress components. Because our emphasis here is on the large scales, we show only spatially low-pass filtered difference fields with all variations on scales smaller than 5000 km removed. The fact that the largest modifications exist close to intense boundary current systems again indicates the difficulties this 2° horizontal resolution model has in producing the proper current separation without extra vorticity input by the modified wind stress. We note however, that even these large changes in the wind are all within the prior error bounds of the NCEP wind stress fields provided by Stammer et al. (2002a) and based upon scatterometer measurements. On the other hand, other changes, such as the increased trade winds over the tropical Pacific, are consistent with prior knowledge of NCEP re-analysis shortcomings (e.g., Milliff et al., 1999), and these adjustments may well be true corrections. Because of the apparent impact of the model defects on some local estimates of wind stress and wind stress curl, we need to make regional distinctions in the comparisons.

The lower row of Fig. 5 shows the difference between the mean wind stress components obtained from the ERS1/2 satellite measurements for 1992 through 2000, and the NCEP fields. To allow a better quantitative comparison of large scale structures in the wind stress difference field, we show again spatially low-pass filtered difference fields.

For the comparison, we used the ERS-1/2 gridded monthly wind stress fields on a global 1° by 1° grid (Bentamy et al, 1998; IFREMER, 2000). Note that the ERS-NCEP difference fields are very similar to the differences seen in the NSCAT comparison to NCEP by Milliff et al. (1999). Satellite observations point generally toward winds being too strong in the NCEP re-analysis: both westerly and easterly winds from the satellites tend to be weaker than in the NCEP fields. Meridional stress is less poleward at latitudes higher than about 30°, while from 10° to 30° latitude, the stress is reduced compared to the NCEP values. An exception can be found in the tropical regions, where the NCEP
fields show too weak easterlies. Another exception is the region of strengthened zonal stress to the west of Drake Passage in both ECCO and ERS values.

Over most of the ocean, the dynamically relevant forcing is the wind stress curl. We therefore compare in the left column of Fig. 6 the differences in mean wind stress curl as they emerge from our results minus NCEP and from and ERS minus NCEP fields, respectively. The comparison is limited to the large-scales over the lower latitudes where we have seen some skill in the ECCO estimates in improving NCEP stress estimates. As is to be expected, differences in the curl adjustments can be found in the vicinity of strong boundary currents, e.g., the Kuroshio. Note also the reversal of the dipole structure around Hawaii between the two fields. Nevertheless, agreements in the curl adjustment are obvious, and include bands of positive and negative curl adjustments that extend zonally across the Pacific. Even some regional changes are similar such as the positive anomaly in the eastern South Pacific that is disconnected from the coast by a negative band.

So far we have concentrated on corrections of biases in the NCEP surface flux fields. But the state estimation procedure adjusts the surface forcing on a daily basis and ultimately, as model physics improve, time-varying weather events may also become correctable. To provide some insight into the time-varying signal, we show in Fig. 7 a comparison of the zonal wind stress component measured at 170° W on the equator by one of the TOGA-TAO buoys with similar NCEP estimates (lower left panel) and ECCO estimates (upper left panel).

Observed winds at a normal height of 4m above sea surface were converted to winds at 10m-level reference height using the logarithmic wind profile equation. The observed parameters used in the calculation of the TAO wind stress are air temperature and relative humidity and their measurement heights, SST and its measurement depth, and surface pressure. Details are documented in papers on the algorithm (Fairall et al., 1996a,b). The algorithm is designed to give estimates of the turbulent fluxes of sensible and latent heat and the stress from inputs of bulk variables. Bulk transfer coefficients are based on the Liu et al. (1979) model with some modifications.
ous, indicating that the estimation at this location actually brings all variation on time scales longer than about one month into better agreement with the TAO measurements. Accordingly, scatter plots of the wind stress magnitude relative to TAO moves from a fairly flat slope of a least-squares fitted line toward the ideal angle of 45°. Similar results can be found along the entire equatorial regions.

For a better understanding of the temporal signals off the equator, we used fields from NSCAT for a comparison. Daily NSCAT wind stress was computed by using the Level 3 wind vectors available from the Jet Propulsion Laboratory. The Level 3 product is a daily average ocean surface wind vector map at a reference height of 10m on a rectangular uniform 0.5° x 0.5° latitude-longitude grid covering latitude range ±75°.

Figure 8 shows a summary of the comparison of 7-day running means with of the NSCAT data in form of a map of the improvement in the RMS difference between NCEP and NSCAT and between ECCO and NSCAT fields. Positive values indicate a reduction (i.e., improvement) of the RMS misfit when going from NCEP to ECCO stress fields. Improvements are limited to about ±20° latitude, and are most obvious for the Pacific Ocean. At most other latitudes, changes in the misfits are unstructured (values fluctuate around zero). But as was discussed before, all boundary current regions clearly appear as regions with large negative amplitudes, i.e., the estimation procedure degrades not only the mean wind stress, but also the weather events there, in order to correct the flow field. Improved ocean model physics are required, particularly through higher spatial resolution, to lead to improved surface flux fields over those regions as well.

6 Summary

Oceanic state estimation has advanced sufficiently that it is now possible to make estimates of the atmospheric forcing fields required to reproduce global ocean observations. Ocean applications of the estimated flow field are are described elsewhere (e.g., Stammer et al. 2002a,b; Ponte et al., 2001). Here we focussed on the surface exchanges with the atmosphere that provide a novel opportunity to obtain a better understanding of errors in existing flux products and ultimately to identify errors in atmospheric models. More
specifically, we used the data assimilation results from nine years of observations during the World Ocean Circulation Experiment to assess the adjustments of surface heat, freshwater and momentum fluxes, necessary to make the a priori ocean-forcing fields from the National Center for Environmental Prediction (NCEP) consistent with the observed ocean.

Because the ocean estimates are preliminary, the fluxes presented here are preliminary as well. They will improve as more ocean data are being included in the estimation procedure (e.g., new ARGO temperature and salinity profiles; Roemmich and Owens, 2000) and the model physics is being improved, e.g., by enhancing spatial resolution. Nevertheless the comparison presented with independent information about NCEP flux products is encouraging and shows the potential at hand.

In summary, the changes in the surface net heat and freshwater fields estimated through our ocean state estimation procedure are, overall, consistent with the independent information provided by LN01 and stay generally within assumed prior uncertainties in the meteorological analyses. Wind stress changes also stay within prescribed error bars, but show substantial small scale structures most simply explained as owing to ocean model error, primarily resolution near boundaries. All changes in surface flux fields are inferred here entirely through ocean observations and an ocean synthesis, illustrating the large reservoir of information about the atmosphere and the climate system residing in the ocean. The long-term goal is to eventually use this information to improve predictability on climate time scales, but also for medium-range weather predictions.

That observations of the ocean circulation could be used to infer atmospheric forcing fields was one of the motivations behind WOCE (e.g., Wunsch, 1984). It has taken the intervening years to deploy an observational system, develop adequate GCMs and estimation tools, to make the calculations practical. The results here are clearly preliminary, as they demonstrate remaining shortcomings of all three of the elements in the system, but considerable progress has nonetheless been made, and rapid improvements should now be possible.

**Acknowledgments:** CW acknowledges the hospitality of the Southampton Oceanography Centre where he worked with SJ on some of the issues presented here. Re-analysis
surface forcing fields from the National Center for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) are obtained through a computational grant at NCAR. The computational support through a NRAC grant from the National Partnership for Computational Infrastructure (NPACI) is acknowledged. Supported in part through NASA grant NAG5-7857, through NSF grant OCE 9730071 and through two contracts with the Jet Propulsion Laboratory (958125 and 1205624). This is a contribution of the Consortium for Estimating the Circulation and Climate of the Ocean (ECCO) funded by the National Oceanographic Partnership Program under grants N00014-99-1-1049 and N00014-99-1-1050.
References


Figure Captions

**Fig. 1:** Schematic for assimilation run 1992 through 2000. Data used as constraints are at the top of the figure, with altimetric observations in red, the Reynolds and Smith (1994) sea surface temperature analysis (SST) in green, the surface boundary data (stress, $\tau$, heat flux $H_q$, fresh water flux, $H_s$) from NCEP in black, and the Levitus et al. (1994) climatologies of temperature ($T$) and salinity ($S$) in blue. Lower part of the figure shows the elements of the control vector, which include the initial conditions on temperature and salinity ($T_0, S_0$), $\tau$, and $H_q$ and $H_s$.

**Fig. 2:** Top row: The mean net surface heat (left) and freshwater flux fields to/from the atmosphere (right) as they result from the optimization over the period 1992 through 2000. Middle row: Mean changes in net surface heat exchange relative to the prior NCEP fields estimated over the one-year period 1993 (in W/m$^2$; left panel), and for the net freshwater exchange (in W/m$^2$; right panel). Bottom row: Mean difference LN01 - NCEP net surface heat flux from 1993 (left panel) and for fresh water flux (right panel).

**Fig. 3:** Zonally integrated heat (top) and and surface fresh water fluxes (bottom), evaluated globally (blue curves) and over the Atlantic, Pacific and Indian Ocean sectors (green, red and magenta), respectively.

**Fig. 4:** Mean river run-off from Large and Nurser (2001). These contributions are compensated in the model by adjusting the over-ocean fluxes. Observations gave the mean runoff at the mouths of about 200 gauged rivers (Perry et al., 1996), which typically accounts for 40 to 60/into each basin from ungauged rivers was evenly distributed along its coast. The runoff was converted into the surface freshwater flux, by spreading it out over an area near its source. This spreading decreased exponentially with a 1000km $e$-folding distance, as suggested by the observations of surface salinity off the mouths of the Amazon and Congo rivers.

**Fig. 5:** Top row: The mean surface zonal wind stress (left) and meridional wind stress fields (right) as they result from the optimization over the period 1992 through 2000 (in N/m$^2$). Middle row: Mean changes in meridional wind stress relative to the prior NCEP.
fields estimated over the six-year period 1992-1997 (in N/m²; left), and for the meridional component (in N/m², right). Bottom row: The left panel shows mean difference in ERS zonal wind stress from 1992 through 1997 minus net NCEP surface heat fluxes from the same period. Right panel is the same, but for the meridional stress. All difference fields have been spatially low-pass filtered to show only structures larger than about 5000km.

Fig. 6: The differences in wind stress curl estimated from mean ECCO wind fields minus those from NCEP (top) and the differences with the ERS 1/2 wind stress fields for the period 1992 through 1997. Units are 10⁻⁷ N/m³.

Fig. 7: Comparison of ECCO and NCEP wind stress fields with TOGA TAO measurements at 170° W on the equator. The top panel shows a timeseries of zonal TOA wind stress components (blue) and ECCO results (purple). The middle panel shows a similar plot, but with NCEP fields in red. The bottom panel shows scatter diagrams of NCEP (red) and ECCO (blue) wind magnitude measurement at the same position with the TAO data.

Fig. 8: RMS reduction in the difference between ECCO and NSCAT wind stress time-series as compared to the original NCEP-NSCAT differences. Positive values indicate a respective misfit reduction (in N/m²).
Figure 1:

Data Constraints

Controls

T^2\,SST

\text{mean TP SSH} - \text{EGM96}

\text{monthly}

\text{monthly}

\text{monthly}

\text{daily TP SSH'}

\text{EGS-3 SST}

\text{EGS-1 SST}

\text{EGS-2 SST}

\text{EGS-4 SST}

\text{EGS-5 SST}

\text{EGS-6 SST}

\text{EGS-7 SST}


Figure 2:
Figure 3:
Figure 4:
Figure 5:
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Figure 8: