Modification of Surface Fluxes from Component Models in Global Coupled Models

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ABSTRACT

The present generation of global coupled ocean–atmosphere GCMs contains considerable systematic errors both in terms of net surface heat flux and simulated SSTs. Here, a global coupled GCM is used to illustrate how systematic errors in the separate coupled model components (atmosphere and ocean) contribute to the simulations of net surface heat flux and SST when the components are coupled together. Features of the coupled model simulation are a combination of errors in the component models and errors introduced due to the dynamic interaction, both local and nonlocal, between atmosphere and ocean. Various regions and latitudinal zones are examined to determine the processes that produce the net surface heat fluxes and SSTs in the coupled simulation. In the coupled model, a good simulation of net surface heat flux does not always produce a correspondingly accurate simulation of SST. Alterations of surface winds and/or ocean currents can introduce SST errors and consequent compensating surface fluxes that have apparent agreement with observed estimates (e.g., near 60°N in the North Atlantic). Additionally, an SST error that occurs due to a combination of surface flux errors from atmosphere and ocean components in the coupled simulation, as well as an alteration of the ocean surface currents, can produce a better agreement of the net surface fluxes in the coupled model with observed estimates in spite of the large SST errors (e.g., near 50°N in the Atlantic and Pacific). Conversely, a good simulation of SST in the coupled model can be associated with surface heat flux errors also due to dynamic adjustments in the atmosphere and ocean in the coupled simulation (e.g., near 20°N and 20°S). A high-quality coupled model simulation does not necessarily require a precise reproduction of observed net surface heat fluxes, even though accurate observed surface fluxes are necessary to calibrate model parameterizations in the components and to provide an index of model performance. Rather, improved coupled model simulations must rely on improvement of the entire thermodynamic and dynamic simulations (and verification of state variables) in the components separately and when coupled.

1. Introduction

It is well known that present-day global coupled climate models contain considerable errors in terms of surface heat fluxes (Gates et al. 1993). These errors are thought to arise from a combination of errors in the component models and errors that result from the dynamic interaction between components when they are coupled (Meehl 1990, 1992, 1995). In light of these errors, some groups have chosen to use flux adjustment techniques to bring the coupled model control integration more into line with the observations. However, these flux adjustments can be as large as the fluxes themselves (Gates et al. 1993).

For example, for the Hadley Centre coupled model (Murphy 1995), the general pattern of net surface heat flux (Fig. 1a) compares qualitatively well with various observed estimates (e.g., Oberhuber 1988; da Silva et al. 1994) with general heat uptake (positive net surface heat flux values in Fig. 1a) in the Tropics and heat release (negative net surface heat flux values in Fig. 1a) in the extratropics. However, to obtain an idea of the magnitude of the errors of the simulated net surface heat flux from that model, Fig. 1b shows that heat flux adjustment terms are near \( \pm 50 \text{ W m}^{-2} \) in the equatorial Pacific and north of Scandinavia, with large negative values in the southern midlatitudes. Positive flux corrections indicate SSTs in the coupled model that were tending to be too cold and that require positive heat input in the form of flux adjustments to bring them closer to the observed values. These net surface heat flux errors are also associated with errors in simulated SSTs in the coupled model in Fig. 1c. The largest SST errors generally occur in the regions of the strongest mean SST gradient. For example, SST errors of greater than 6°C occur in the region of the Gulf Stream and the Kuroshio current systems in the Northern Hemisphere.

From this example, it can be seen that net surface heat flux errors in a global coupled model are accompanied by a set of simulated SST errors as well. There are various compensating and interacting processes that produce this unique combination of net heat flux and SST errors.
Fig. 1. (a) Annual mean net surface heat flux from the Hadley Centre global coupled GCM, with a contour interval of 50 W m\(^{-2}\); positive values indicate heat flux into the ocean, and negative values are stippled. (b) Annual mean flux correction terms in the Hadley Centre global coupled model indicative of errors in the net surface heat fluxes, with a contour interval of 50 W m\(^{-2}\); positive values indicate areas where heat flux must be added to the ocean to bring the model close to observations, and negative values are stippled. (c) Annual mean SST errors, computed minus observed, for the Hadley Centre global coupled model run without flux correction, with a contour interval of 1°C; negative values are stippled and indicate that the coupled model SSTs are less than observed. (All after Murphy 1995).
SST errors (similar arguments apply to net freshwater flux and wind stress error fields as well; for the purposes of illustration, we concentrate on net surface heat flux errors in this paper). Some can be directly traced to preexisting systematic errors in either one or both component models of the atmosphere and ocean. However, the net heat flux values in a coupled model are not always indicative of the heat flux errors in the component models. Instead, the interactions noted above result in simulation errors that have been addressed in atmospheric models coupled with nondynamic slab oceans (Meehl and Washington 1985) and in atmospheric models coupled with dynamic ocean models without flux adjustments (Meehl 1989). The product is a set of surface heat fluxes with various systematic errors and a simulated SST field with its own set of systematic errors, as we have noted in Fig. 1.

Thus, the coupled interaction between components produces a unique set of SSTs and surface heat fluxes, with their own set of respective systematic errors, representing various compromises between competing processes as the models communicate and adjust to each other’s set of forcings. This paper will include analyses of these processes in terms of systematic errors in the components and how those interactions occur to produce the simulated surface heat fluxes and SSTs in a coupled model. The purpose is to show that, for coupled modeling, precise details are not necessarily required of observed surface fluxes, although accurate observed surface flux datasets are useful for calibrating coupled model performance. Rather, a successful coupled modeling strategy requires improvements to the model component simulations in stand-alone mode and in concert with improvements to the coupled model simulations. Therefore, high-quality observed datasets of state variables are at least as important if not more so than observed surface flux datasets of comparable quality. Here, we illustrate how these features end up being manifested in a global coupled GCM.

2. The model and experiments

The global coupled model used in this study was developed at the National Center for Atmospheric Research (NCAR), and is a predecessor to the NCAR Climate System Model (CSM). The atmospheric component is a global spectral R15 9-level GCM with mass flux convection, parameterized cloud albedo feedback, computed clouds, and bucket soil moisture (Meehl and Washington 1995; Washington and Meehl 1996). The ocean is a global 1° 20-level ocean GCM (Washington et al. 1994) that includes dynamic (Flato and Hibler 1990) as well as thermodynamic (Semtner 1976) sea ice. The models were first run alone with their respective observed boundary forcings, and then coupled and run for a 145-yr control integration (Washington and Meehl 1996).

Three simulations will be examined here. Annual averages are computed over the last 5 yr of a 7-yr simulation with the atmospheric model run alone with observed monthly mean SST and sea ice distributions (Alexander and Mobley 1976). This integration will be termed the “atmosphere-only” experiment. This simulation, when compared to observations, should provide information concerning systematic errors in the atmospheric model. These errors, of course, will be passed to the ocean component when coupled.

The “ocean-only” experiment will refer to annual mean 10-yr averages computed at the end of the spinup integration of the ocean GCM run by itself with observed forcing (see below). The ocean model was first initialized with the three-dimensional observed temperature and salinity data from Levitus (1982), and then run with only surface forcing from observed monthly mean wind stress (Hellerman and Rosenstein 1983) and the Levitus SST and surface salinities (SST and salinity have a 1-month restorative timescale) for 100 yr (Washington et al. 1994). The Levitus (1982) forcing data have been adjusted to correct for inconsistencies in time and space, and these data were used by Semtner and Chervin (1988) and Washington et al. (1994).

Using such an initialization spinup procedure, many of the fast drifts in the ocean model occur during the initial 50 yr of integration. Washington et al. (1994) and Washington and Meehl (1996) discuss residual drifts in the ocean of mostly less than 0.1°C century⁻¹ by the end of the 100-yr spinup. Though the globally averaged surface air temperatures are relative stable (Washington and Meehl 1996), small flux imbalances (less than 2 W m⁻²) between the atmosphere and ocean are partially compensated for by such slow drifts in the system. A discussion of surface flux components by latitudinal zone is included in section 5. An alternative spinup scheme such as that used in the NCAR CSM can use forcing from the atmosphere model to run the ocean alone prior to coupling. Then, on coupling, there is less “coupling shock” since the ocean has already had a chance to adjust to systematic errors in the atmospheric model. However, errors are introduced into the ocean model, and the coupled system has those errors built in to its initial state. Yet, the opportunities for slow drifts of the coupled model are reduced as a consequence (Boville and Gent 1997).

The values of net surface heat flux from the ocean-only experiment (see Meehl et al. 1982 for technique) provide information concerning which heat fluxes the ocean requires to keep SSTs near observed values. A comparison with observed estimates (e.g., Oberhuber 1988) of the heat fluxes from the ocean-only experiment gives an indication of systematic errors in the ocean component that will affect interactions with the atmospheric component when coupled.

The third simulation will simply be termed “coupled” to refer to annual mean 10-yr averages from years 65–74 of a control run intended to simulate present-day climate with the atmosphere, ocean, and sea ice com-
components coupled together. This 10-yr period is from a similar time interval studied in the climate sensitivity experiments of Washington and Meehl (1996). A comparison of net surface heat fluxes from the coupled integration with observed estimates (e.g., Oberhuber 1988) and the atmosphere-only and ocean-only integrations will provide insight into how the systematic errors in each of the components contribute to errors in the coupled integration. Additionally, we will see how dynamic interaction between the components leads to net heat fluxes and SSTs in the coupled model that sometimes compare favorably to observed estimates and sometimes have produced their own set of errors that could not have been anticipated from examination of systematic errors in the components alone.

It should also be noted that the analyses presented here are intended to illustrate some of the problems that arise when component models are coupled, but, as such, the model errors are specific to this model. However, some model errors described here are typical of errors in other coupled models, and these will be noted where appropriate.

An uncertainty inherent in any comparison of model fluxes to “observations” is the difficulties in producing observational datasets. The question of what “truth” is for the observations of surface fluxes is not insignificant and arises from many sources. These include measurement errors, sampling errors, flux formulation errors, etc. (Oberhuber 1988; da Silva et al. 1994; WCRP 1996). This must be kept in mind whenever model-generated fields are compared to the observed estimates of surface fluxes in this paper. Though other observed estimates have been derived more recently, all still have similar problems (WCRP 1996). Thus, the observed estimates of Oberhuber (1988) are used here mainly as a point of reference to help interpret the model simulations and to illustrate the role of surface flux datasets in global coupled modeling.
3. Net surface heat flux simulations

Figure 2 depicts annual mean net surface heat flux values from the atmosphere-only simulation (Fig. 2a), the ocean-only simulation (Fig. 2b), the coupled simulation (Fig. 2c), an estimate of observed annual mean net surface heat flux from Oberhuber (1988, Fig. 2d), and the difference between atmosphere-only and ocean-only net surface heat fluxes (Fig. 2e). The net surface heat flux is defined here (following Meehl 1992) as

\[ F_{\text{net}} = S - LE - H - (F_\uparrow - F_\downarrow) \]

where \( F_{\text{net}} \) is the net surface heat flux, \( S \) is the solar radiation absorbed at the surface, \( LE \) is the latent heat flux, \( H \) is the sensible heat flux, and \( (F_\uparrow - F_\downarrow) \) is the net infrared radiation at the surface. Note that absorbed solar radiation warms the surface, and the other three components act to cool the surface. As described in Fig. 1 and depicted in Fig. 2d, the effect of net surface heat flux is to warm the tropical oceans (positive heat flux into the ocean) and cool the high-latitude oceans (negative heat flux out of the ocean). In the absence of long-term drift, the net heat flux is an indicator of the contribution of ocean processes to the surface energy balance. Thus, from this first-order description of net heat fluxes, it can be inferred that the oceans transport excess heat from the Tropics poleward, and errors in the net heat fluxes can affect the implied heat transport by the ocean (see discussion below in relation to Fig. 5). 

Comparing Figs. 2a–c, all three model simulations appear to have simulated the qualitative representation of surface heat fluxes, yet with important differences regionally and in magnitude. The atmosphere-only simulation produces net surface heat fluxes of greater magnitude than the ocean-only simulation at almost all lo-
Fig. 4. (a) Annual mean surface wind stress from the atmosphere-only simulation, with a scaling vector at lower right in $10^{-2}$ dynes cm$^{-2}$ or N m$^{-2}$. (b) Same as (a) except for the ocean-only simulation. (c) Difference of the atmosphere-only minus ocean-only surface wind stress [(a) minus (b)]; note that ocean-only wind stresses are observed estimates from Hellerman and Rosenstein (1983). (d) Difference of the coupled minus the ocean-only simulation.

cations, with generally larger heat flux values than the observed estimates in Fig. 2d. For example, in Fig. 2a, the negative values in the subtropical ocean regions of $-100$ W m$^{-2}$ are near twice the the peak values of about $-50$ W m$^{-2}$ in the observed estimates. There also appear to be positive net heat flux values in the atmosphere-only version in Fig. 2a in the North Pacific and near 50°S in the southern oceans that do not appear to be consistent with the observed estimates (though there are few observations near 50°S in Fig. 2d). Additionally, there are large positive values of net heat flux off the west coasts of the subtropical continents that do not agree with the observed estimates in Fig. 2d. This is likely related in part to a problem with the simulation of boundary layer stratocumulus clouds (Robertson et al. 1995).

The ocean-only net heat fluxes (required to keep the ocean SSTs close to observations) in Fig. 2b are of smaller magnitude than the fluxes in the atmosphere-only simulation and in closer agreement (in terms of magnitude) with the observed estimates in most regions. For example, the subtropical minima are near $-25$ W m$^{-2}$, with peak values of near $-50$ W m$^{-2}$, as in the observed estimates. Negative values in the high southern latitudes and in the North Pacific and North Atlantic also appear to be more in agreement with the observed
estimates. However, there are still regional values that do not agree with observed estimates. These suggest that there could be problems simulating correct SSTs in those regions when the ocean is coupled to the atmosphere. In particular, in the equatorial central Pacific, the ocean model requires more than $+50 \text{ W m}^{-2}$ to keep SSTs near observed, while the observed estimates show values of around $+25 \text{ W m}^{-2}$. Meanwhile, in the eastern equatorial Pacific, the ocean model requires net heat loss (values of about $-25 \text{ W m}^{-2}$), while the observed estimates show large positive values of about $+75$ to $+100 \text{ W m}^{-2}$, a disagreement in magnitude but also in sign with the ocean-only simulation in Fig. 2b.

From examination of Figs. 2a and 2b, at the very least it would appear that there would be large surface heat flux errors at almost all latitudes in the coupled simulation. However, inspection of the coupled simulation of net heat flux in Fig. 2c shows that there is qualitative agreement with the observed estimates in Fig. 2d both in terms of the magnitude and sign of net surface heat fluxes over many regions. This implies that compensations have occurred as the components interact, with one consequence being the SST errors discussed below.

The difference between ocean-only and atmosphere-only in Fig. 2e shows that the large-amplitude heat fluxes in the atmosphere-only simulation have been damped in the coupled simulation. In particular, the subtropical negative heat flux values in the coupled simulation in Fig. 2c are more in line with the $-25$ to $-50 \text{ W m}^{-2}$ values of the observed estimates in Fig. 2d and are manifested by negative differences in those areas in Fig. 2e. The large positive values off the west coasts of the subtropical continents have been reduced and even

![Fig. 4. (Continued)](image-url)
changed to net heat loss in better agreement with the observed estimates near 20°S west of South America, thus producing positive differences in Fig. 2e. There are more regions in the North Pacific showing net heat loss (negative heat flux values) in the coupled model than in the atmosphere-only simulation, which is also in better agreement with observed estimates and evidenced by positive differences in Fig. 2e. In the tropical Pacific, where major problems were identified in both the atmosphere-only and ocean-only simulations above, the coupled model has a better simulation of the fluxes than in the components separately (though with resulting SST errors, as discussed below). The coupled model in Fig. 2c shows that in this region there is good agreement with observed estimates, as both show maximum positive heat flux values of +75 to +100 W m⁻² from about 5°N to 5°S and from 140° to 90°W. Therefore, in spite of surface flux errors in the components, the coupled model does not stray into a totally foreign climate regime in terms of surface fluxes, though SST errors in certain regions are a consequence.

4. SST errors

The net surface heat flux simulation is only one aspect of the adjustments made in the coupled model as the components interact with each other. The other, as noted above, is SST error. Figure 3a shows the SST simulation from the ocean-only model. Recall that the SST forcing has a 1-month restorative timescale (Washington et al. 1994). This means that the SSTs are not restored exactly to the observed values of Levitus (1982). The atmosphere-only simulation (Fig. 3b) used a somewhat different observed SST dataset (Alexander and Mobley 1976). Small differences in the details of the observed SST datasets (e.g., see Shea et al. 1990) could contribute to initial inconsistencies between components. But the SST errors from the ocean-only model (due to the 1-month restorative timescale of the forcing with their own set of relatively large systematic errors—on the order of several degrees in Fig. 3d) overwhelm small inconsistencies between observed SST datasets.

Figure 3c shows the annual mean SST simulation from the coupled model, which can be compared to a plot of annual mean observed SSTs in Fig. 3b. There is overall qualitative agreement between the two, with evidence of an equatorial cold tongue in the eastern Pacific in the coupled model and intensified SST gradients in the North Pacific, North Atlantic, and around Antarctica, indicating some representation of the Kuroshio, Gulf Stream, and Antarctic Circumpolar Currents, respectively. Closer inspection reveals the type of error noted in Fig. 1, involving small shifts of strong SST gradients. The strong SST gradient regions in the midlatitude oceans in the coupled model in Fig. 3c appear to be shifted poleward compared to the observed estimates in Fig. 3b. This results in large positive SST errors in those regions in the difference field, coupled model SSTs minus the observed, in Fig. 3e. This is opposite to the systematic SST errors in those regions in the ocean-only simulation (Fig. 3d).

It is likely that problems simulating the boundary layer stratocumulus regimes, especially off the coast of South America, contribute to the problems with simulating the proper orientation of the cold tongue in that region (Mechoso et al. 1995). This produces a positive SST error in Fig. 3e in the eastern tropical Pacific and eastern tropical Atlantic. Elsewhere, however, SST errors in this non-flux-corrected simulation are generally less than 1°. This indicates that areas of strong SST gradient in the midlatitudes and the eastern equatorial oceans have the greatest SST errors, while elsewhere, in the less dynamically active regions of the oceans, the simulated SSTs are within about 1° of the observed values.

A comparison of Figs. 2e and 3e suggests that surface flux errors are a contributing factor to coupled model SST errors in some regions. For example, some of the positive SST errors in the midlatitudes in both hemispheres are associated with positive heat flux anomalies in Fig. 2e. There is also some correspondence between the positive SST errors off the west coasts of North and South America and Africa, and the positive heat flux differences in Fig. 2e. Thus, local flux imbalances provide part of the explanation for the SST errors. However, nonlocal dynamic effects, as evidenced by errors in surface wind stress in Fig. 4, are also playing a role.

Surface wind stresses from the atmosphere-only simulation (Fig. 4a) appear to qualitatively reproduce the major surface wind regimes depicted in Fig. 4b (a plot of an observed estimate of surface wind stress used to drive the ocean-only simulation) in terms of the approximate strength and location of the midlatitude westerlies and easterly trade wind regimes. However, significant systematic errors in the wind stress simulated for the atmosphere only are shown in Fig. 4c, which is a plot of the differences between atmosphere-only and observed estimates of surface wind stress in Fig. 4b. The subtropical and midlatitude westerlies in the North Pacific and North Atlantic are weaker than the observed estimates by around 0.5 dynes cm⁻². This implies that, when coupled to the ocean, the atmospheric wind stresses would cause the ocean circulation to slow down, with a resulting shift of SST gradients. (This aspect will be discussed later in this paper in regards to Fig. 8.) A further examination of the changes in the coupled model in comparison to the components of the surface fluxes is given in the next section.

5. Zonal mean analyses

We now examine the interaction between model components to address further why these various heat flux and SST errors are manifested the way they are in the coupled model. Figure 5 is a plot of annual and zonal mean total poleward ocean heat transport from the
ocean-only model, a similar plot from the coupled model, and the implied ocean heat transport from the atmosphere-only model. It was mentioned earlier in regard to Fig. 2 that the atmosphere-only model appeared to have excessive heat flux into the ocean in the midlatitudes and that such surface heat flux errors could have implications for the implied ocean heat transport. Indeed, as shown in Fig. 5, there is excessive equatorward implied ocean heat transport in the atmosphere-only simulation near 40°N and 40°S. The ocean-only heat transport thus exceeds the northward heat transport of the coupled model near 40°N, while near 40°–50°S the coupled model ocean heat transport is northward. Observed estimates (e.g., Trenberth and Solomon 1994) show poleward heat transport in each hemisphere, with maximum values of near 2 pW at about 15°N and 15°S. The systematic errors from the atmosphere clearly influence the coupled model ocean heat transport, at least in part as a result of errors in the surface heat flux in the midlatitudes, particularly in the Southern Hemisphere. Such systematic errors are typical of atmosphere-only simulations (e.g., Gleckler et al. 1995) and are thought to arise, in part, due to errors in the simulation of clouds.

Zonal mean net surface heat flux from the components and the coupled model compared to the observed estimates from Oberhuber (1988), as well as the resulting zonal mean SST errors in the coupled model, are shown in Fig. 6. Here, the large positive surface heat fluxes from the atmosphere are clearly seen near 40°N and 40°–50°S in the atmosphere-only simulation. These exceed both the ocean-only heat fluxes and the observed estimates as well, pointing to the likelihood of errors in the atmospheric model. Not surprisingly, the largest SST errors simulated by the coupled model occur at those same latitudes (with peak values of around 5°C near 50°S and maxima of about 5°–8°C near 40°–60°N). However, it was noted earlier that errors in surface heat flux are only part of the problem. Figure 4 showed errors in surface wind stress simulated by the atmospheric model that also contribute to weakening the ocean circulation in the subtropical gyres, thus contributing to shifts in the maximum SST gradients. Such shifts also contribute to the SST errors seen at those latitudes, as shown in Fig. 6b.

To help quantify these effects, a number of zonal mean sectors are chosen to examine the various contributors to these errors in terms of the surface energy
Fig. 7. Zonal mean latitudinal band annual averages, ocean points only, of surface energy balance components, net surface heat flux, zonal surface wind stress and SST; observed estimates (dark-shaded bars) for surface energy balance components and surface heat flux from Oberhuber (1988), SST from Alexander and Mobley (1976), and zonal (u component) wind stress from Hellerman and Rosenstein (1983), light-shaded bars are ocean only, single-hatched bars are atmosphere only, and double hatched bars are coupled, for latitude bands (a) 60°S–50°S, (b) 25°–15°S, (c) 15°S–15°N, (d) 15°–25°N, (e) 45°–55°N, and (f) 55°–65°N. There are no observed estimates of surface heat flux and surface energy balance components given for part (a). Ocean-only net heat flux is formulated as the magnitude of the restorative term derived from the surface temperature forcing, and consequently no surface energy balance components are given. Note the different vertical scales—scale at far left is dynes cm⁻² for u-wind stress, scale second from left is W m⁻² for net heat flux components, and scale at right is °C for SST.
balance components [Eq. (1)], net surface heat flux, \( u \)-component wind stress, and SST. First, concentrating on the zonal mean latitudinal band averaged from 60° to 50°S in Fig. 7a, there is a positive SST error in the coupled simulation of 3.2°C. This region was identified in Fig. 3 to have a southward-shifted SST gradient that contributed to these positive SST errors. Looking at the zonal mean net heat fluxes in Fig. 7a, the ocean-only simulation requires near-zero heat flux; the atmosphere alone provides \( +22 \) W m\(^{-2} \), but the coupled model ends up with near-zero net surface heat flux. Thus, the coupled model net surface heat flux ends up close to that from the ocean-only simulation, but with a positive SST error. This arises due to problems noted above in simulating cloud cover, and additionally, the surface wind stress is weaker than the observed estimates (near 0.7 dynes cm\(^{-2} \)) for both the atmosphere-only and coupled simulations, with the observed estimate being about 1.0 dynes cm\(^{-2} \); this observed value has also been estimated to be even higher by Trenberth et al. 1990). This causes the ocean circulation to slow down in the coupled model compared to the ocean-only simulation (not shown) and contributes to the southward shift of SST gradients. Though there are no reliable observed estimates of the surface heat flux at these latitudes, the large positive value of this quantity from the atmosphere-only simulation is a typical systematic error in atmospheric GCMs (Gleckler et al. 1995). In addition, the apparent errors in the implied ocean heat transport values in Fig. 5 point to surface wind stress and surface heat flux errors that combine to contribute to higher-than-observed SSTs at these latitudes (for this latitudinal band, the coupled model SST value is 6.4°C, while the observed estimate and ocean-only values are 3.2°C and 3.3°C, respectively). The various surface heat flux components are comparable between the atmosphere-only and coupled simulations, with the exception of latent heat flux (58 W m\(^{-2} \) in the coupled model vs 42 W m\(^{-2} \) for the atmosphere-only). This discrepancy is mainly due to the higher SST values in the coupled model contributing to increased evaporation from the warmer surface. This extra 16 W m\(^{-2} \), removed from the warmer ocean surface, helps to balance the positive downward net heat flux from the atmosphere (22 W m\(^{-2} \)), resulting in near-zero net surface heat flux in the coupled model near 55°S, but with the positive SST error noted above.

The latitudinal zones averaged over 25°–15°S and 15°–25°N (Figs. 7b and 7d) experience similar simulation traits and will be discussed together here. The atmosphere-alone simulation has negative values of net surface heat flux (\(-22\) and \(-28\) W m\(^{-2} \), respectively), compared to the observed estimates at those latitudes of \(-10\) W m\(^{-2} \). The ocean-only and coupled models have near-zero net surface heat flux values (Fig. 4a). Interestingly, the zonal mean SST errors at 20°N and 20°S in Fig. 5b are less than 1°C, indicating that there is almost no error in simulating zonal mean SST near 20°N and 20°S (Fig. 6b), even though the zonal mean coupled model net heat fluxes are in error. Part of the reason for this can be seen in Fig. 3. The SSTs in the eastern tropical portions of the oceans increase in the coupled model (Fig. 3e) due to a combination of erroneously
large positive net heat fluxes in the atmosphere-only (Fig. 2e) simulation and the erroneously negative net surface heat flux in that region required by the ocean (Fig. 2b). Precipitation and surface winds respond to these SST changes in the coupled model, with positive precipitation anomalies (not shown) forming over areas where the SSTs have increased in the eastern tropical basins (Fig. 3e). The surface wind stress changes associated with these precipitation anomalies involve decreases of the strength of the easterly trade winds (from $-0.7$ to $-0.6$ dynes cm$^{-2}$), thus contributing to a decrease in evaporation and latent heat flux (from 176 to 170 W m$^{-2}$ at 25$^\circ$–15$^\circ$S and from 191 to 173 W m$^{-2}$ at 15$^\circ$–25$^\circ$N in Figs. 7b and 7d), a reduction in cloud (not shown), and a corresponding increase in absorbed solar radiation at the surface (from 209 to 221 W m$^{-2}$ at 25$^\circ$–15$^\circ$S and from 211 to 222 W m$^{-2}$ at 15$^\circ$–25$^\circ$N). This effectively increases the amount of heat flux into the ocean surface in the coupled model, thus helping to compensate for the negative surface heat flux in the atmosphere-only simulation and bringing about more consistency with the near-zero values required by the ocean-only simulation. The result is net surface heat fluxes in error by around 20 W m$^{-2}$, but SSTs close to the observed estimates. Note also the compensating errors in absorbed solar (around 35 W m$^{-2}$) and latent heat fluxes (roughly 40 W m$^{-2}$) in both the atmosphere-only and coupled simulations, with the excess absorbed solar flux nearly balanced by excess heat removal by the latent heat flux.

For net surface heat flux in the equatorial zone in Fig. 7c, the ocean requires 18 W m$^{-2}$, while the atmosphere provides 28 W m$^{-2}$. The coupled model ends up with the ocean-only value of 18 W m$^{-2}$, which is nearly that of the observed estimate of 20 W m$^{-2}$. However, increases of absorbed solar flux (223 to 231 W m$^{-2}$) contribute to higher SSTs in the coupled model by about 1$^\circ$, with corresponding increases of latent heat flux (154 to 171 W m$^{-2}$). This increased removal of heat from the ocean surface reduces the value of net heat flux from the atmosphere-only simulation, making it coincidentally close to the observed estimate, as noted above.

For the latitudinal zone from 45$^\circ$–55$^\circ$N (Fig. 7e), the ocean requires near-zero net heat flux, while the atmosphere-only simulation provides 25 W m$^{-2}$. However, the coupled model ends up with a net heat flux value of $-11$ W m$^{-2}$, a seemingly unlikely outcome given the fluxes from the component models. Simulated SSTs from this latitudinal zone are 6.2$^\circ$C warmer than those observed in the zonal mean (Fig. 7e; also refer to Fig. 3e for pattern of SST errors), even though the ocean-only SSTs correspond very closely to the observed estimates (Fig. 7e). The SST errors at this latitude arise, in part, due to a northward shift of the SST gradient in the North Pacific and North Atlantic, as suggested earlier in relation to the surface wind stress errors. Zonal wind stress is 0.74 dynes cm$^{-2}$ in the coupled model, compared to 0.83 dynes cm$^{-2}$ in the observed estimates in Fig. 7e. As a consequence, the subtropical gyres spin down in the ocean model component. In the coupled model, the Gulf Stream (Figs. 8a and 8b) and the Kuroshio currents (Figs. 8c and 8d) are weaker by about 40% and tend to have a more northward track than in the ocean-only simulation, thus advecting warmer water farther north than in the ocean-only run.

![Ocean Surface Currents](image)
result is a contribution to a positive SST error through a northward shift of the strong SST gradients in the North Pacific and North Atlantic, in addition to the contribution of positive net heat flux from the atmosphere-only noted above in Fig. 7e. As the coupled model experiences these warm SSTs at that latitude, the latent heat flux increases in Fig. 7e (the atmosphere-only and the observed estimates are close at 56 and 54 W m\(^{-2}\), respectively, but the coupled model has an increased latent heat flux of 91 W m\(^{-2}\)). This greater heat loss results in negative net surface heat flux in the coupled model, in closer agreement with the observed estimates, since there are negligible changes in the other components of the surface energy balance in Fig. 7e. Yet the coupled model is still left with a positive SST error due to the combination of errors in the atmosphere and the dynamic response of the ocean in the coupled simulation.

In the latitudinal zone 55°–65°N, there is another peak in SST error that is evident in Fig. 6b and quantified in Fig. 7f (almost entirely from the North Atlantic) of 6.2°C. Strong Norwegian and Spitsbergen currents (Washington and Meehl 1996) advect the warm water from the Gulf Stream and North Atlantic currents (flowing farther north than in the ocean-only integration, as noted in Figs. 8a and 8b) into the areas north of England and Scandinavia. The result is warmer water there in the coupled model and increases of the latent heat flux in Fig. 7f from 51 W m\(^{-2}\) in the atmosphere-only model to 77 W m\(^{-2}\) in the coupled model (compared to the
observed estimate of 65 W m\(^{-2}\)). This increased heat removal from the warmer surface pushes the zonal mean net heat flux value for the coupled model (\(-46\) W m\(^{-2}\) in Fig. 7f) into closer agreement with the observed estimate of \(-58\) W m\(^{-2}\), compared to net surface heat flux values of only \(-21\) and \(-24\) W m\(^{-2}\) in the atmosphere-only and ocean-only simulations, respectively. Thus, near 60\(^\circ\)N, the atmosphere-only and ocean-only simulations are in relatively good agreement in terms of net heat flux in Fig. 7f, but there is excessive absorbed solar radiation at the surface in the atmosphere-only simulation (111 W m\(^{-2}\), compared to the observed estimate of 78 W m\(^{-2}\)). This excess heat input combines with the changes in ocean advection from the northward-shifted North Atlantic current that provides warmer water to the Norwegian and Spitsbergen currents to produce a positive SST error in the coupled model.

6. Conclusions

Global coupled models are typically characterized by significant net surface heat flux and SST errors. These errors are a combination of errors in the component models and those resulting from the dynamic interaction between the components when they are coupled. Here, we use a global coupled model developed at the National Center for Atmospheric Research (NCAR) to illustrate some of these interactions for various regions and latitudinal zones. What the ocean requires and what the atmosphere provides result in compromise values of net surface heat fluxes and simulated SSTs in the coupled model that are dependent on both the region and the processes involved in producing those fluxes and SSTs. The combination of these affects the zonal mean ocean heat transport implied in the atmosphere-only simulation and calculated for the ocean-only and coupled models. This analysis is intended to demonstrate the role played by observed estimates of net surface heat fluxes in a coupled modeling strategy that involves not only calibrating coupled model performance, but also improving separate component model simulations.

In some cases, an error from the atmosphere results in coupled model SSTs that are too warm, but with net heat fluxes that remain near the ocean-only values due mainly to increases of latent heat flux from the warmer ocean surface in the coupled model (e.g., near 55\(^\circ\)S). In other regions (e.g., near 20\(^\circ\)N and 20\(^\circ\)S), in response to SST errors in the coupled simulation produced by both the ocean and atmosphere components, precipitation anomalies over the warmer water in the eastern tropical ocean basins produce weaker trade winds, reduced latent heat flux, and near-zero net surface heat flux. Since the ocean-only component required such anomalously small values of net surface heat flux, the result in the coupled model is very good agreement with observed SSTs near 20\(^\circ\)N and 20\(^\circ\)S, in spite of net heat flux errors of about 60%.

In still other regions (e.g., near 50\(^\circ\)N), changes in oceanic circulation (the Gulf Stream and Kuroshio currents and the associated strong SST gradients) due to a spin-down of the subtropical gyres in the coupled model related to weaker-than-observed zonal wind stresses from the atmosphere combine with positive heat flux errors from the atmosphere-only component and cause warmer-than-observed SSTs in the coupled model. The result is an increase of latent heat flux (greater amounts of heat are removed from the ocean surface). Net surface heat flux values in the coupled model end up with an agreement in sign with the observed estimates, but with a reversal in sign from the positive net heat flux values in the atmosphere-only and ocean-only simulations. Finally, near 60\(^\circ\)N (mainly in the North Atlantic), there is qualitative agreement in net surface heat flux among atmosphere-only, ocean-only and coupled models and the observed estimates, yet there is a positive SST error. The northward-flowing currents in the North Atlantic, shifted poleward in the coupled model compared to the ocean-only simulation due to surface wind errors noted above, advect anomalously warm water north of England and Scandinavia, thus resulting in a northward shift of the strong SST gradient in those regions. This raises the values of latent heat flux and makes the net surface heat flux more negative in the coupled model (and in better quantitative agreement with the observed estimate than in either component alone). However, the coupled model is left with a positive SST error in spite of the seemingly good agreement of the net surface heat flux with observed estimates.

It is evident from this analysis that net surface heat flux is a moving target, and it is difficult to predict how it will be simulated before coupling, even after examining the components alone. This is because errors in the components end up affecting the dynamic interaction between the components when coupled, with sometimes unanticipated results both locally and nonlocally. In some cases, the compromises between circulation, heat flux, and SSTs result in errors in fluxes that produce the best SST simulations (e.g., near 20\(^\circ\)N and 20\(^\circ\)S), while in other regions, the errors in fluxes and ocean circulation changes combine to produce large SST errors, but better agreement of the net surface heat flux with observed estimates than in either component alone (e.g., near 50\(^\circ\)N and in the eastern tropical Pacific). In still other regions, agreement between coupled model surface fluxes and observed surface fluxes is associated with large SST errors due to alterations of the ocean circulation in the coupled simulation (e.g., near 60\(^\circ\)N).

One possible way to address this problem is in the spinup of the coupled system. In order to reduce the flux mismatches at the beginning of the coupled simulation, net fluxes from the atmospheric model can be used to spin up the ocean as in the NCAR CSM (Boville and Gent 1997). The ocean will come into agreement with the atmospheric forcing, thus reducing the “coupling shock” on coupling. However, errors in the atmospheric model.
will then introduce comparable errors into the ocean model, building systematic errors into the coupled system. Due to the problems documented in this paper involving flux mismatches, such a spinup technique may be preferable to the alternatives. Yet, no matter what spinup technique is used, the dynamic interactions between the model components of the atmosphere, ocean, and sea ice can produce unanticipated simulation features in the couple model.

Improvements in simulating surface heat fluxes could benefit from better estimates of observed net surface heat fluxes as a point of reference. In particular, when testing model parameterizations in component models run alone, the observed surface fluxes are useful in verifying model performance. However, the problem for coupled models is really one of simulating the state variables representing the thermodynamics and dynamics of the components correctly. As a direct consequence, this kind of holistic model improvement will then naturally result in better-simulated surface fluxes and SSTs in the coupled simulation.

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REFERENCES


