Simulation Errors Associated with the Neglect of Oceanic Salinity in an Atmospheric GCM

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ABSTRACT: In all the atmospheric general circulation models (GCMs) at the Goddard Laboratory for Atmospheres (GLA), the influence of oceanic salinity on the saturation vapor pressure of seawater is ignored. Since the relative humidity in the oceanic boundary layer is generally high while the saturation vapor pressure of seawater is lowered by salinity, its neglect could have a nontrivial influence on the near-surface specific humidity gradient, a primary determinant of oceanic evaporation. Such an approximation might affect the simulated circulation and rainfall systematically. To evaluate this idea, we carried out a 5-yr-long salinity simulation (S) with the GLA GCM in which the influence of salinity on the saturation vapor pressure of seawater was included. Corresponding to this, a control simulation (C) with the GLA GCM in which the salinity effect was ignored was also available. Analyses of S-minus-C fields have shown some evidence of discernible systematic errors in the global evaporation, boundary layer specific humidity, and several key parameters that affect the onset of moist convection, for example, convective available potential energy. Several other systematic effects are also evident even though they remain small compared to the interannual variability of climate. The systematic interactions caused by the neglect of salinity are evidently spurious, and even though the final outcome is less dramatic than

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anticipated originally, several persistent systematic errors can be noted in the 5-yr mean fields. Based on these results, we infer that coupled ocean–atmosphere models that ignore the influence of salinity on ocean evaporation might also benefit from the salinity correction. Indeed, the correction is so trivial to include, its neglect in the modern state-of-the-art GCMs is unwarranted.

KEYWORDS: Modeling; General circulation; Air–sea interaction; Convective processes

1. Introduction

Even though the influence of oceanic salinity on the saturation vapor pressure of seawater is to depress it by roughly 2.0% (the actual value at a place being a function of the magnitude of salinity), climate modelers have chosen to ignore this effect presuming it to be insignificant compared to other parameters that affect air–sea fluxes. On the other hand, so much research and tuning effort has been devoted to better estimate the bulk surface transfer coefficients, $C_D$ and $C_H$. The above assumption is common to almost all atmospheric general circulation models (GCMs) both within the United States and abroad. As soon as we found out about the magnitude of the saturation vapor pressure error caused by the neglect of salinity, we began to question the assumption. We contemplated that if the relative humidity at the anemometer level for ocean-covered regions is taken to be roughly 80% [a statistical number derived from analysis of observations generated by the Data Assimilation Office (DAO) as well as the Goddard Laboratory for Atmospheres (GLA) GCM], a 2% error in the surface vapor pressure could amplify into a fivefold error in the humidity gradient between the surface and the boundary layer aloft leading to a roughly 10% error in ocean evaporation. Consequently, we began to argue that the neglect of the salinity effect might have a systematic and discernible influence on the surface evaporation and consequently the hydrologic processes. However, we also realize that reduced evaporation in an interactive GCM will induce atmospheric adjustments that will alter the course of future air–sea exchanges through spurious interactions; for example, at the very basic level, reduced evaporation will reduce the specific humidity within the boundary layer, which will tend to mitigate the reduction of boundary layer fluxes as well as suppress the moist processes aloft. We further contend that the reduced boundary layer humidity might weaken the water vapor flux transports that in turn can stifle the moisture convergence into raining regions, such as the intertropical convergence zone (ITCZ) over the warm pool region of the ocean. Hence, we argue that salinity might have an important effect on the simulated hydrologic cycle even though its influence on the surface evaporation may, in itself, be somewhat suppressed after the initial adjustment of the boundary layer humidity. Moreover, because the salinity error systematically affects fluxes over all the oceans, it could have nonnegligible consequences for the atmospheric circulation and dynamics.

Based on the above considerations, we have used the current version of the GLA GCM to investigate the influence of salinity on the simulated circulation
and hydrologic processes. Our primary motivation is to determine whether there
is enough justification for ignoring the effect of ocean salinity on the saturation
temperature-pressure of seawater. This investigation should not be viewed as a scientific
investigation of the role of salinity in influencing or maintaining the global cir-
culation, even though there is some element of it in our analysis. We prefer to
postpone addressing that question until a better GCM with cloud water substance,
gravity wave drag, and refined cloud physics replaces the current one. Nonethe-
less, we must emphasize that the current GCM is amply suitable for determining
whether oceanic salinity could (or could not) be ignored in GCMs. We have
compared as well as differentially analyzed two simulations, one with and one
without the salinity effect, to address the issue of oceanic salinity on the air–sea
exchanges and global circulation. Section 2 of this paper describes the details of
the model and simulation experiment; section 3 shows the results with differences
in the hydrologic cycle, and section 4 makes a case for modifying all GCMs to
include the effect of oceanic salinity on the saturation temperature-pressure of seawater.

2. Model and experimental design

2.1. The GCM

The model used for this investigation is the GLA GCM that participated in the
Atmospheric Model Intercomparison Project (AMIP-1). The only subsequent up-
tate to the model is the addition of convective downdrafts by Sud and Walker
(Sud and Walker, 1993). Recently, the model was used in three studies (viz.,
Walker et al., 1995; Sud et al., 1996a; Sud et al., 1996b), aimed at understanding
the influence of changes in the biosphere–atmosphere interaction fluxes, due to
tropical and/or Amazonian deforestation, on the atmospheric circulation and rain-
fall. The key parameters such as solar constant and carbon dioxide concentration
are 1365 W m\(^{-2}\) and 345.0 parts per million (ppm), respectively. These values
were recommended for use in the AMIP project. For further details on the rest
of the disposable parameters, particularly in the simple biosphere model (SiB),
as well as a description of all the physical parameterizations, refer to Phillips et
al. (Phillips et al., 1994) on the World Wide Web at URL: http://
www.pcmdi.llnl.gov/phillips/modldoc/amip/amip.html

2.2. Salinity–vapor pressure relationship

A monthly mean salinity dataset produced by Levitus (Levitus, 1982) was avail-
able at the Laboratory for Hydrospheric Sciences at Goddard Space Flight Center.
The dataset represents the best known climatology of salinity and is in good
agreement with the other salinity datasets. The data show a relatively small tem-
poral variation of salinity [maximum-minus-minimum difference of roughly 0.5
parts per thousand (ppt)] as compared to a much larger spatial variation (maxi-
mum-minus-minimum difference of up to 10.0 ppt). This helped us to conclude
that, as a first approximation, we could use annual mean salinity values shown
in Figure 1a for including the salinity effect in our investigation. This figure shows
a large spatial variation in salinity with a global mean of approximately 35 ppt.
Some regions receive significant amounts of freshwater flux produced by ice melt
and rainfall, for example, the Norwegian Sea; consequently they yield low values of salinity. On the other hand, subtropical regions, where evapotranspiration exceeds precipitation, have relatively higher salinity. Indeed, we can also see sharp salinity gradients near the discharge locations of major rivers, for example, the Amazon, Mississippi, and Congo Rivers. We use the following well-known relations (Witting, 1908) to calculate the reduction of saturation vapor pressure due to salinity:

(a) salinity $S_I$ (ppt) and chlorinity $C_I$ (ppt) relation

$$S_I = 1.805 C_I + 3.0 \times 10^{-2}\quad (1)$$

and

(b) chlorinity (ppt) and saturation vapor pressure factor $F$ relation

$$F = 1.00 - 0.001 C_I.\quad (2)$$

From (1) and (2), we note that for a 32–37-ppt variation of salinity, the computed value of $F$ will be approximately 0.98. Figure 1b shows the field of saturated water vapor pressure depression, that is, $1 - F$, over the oceans.

### 2.3. Simulation experiments

First, we produced a 24-month-long salinity integration $(S)$ using the salinity values from the data shown in Figure 1a. For this period, a control case $(C)$, in
which the salinity effect is ignored, was available to us from our earlier simulations (Walker et al., 1995; Sud et al., 1996a; Sud et al., 1996b). After finding several systematic differences between the S and C simulations, the integration period was extended to 60 months: 0000 UTC 1 January 1979 to 0000 UTC 1 January 1984. Both simulations employed the GLA GCM. An integration period of 60 months enabled us to examine the interannual and intraseasonal differences in the simulations so as to assess the influence of salinity within the framework of full natural variability of climate. Both simulation used the identical SST and sea-ice datasets. The boundary conditions datasets were the same as employed in the AMIP simulation (Gates, 1992) and contain the analyzed interannual variability in SST that were produced by the National Meteorological Center [currently the National Centers for Environmental Prediction (NCEP)] using satellite and in situ observations. The soil moisture initialization assumes that the Mintz and Walker (Mintz and Walker, 1993) analysis yields the volumetric soil moisture for the root zone regions at the initial time. The soil moisture in the GCM is a prognostic variable and it evolves through the soil hydrology submodel of the simple biosphere model (SiB). Nevertheless, accurate soil moisture initialization may not be crucial for the current investigation. However, we would very much like that other modeling groups also perform similar studies to either confirm or invalidate our findings.
3. Results

3.1 Zonal mean fields

At the outset we found that, on the timescale of a season, the local differences in S-minus-C fields at any grid point were so small that they could not be easily discerned from the natural variability of the simulated climate variability. However, the differences in the monthly and 5-yr average global mean evaporation and precipitation did show several systematic differences. Specifically, in the first 2-yr average, during which the tropical SSTs had no ENSO episodes (that dominated the picture in the subsequent 3 years), the global ocean evaporation reduced by roughly 1.6 mm month$^{-1}$, which amounts to 1.6% reduction in evaporation since the simulated ocean evaporation in the control C is roughly 100.0 mm month$^{-1}$. Even this reduction is not stable from one month to the other. Such large fluctuations in the monthly mean fields is consistent with the well-known high variability of the simulated as well as the observed climate; therefore, it is a difficult to isolate the salinity signal that is undoubtedly weaker than the observed interannual variability. It was perhaps for this reason that a few undocumented attempts to determine the oceanic salinity effects in GCMs led to its universal neglect in almost all GCMs throughout the world. The time series of
zonal mean fields also had a number of inter- and intraseasonal fluctuations that showed that there are no large and/or persistent systematic errors caused by neglect of salinity that would be easily reproduced year after year. The small but systematic differences between S and C simulations are well portrayed in the differences in the zonal mean evaporation (Figure 2a), zonal mean precipitation (Figure 2b), zonal mean specific humidity (Figure 2c), zonal mean surface temperature (Figure 2d), and zonal mean convective available potential energy (CAPE, Figure 2e). Since these differences are not only consistent but are easily understandable from salinity influenced biases in sea–air water vapor exchanges; they are also reflected in the zonal mean surface pressures, winds, and other circulation fields (not shown). We note that the CAPE differences in the Tropics (Figure 2e) are not only substantial (up to about 10% of the corresponding values in C simulation) but may in fact be responsible for the substantial S-minus-C differences in the Hadley circulation (section 3.3). Nevertheless, the salinity produced effects are small as compared to interannual variations of climate, but, in subtle ways, the salinity is seen to produce a discernible systematic influence on the simulated ocean surface fluxes. This is not an unexpected result because the salinity error tantamounts to an SST equivalent error of less than 0.4°C (to be distributed according to variation of salinity), whereas the interannual fluctuations

![Figure 2b. Zonal mean S-minus-C differences in the 5-yr means for precipitation (mm month⁻¹).](image-url)
of tropical SST are an order of magnitude larger than the equivalent salinity magnitudes and horizontal gradients.

3.2 Global fields

One may expect to see a well-organized field of a systematic reduction in evaporation, precipitation, and even boundary layer humidity over the entire ocean; however, not only did it not happen but it took almost a year of integration to fully reveal the small differences in some of the fields being discussed in the paper. In the first 2-yr mean, for example, S-minus-C differences in the global distribution of precipitation and evaporation were barely discernible. The simulated climate variability is so large (not shown) that one finds it largely obscures the salinity signal. However, some clear as well as expected differences emerge in the 5-yr averages; and a few of those along with other key features of the simulation are discussed below.

3.2.1. Evaporation and precipitation

The evaporation and precipitation difference fields (Figure 3a) over oceans show that oceanic salinity reduces oceanic evaporation in the Tropics by about 5 W m$^{-2}$, while the accompanying changes in precipitation are much larger, particu-
larly in regions of high moisture convergence such as the ITCZ and the warm pool region of the western tropical Pacific. A system that does not produce any feedback other than altering the evaporation would have reduced the evaporation by about 10% (as argued in the introduction), but to our amazement, the simulated reduction in evaporation is less than 2%. As we shall see later, the reduced evaporation recovers significantly after the atmospheric humidity fields adjust to the new lower value of saturation vapor pressure of the saline oceans. Since evaporation provides the moisture that is required for condensation heating of the atmosphere that is needed to mitigate atmospheric cooling by outgoing longwave radiation, the atmosphere adjusts PBL temperature and water vapor profiles slightly but systematically to gain back the reduction in evaporation. As expected, the evaporation generally reduces in the 5-yr average over most of the oceanic regions; however, there are some regions of increased evaporation, such as the region between the west coast of Canada to the Bering Sea (east of Siberia), but these reflect the influence of substantially invigorated near surface (PBL) winds (Figure 3b); undoubtedly, this is a systematic effect of changes in circulation that emerge in the 5-yr averages. Regardless, the influence of salinity on ocean evaporation is easy to understand from overall energy budget considerations, whereas its influence on the precipitation is far more complex reflecting an outcome of changes in both circulation and moist processes.
3.2.2. Surface layer and boundary layer specific humidity

As stated earlier, the oceanic evaporation did not reduce as much as expected from basic humidity gradient considerations (see section 1). The calculations provided to us by Sui (see section 3.4) for the Tropical Oceans Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA COARE) region indeed yield a much larger change than that simulated by the GCM; indeed, these accord the expected results. The reason for the discrepancy is that the oceanic evaporation strongly recovers after the boundary layer specific humidity reduces in response to reduced evapotranspiration. However, such a reduction in the boundary layer humidity (Figure 3c) in turn influences the water vapor transport convergence and that does happen as we shall see. Finally, after everything has had a chance to equilibrate, both the simulated boundary layer and the surface layer specific humidities are found to reduce by 0.2–0.4 g kg\(^{-1}\), and this effect is almost uniform over the entire ocean. Indeed, the specific humidity reduces in the entire atmosphere as shown in the zonal averages (Figure 2c). As a consequence of this adjustment, the net change in oceanic evaporation turns out to be much smaller than that inferred from Sui’s noninteractive calculation; nevertheless, the spurious feedbacks generated by neglect of oceanic salinity can not be
viewed as trivial. Indeed, as shown in Figures 2c–e, these changes influence the water vapor, surface temperature, and CAPE and in that way they influence the global hydrologic processes.

3.2.3. Winds and sea level pressure

The difference fields for PBL winds and surface pressure are shown in a composite plot in Figure 3b. The differences in the equatorial regions are small, while the midlatitudes show large changes in subtropical highs. There are some changes even in the polar circulation. Surface and sea level pressure increases over Antarctica cause a weakening of westerlies over the latitude of roaring 60s. We argue this is a response to the changes in the Hadley circulation that would affect the mean meridional transports. Such changes in mean meridional circulation, in turn, lead to readjustment of the magnitude and location of the polar and Ferrell cells. Although the statistical significance of the polar region changes is not analyzed, the fact that salinity can influence global-scale circulation has not come as a surprise because small systematic differences such as introduced in tropical deforestation studies have been found to produce global-scale effects (Sud et al., 1996b). We shall examine these changes further with regard to its effect on mean meridional circulation.

3.3. Hadley circulation changes

As compared to the analysis in Peixoto and Oort (Peixoto and Oort, 1992,159), the GLA GCM simulates a reasonable mean meridional circulation (MMC) in the
Figure 3b. The S-minus-C differences in (i) surface winds (m s⁻¹) and (ii) surface pressure (hPa).

annual mean (Figure 4a) as well as in the winter season (not shown); however, its summer (June–July–August or JJA) MMC (Figure 4b) is much stronger than the observed. This causes too high MMC mass flux in JJA as compared to the analysis of observations. In the past we have argued that this situation relates to the deficiencies in convection. In particular, if the cumulus scheme produces too vigorous a moist convection, it will naturally lead to a strong Hadley circulation. In the first 2 years, the S-minus-C simulation produces a weaker annual mean Hadley circulation particularly near the equator: note the differential sinking motion around the northern side of the equator in Figure 5a. The picture becomes much clearer in the Hadley cell differences for JJA (Figure 5b), which shows a much broader ITCZ with a weaker rising branch at the latitude bands of the equator (0°–15°N) and a stronger rising motion in the subtropics (15°–30°N). Thus, salinity effect not only weakens the tropical ITCZ but promotes its northward excursion in the summer. However, in the subsequent 3 years, the S-minus-C differences of the first 2 years were not well reproduced, Figure 5c. This may not be difficult to understand, if one recognizes that the later 3 years were overwhelmed with El Niño/La Niña episodes, which would have a far stronger influence on modulating the MMC than the salinity. The Hadley cell differences in Figure 5b are consistent with the rainfall reduction over the tropical oceans, which can be seen in the zonal (Figure 2b) as well as the global precipitation differences (Figure 3a). Thus, the changes in the tropical precipitation are not too large, but
when viewed together with the systematic errors in the other zonal average quantities, their relationship with the salinity effect begins to emerge. Even though the influence of salinity in affecting the MMC is strongly masked by interannual variability of climate, it is quite evident in the internal consistency of various difference fields discussed above as well as the physical basis of some of the spurious errors.

3.4. Evaporation and precipitation in the warm pool region

To compare the differences between the ocean evaporation with and without the salinity correction, we focused on the warm pool region (also called TOGA COARE region: $10^\circ$S–$10^\circ$N and $140^\circ$E–$180^\circ$) of the western Pacific. We requested C.-H. Sui of the Climate and Radiation Branch to produce two calculations of the surface evaporation fields: one with and one without the salinity correction for the TOGA COARE campaign period using the TOGA COARE surface evaporation scheme. The results are shown in Figure 6a. In these graphs, the evaporation with salinity corrected sea surface vapor pressure reduces by 9.95 W m$^{-2}$; the total evaporation is 105.05 (115.00) W m$^{-2}$ for the cases with (without) salinity correction in which a salinity factor $F$ equal to 0.98 is used. We note that this represents roughly a 10% decrease in evaporation in response to only a 2% decrease in saturation vapor pressure. This result is so akin to that expected from simple arguments. However, such a large decrease in evaporation did not occur in the GCM simulation as is evident in S and C comparisons of Figure 6b. We
Figure 4. (a) Time mean meridional circulation (MMC) for the control case for the 5-yr mean in $10^9$ kg s$^{-1}$. (b) Same as a except for JJA MMC for the 5-yr mean in $10^9$ kg s$^{-1}$.

have shown that this happens because the boundary layer humidity reduces in response to reduced evaporation, which in turn mitigates the reduction in evaporation. This is also confirmed by the fact that the reduction in evaporation in the first time step of the GCM was about 7%; however, at the end of the first day it became a little more than 4%, and in the long-term climatology, finally it settled at a value of only 2%. However, because the overall effect of this correction was to reduce the evaporation over the entire ocean, it leads to lower specific and relative humidity of the boundary layer as well as reduced moisture convergence into intensely raining TOGA COARE region, particularly in the first 2 years. Figure 6b also shows that the local precipitation also reduces as a consequence of reduced moisture flux convergence. In the first 2 years, it reduced by
24.5 mm m$^{-1}$ from a value of 263.5 mm month$^{-1}$ for the control run. The corresponding reduction in local evaporation was only 1.2 W m$^{-2}$, (roughly equal to 1.2 mm month$^{-1}$) from a value of 136.8 W m$^{-2}$ for the control case. In this way, the reduced specific humidity in the boundary layer reduces the specific humidity of the wind transports into the primary moisture convergence regions and thereby reduces the rainfall. The same was evident in the rainfall differences in the ITCZ regions, which show that the reduction in rainfall is much larger than the reduction in local evaporation (not shown). However, a much more complex scenario emerged in the next 3 years suggesting that the influence of 2$^\circ$–4$^\circ$ changes in SSTs is much stronger than the systematic effect of salinity that translates into an effective SST change of less than 0.4$^\circ$C spread more or less uniformly over the entire ocean (Figure 1a). It is not clear whether the salinity effect can

Figure 5. (a) First 2-yr time average S-minus-C differences in MMC in 10$^9$ kg s$^{-1}$. (b) First 2-yr JJA average S-minus-C differences in MMC in 10$^9$ kg s$^{-1}$. 

Pressure (hPa)

Latitude

Pressure (hPa)
make a difference for the simulation deficiencies of the GLA GCM that we inferred initially from the first 2 years of the integration, but that it introduces spurious systematic errors is abundantly clear. A small but corrective reduction of the summer Hadley circulation in the first 2 years of simulation was completely wiped out in the subsequent 3 years. Nevertheless, the differences between these simulations are large in the Tropics, and that suggests that the simulated circulation, particularly tropical, shows discernible sensitivity to oceanic salinity. This provides sufficient grounds for including the correction.

4. Summary and conclusions

Our simulations with the GLA GCM have shown that oceanic salinity has a small but systematic influence on oceanic evaporation, oceanic boundary layer humidity, mean meridional circulation, and hydrologic cycle including the ITCZ rainfall. Consequently, its effect should not be ignored even if the evaporation differences it causes turn out to be small. Sui’s calculations clearly show that a 2% error in the saturation vapor pressure caused by neglect of salinity can lead to as much as 10% error in evaporation. However, when the atmospheric interactions evolve and adjust in response to the error due to neglect of salinity, the GCM simulation shows a mere 2% error in evaporation; evidently, this hides more complex and essentially spurious physical adjustment processes that partake to reduce the overall error in the evaporation. These adjustments lead to reduced PBL water vapor and moist static energy, and air-column CAPE, and even the surface temperature. All of these are important for the physical processes and the evolution of the earth–atmosphere system. Thus, the ensuing adjustments make the task of detecting the errors rather difficult and might have led modelers to ignore this effect. We believe that the only logical way to develop a complex model, such as the
Figure 6a. Evaporation (W m$^{-2}$) for the TOGA COARE site during the field campaign period calculated from the TOGA COARE algorithm (top) with and without the salinity correction factor.

earth–atmosphere system model, is to include all the key effects, however trivial, and then ignore them only after clearly delineating the sensitivity of the system to the specific assumptions. This lesson has surfaced in the past when modelers made unwarranted assumption that later came to haunt them, for example, ignoring momentum and moisture mixing in dry convection (Sud and Molod, 1988), and ignoring the saturation water vapor pressure for ice-phase below subfreezing temperatures (Starr et al., 1995). The salinity correction is so trivial to invoke that its neglect in GCMs must be corrected. In future simulation studies we will attempt to address issues such as the influence of salinity on maintaining the global circulation and hydrologic cycle and prediction of climate scenarios. These questions require state-of-the-art GCMs as well as a participation by several modeling groups because the magnitude of salinity correction for most regions is quite small. Consequently, we urge that we have made a strong case for including the influence of salinity on the saturation vapor pressure over seawater. Nevertheless,
the correction of salinity error is not a substitute for several major systematic errors due to deficiencies in the physical parameterizations, such as clouds, which need substantial research and development.

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