The Effect of Salinity on Tropical Ocean Models

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ABSTRACT

The effect of horizontal salinity gradients on the tropical ocean circulation has not previously been evaluated. It is shown that there are noticeable differences between the dynamic height field calculated with and without the inclusion of salinity variations. Hence salinity has a significant contribution to the geostrophic velocity field. This conclusion is illustrated by running two identical Indian Ocean models: one initialized using a climatological salinity field while the other has no horizontal salinity gradients. The differences in the temperature and velocity fields after 110 days are of the order of 0.5°C and 10 cm s^{-1} over some regions of the ocean. Further experiments using the same model for data updating studies showed that the absence of salinity data greatly reduces the usefulness of temperature data. It is concluded that for an accurate simulation of the tropical ocean the salinity field needs to be included.

1. Introduction

The ocean has two parameters that determine the density at a given pressure–temperature and salinity. It is known that both of these have important effects in determining the ocean circulation and thereby the Earth's total climate system. The ocean has a dominant role in the meridional redistribution of thermal energy (Ellis et al. 1978), and the sea surface temperature has important effects on the atmospheric circulation. This is seen in the tropics in the El Niño–Southern Oscillation (ENSO) phenomenon (Rasmusson and Carpenter 1982; Gill and Rasmusson 1983) and also at midlatitudes (Palmer and Zhaobo 1986). Although salinity has little direct effect upon the atmosphere, it is important in the formation of water masses and thus in the global ocean circulation (Levitus 1986).

While salinity is known to be the dominant cause of ocean currents at high latitudes, by forcing the thermohaline circulation (Bryan 1986), its importance in tropical dynamics is not known. In tropical ocean models the density is calculated either with salinity fixed (Latif et al. 1985; Rowe and Wells 1985) or else with salinity advected and diffused from a prescribed initial state (Philander and Siegel 1985). It is only recently that a tropical ocean model has included fully active salinity with surface water flux (C. Gordon, personal communication).

Why is it that salinity has been neglected in tropical ocean models? One reason is the poor data coverage of both salinity within the ocean and the water mass flux across the ocean surface. In particular both evaporation (Esbensen and Kushnir 1981; Reed 1985) and precipitation (Jaeger 1976) over the tropical ocean are very poorly known. This lack of data means that the surface water flux is a subjective quantity and hence the forcing of the surface salinity field is unknown. Even so, it would be expected that the data we have are better than none at all. Indeed, while the heat flux over the ocean is poorly known, with variations between different estimates as large as a factor of two (compare Esbensen and Kushnir 1981, with Reed 1985), the data are still used in ocean modeling.

A second reason why salinity has been ignored in tropical ocean models is the assumption that it has negligible effect on the dynamics. For while at mid- and high-latitude ocean temperatures the thermal expansion coefficient of sea water is small, at tropical temperatures it becomes much larger (for instance, see Gill 1982, appendix A). In contrast the salinity "expansion" coefficient is almost constant for all the ocean. Hence the same temperature change will have a larger effect on the density in the tropics than in colder water, leading to the conclusion that the effect of salinity variations in the tropics is negligible. This neglect is encouraged by the lack of a direct effect of salinity on the atmosphere. However, the above considerations should not lead us to neglecting salinity but just to realizing that temperature is more important than salinity in the tropics.

To quantify the importance of salinity in the tropics, its contribution to the dynamic height was calculated (section 2). Experiments were then performed in a global circulation model (GCM) of the Indian Ocean to compare a model initialized with and without horizontal salinity gradients (section 3). The results show important quantitative differences, though the ocean

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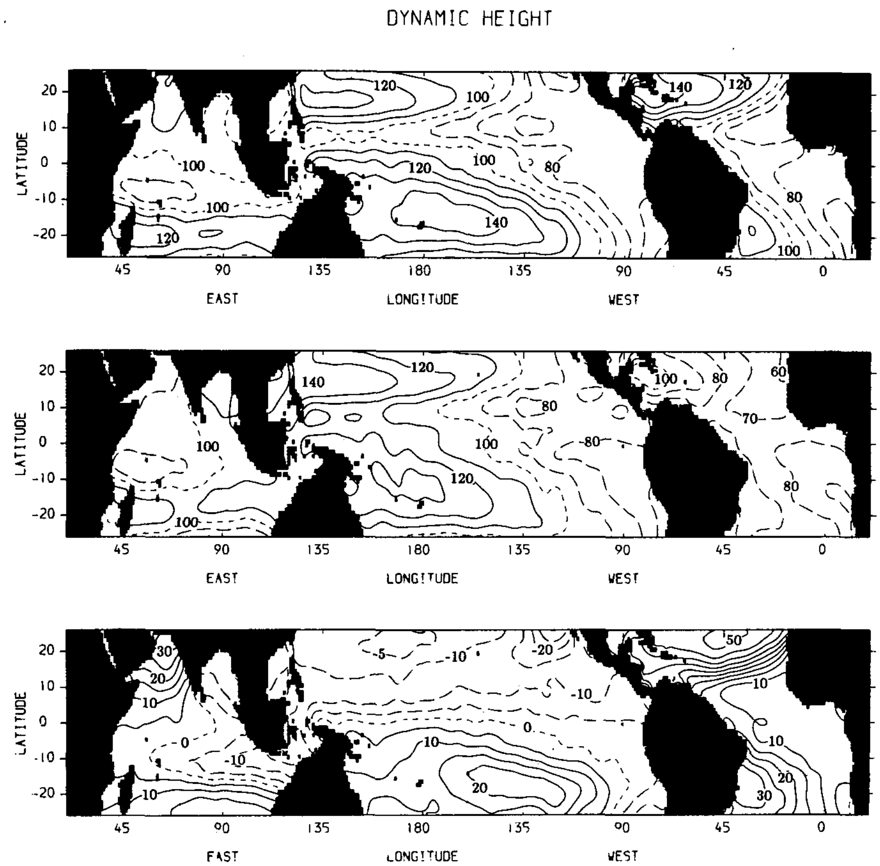


FIG. 1. The dynamic height of the tropical ocean surface relative to 400 m depth calculated from Levitus data. (a) Salinity set to a constant value of 35, (b) salinity varying as in Levitus data and (c) difference between (a) and (b). Contour intervals are 10 dyn cm in (a) and (b) and 5 dyn cm in (c).

fields are still similar. To investigate the effect of neglecting salinity on models which include observational data, data assimilation studies were performed using the method of Moore et al. (1987). The results are presented in section 4, and show that lack of salinity data can lead to severe degradation of the model when run in an "operational" mode. Suggestions of how to overcome the problem of our lack of salinity data are given in the conclusions, section 5.

2. Effect of salinity on the dynamic height

The surface salinity field in the tropical ocean shows some large variations in space and time (see Figs. 5 and 8 in Levitus 1986). The strongest salinity gradients, both horizontal and vertical, are in the Bay of Bengal and near the outflows from the Amazon and Congo rivers. The annual signal is large in these regions and also near the tropical convergence zones. The annual cycle has a maximum amplitude of 8 (in the SI-recommended units, which are pure numbers and do not have the attached units of parts per thousand) while the large scale field can have gradients of over 1 in 1000 km. The cause of these variations is the influx of vast amounts of rain and river water. Thus it might be thought that the salinity perturbations are significant only close to the surface. However, the horizontal salinity gradients are large down to below 400 meters (see Fig. 11 of Levitus 1986). Because evaporation from the surface of the ocean is much more spatially and temporally uniform than the water flux into the ocean, it has less effect on the horizontal gradients in salinity, except at the very longest length scales. However, the rainfall over the ocean is very poorly known and hence any operational model will have difficulty in predicting the correct salinity field in the upper 100 m.

Temperature, salinity, and pressure determine the density of sea water, which can be calculated using one of many formulae (Lewis and Perkin 1981). Thus at a given pressure, and hence depth, the density depends only on temperature and salinity and therefore the horizontal density gradient is due to gradients in temperature and salinity. At high latitudes salinity is the dominant component of the density gradient, but in the tropics the coefficient of thermal expansion is much

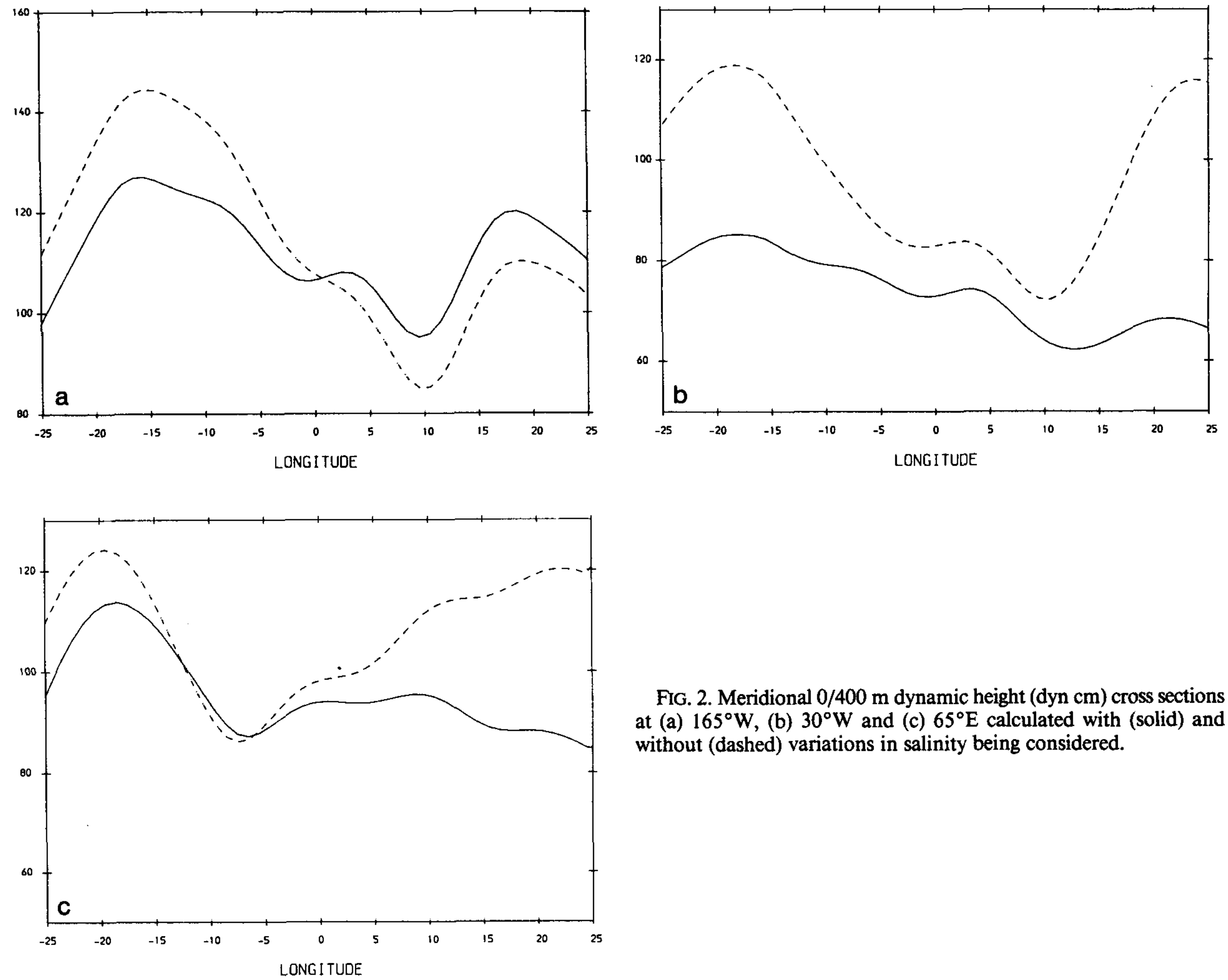


FIG. 2. Meridional 0/400 m dynamic height (dyn cm) cross sections at (a) 165°W, (b) 30°W and (c) 65°E calculated with (solid) and without (dashed) variations in salinity being considered.

higher and the density is more sensitive to temperature than salinity variations. To illustrate the relative magnitudes of the effect of temperature and salinity gradients in the tropics, the equation of state may be approximated to:

$$\rho = \rho_0[1 - \alpha(T - T_0) + \beta(S - S_0)]$$

where $\alpha = 0.304$ and $\beta = 0.756$. (Values based on Table A3.1 in Gill 1982 with $T_0 = 25°C$ and $S_0 = 35$.) Thus in the tropics the density gradient due to a unit change in salinity is equivalent to a change in temperature of about 2.5°C.

Salinity affects ocean dynamics through the pressure gradient terms in the momentum equations. Using annual mean data from Levitus (1982) the 0/400 m dynamic height was calculated throughout the tropical upper ocean using the density formula A3.2 from Gill (1982), in two cases. In the first case the full information about both temperature and salinity was used, while in the second the salinity was assumed to be constant at 35. The dynamic height calculated indicates the sea level variations which would be observed assuming that 400 m is a level of no motion. It is evaluated from the density profile $\rho(x, y, z)$ by the formula

$$h(x, y) = g \int_{-400}^{0} \left(1 - \frac{\rho(x, y, z)}{\rho_0}\right) dz$$

where ρ_0 is the density of sea water at temperature 0°C and salinity 35, x is longitude, y latitude and z depth.

Figure 1 shows three plots of 0/400 m dynamic height field for the tropical oceans, the upper two calculated without and with salinity variations, respectively, while the third is the difference between the two fields and hence shows the effect of salinity gradients. It can be seen from Fig. 1a that the dynamic height reflects the ocean thermal field; for instance, the cold ocean temperature in the eastern Pacific showing as lower dynamic heights. Figure 1b, in which the effect of variations in salinity are included, shows similar large scale patterns but also many differences. The meridional variations are less than in Fig. 1a, the largest values in the Pacific are now north of the equator and the

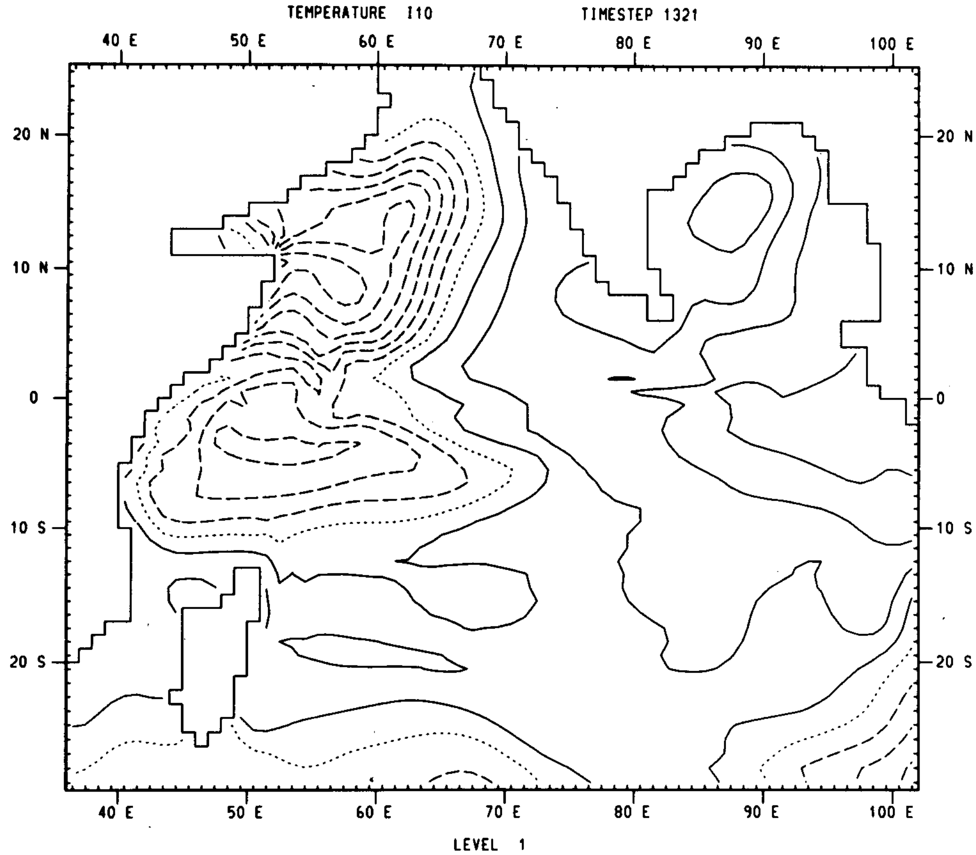


FIG. 3. (a) Temperature field in the ocean model I10 (with full salinity) at the end of September. The contour interval is 1°C, with solid contours for $T > 25°C$, dashed for $T < 25°C$ and dotted where $T = 25°C$. (b) Salinity field with c.i. of 0.5, solid contours for $S > 35$, dashed for $S < 35$ and dotted for $S = 35$. (c) Velocity field with c.i. of 1 m s^{-1}.

Atlantic shows much smaller gradients. These differences are highlighted in Fig. 1c, which, as would be expected, shows a striking similarity to the surface salinity climatology (Fig. 5 of Levitus 1986). There are large positive anomalies in the dynamic height field where surface salinity is above 36 and the saline Atlantic Ocean has only positive anomalies, with very steep meridional gradients north of the equator.

Comparison of Figs. 1a and 1c shows that in much of the tropical ocean the meridional variations in temperature and salinity are positive correlated, with the ocean tending to be more saline where it is warmer. The consequence of this (with the equation of state) is that they have opposite effects on the dynamic height, and thus including salinity reduces the meridional pressure gradients. This effect is clearly seen in Fig. 2 which shows three north–south cross sections, one in each ocean basin. In each case the meridional gradient is less when salinity included. In the Pacific (Fig. 2a) the effect is smallest. Even so, in the region between the equator and 15°S the gradient is halved, and thus so is the zonal geostrophic current [which has value $u_g = -(g/f)dh/dy$ where g is gravity and f is the Coriolis parameter]. At 10°S, $u_g = 5$ cm s^{-1} when salinity is neglected but is only 2 cm s^{-1} when salinity is included.

In the Atlantic Ocean there is a large temperature gradient between 10° and 20°N, clearly visible on Fig. 1a. However, a complementary gradient exists in the salinity (Fig. 1c) and consequently there is very little gradient in the dynamic height when both temperature and salinity are considered. This is illustrated in the meridional cross section, Fig. 2b. The geostrophic currents at 15°N and at 10°S are each about 10 cm s^{-1} when only temperature is considered, but the value is reduced to 2–3 cm s^{-1} when salinity is also taken into account. Similar effects also occur in the Indian Ocean north of the equator (Fig. 2c), though in this case the salinity gradient is so steep as to reverse the geostrophic flow from 2 cm s^{-1} to -3 cm s^{-1} at 15°N.

From these results it is clear that salinity counteracts thermal effects on the density, thus reducing density gradients and making isopycnals more horizontal. In some areas the effect of salinity is larger than that of temperature and hence must not be neglected. Climatological salinity, however, has very little effect on the zonal pressure gradient in the equatorial Pacific, and

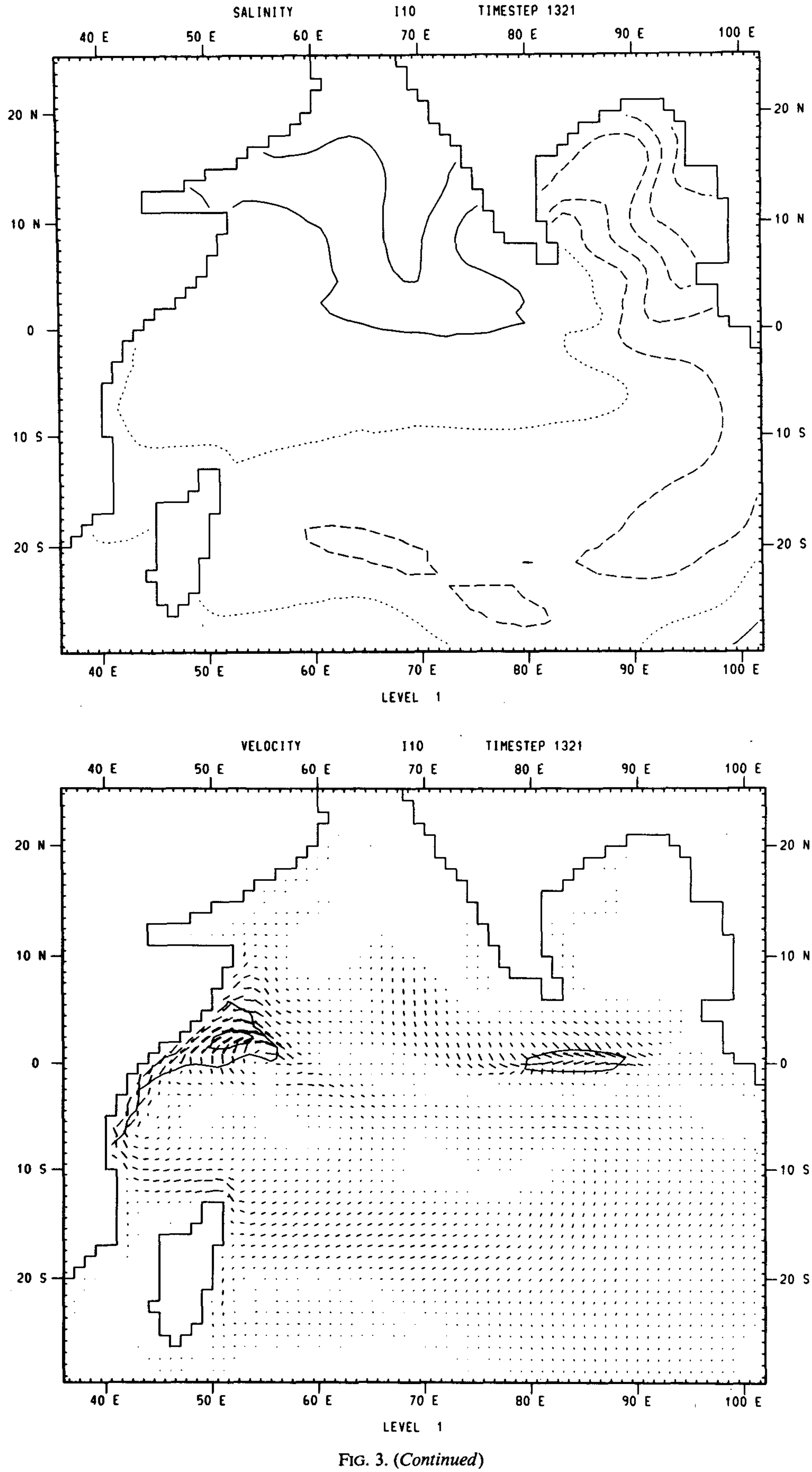


FIG. 3. (*Continued*)

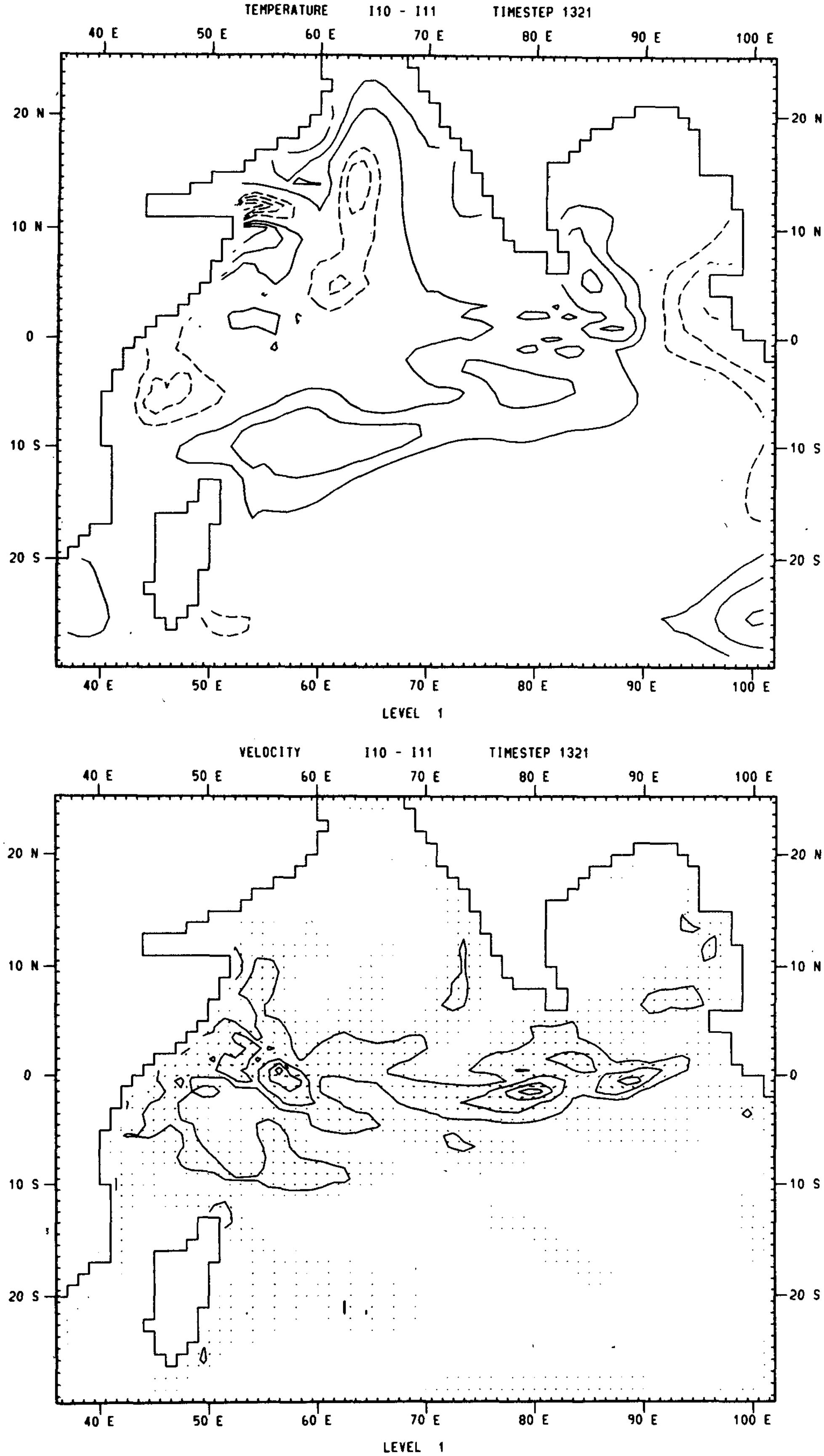


FIG. 4. (a) Temperature and (b) velocity difference fields between runs I10 and I11, thus showing the effect of ignoring salinity. Contour intervals are 0.5°C and 10 cm s^{-1}. For clarity the zero temperature contour is not plotted. Solid contours are for positive differences and dashed contours for negative differences.

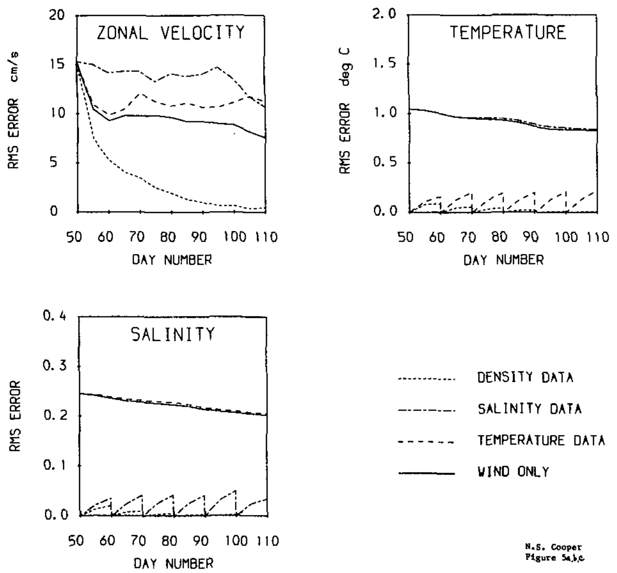


FIG. 5. The rms errors in (a) the velocity field, (b) the temperature field and (c) the salinity field for the four updating experiments listed. Note that providing just temperature or just salinity data produces larger errors than using no data from the truth run.

thus the equatorial undercurrent may be insensitive to the annual mean salinity distribution.

The plots shown (Figs. 1 and 2) are for the annual mean salinity and temperature. Further calculations (not presented here) show that the seasonal variations in the pressure gradient also have a significant (in some places dominant) contribution from salinity. Similarly, when there is an anomalous rainfall pattern the pressure gradients will be altered due to the induced salinity gradients. For example, the rainfall pattern of El Niño alters the near-surface salinity (Donguy and Eldin 1985) and hence the dynamic height may be changed more by this effect than by ocean warming (Donguy et al. 1986). Both simple El Niño models and global circulation models have not included the effect of precipitation on the density, yet it may significantly change the ocean dynamics.

Thus it has been shown that salinity is important in the tropics due to its large contribution to the horizontal pressure gradients. However, in the upper ocean, where this effect is largest, the circulation is predominantly wind driven hence in the following sections the effect on a tropical Ocean Global Circulation Model (OGCM) of neglecting salinity is investigated.

3. Effect of initial salinity on an ocean model

The model used was that of Bryan (1969) in the Semtner (1974) coding. The coastal geometry used was that of the Indian Ocean with a 1° resolution, while the bottom was assumed flat at a depth of 4.59 km. There were six vertical levels at depths of 10, 30, 75, 150, 300 and 2500 m. To increase the computational speed, the barotropic component (which is small in the tropics) was neglected. The vertical mixing was calculated using the Richardson number parameterisation of Pacanowski and Philander (1981), while the horizontal mixing was a Laplacian operator with coefficients of 1×10^8 m^2 s^{-1} (viscosity) and 5×10^7 m^2 s^{-1} (diffusivity). The wind forcing is based on the 12-hourly data supplied by J. J. O'Brien (personal communication) for July to October 1979, but averaged over 5-day periods to remove high frequency forcing. The timestep was 2 hours with an Euler forward timestep every 17 timesteps. There was no surface heat or fresh water (salinity) flux.

Results from two model integrations, both of which started from a state of rest, will be described in this section. In one case both the initial temperature and salinity fields had the full 3-dimensional structure of the Levitus (1982) annual mean climatology while in the other integration the initial salinity was horizontally uniform, being assigned the mean value at that level for the whole basin. For ease of reference the integrations are called I10 and I11, respectively. Values of temperature and salinity in the model ocean below the actual ocean floor were interpolated from the nearest ocean grid points at the same depth. Each integration was run for 110 days.

We are interested in the difference between the two runs, but first we confirm that the model is reasonable. The main phenomena in the Indian Ocean in June to September is the Somali Current, forced by the southeast Monsoon winds. This current, with its associated upwelling, is reproduced in the model (see Fig. 3) with a maximum velocity of over 2 m s^{-1} compared to the observed 2.5 m s^{-1} (Swallow et al. 1983). However, the model Somali current is too wide due to the low model resolution and also the temperature anomaly of the upwelling is too extensive due to both the poor resolution and the neglect of surface heat flux. The other surface currents are reproduced in response to the wind forcing. The qualitative results from both models are similar and the inclusion of salinity has little effect on the accuracy of the model, as it does not mitigate the other errors which are present.

When the two integrations are compared with each other then significant differences are found, as is illustrated in Fig. 4. Despite the integrations beginning with identical thermal distributions, there are differences of up to 2°C after 110 days. The largest differences are localized in the Somali current region where large velocity differences (due to differences in the density field) result in temperature advection differences. Much of the ocean has a difference in surface temperature between the two integrations of over 0.5°C. The cause of this may be the differences in the vertical mixing in the two integrations as salinity affects the vertical stability. Differences in surface velocity of up to 40 cm s^{-1} occur on the equator, and differences of over 10 cm s^{-1} occur between 10°N and 10°S. This is because it is in this region that the geostrophic current which,

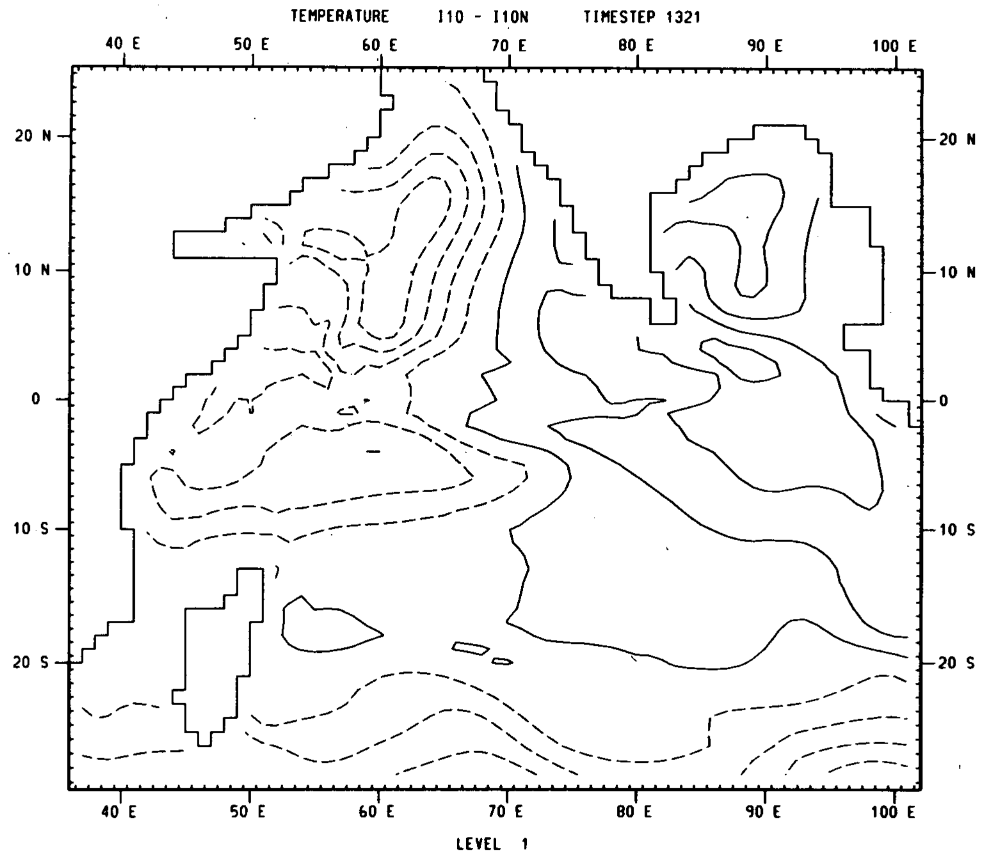


FIG. 6. Temperature difference fields at the end of the integration for three different updating experiments (a) no data updated (b) temperature data updated and (c) salinity data updated. Solid contours are for $T = 0.5°C$, $1.5°C$ · · ·, dashed lines are for $T = -0.5°C$, $-1.5°C$ · · ·. Note that as updating with density data produces negligible errors the diagrams are blank and thus not plotted.

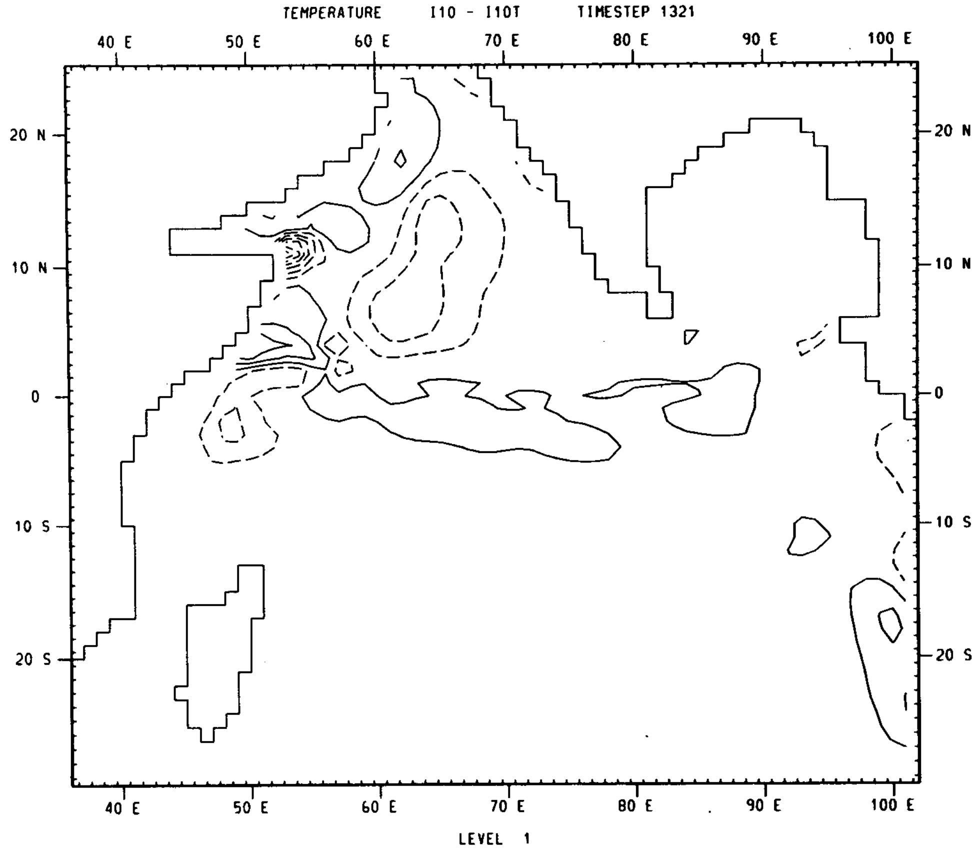
as can be seen from Fig. 2c, is completely altered by neglecting salinity.

Thus we see that the neglect of salinity causes errors in both the velocity and temperature fields. For theoretical ocean models used for research, such errors are insignificant. However, ocean models are now beginning to be used to produce high quality data. This occurs in both coupled models and prognostic ocean models. In the former, an ocean GCM is coupled to an atmospheric GCM for long integrations, with the fluxes (of momentum, heat and water) at the ocean–atmosphere interface being predicted by the physics of the models. Prognostic ocean models are being developed in which observed, real-time data is assimilated into an ocean model to produce a forecast in the same way as in numerical weather prediction. For both coupled and prognostic models the errors found in this section would be serious.

4. Effect of salinity in data updating experiments

In reality, when running a prognostic ocean model the data coverage is very limited. In this section we investigate the situation when data coverage is incomplete using an idealized scheme in which it is assumed that the velocity, temperature, and salinity fields are each either known at every grid point or at none. We then investigate which data are most useful in reproducing the "truth" integration. The method used follows that of Moore et al. (1987, hereafter referred to as MCA) where the relative usefulness of velocity and temperature data was considered with salinity kept constant. It was found there that density data is much more important than velocity data in reconstructing the model fields.

The model was integrated for 110 days (the truth run) and the fields (u, v, S, T) stored every 5 days. This data is then taken to represent reality, and is referred to as the truth field. Further integrations are then performed with the same model but with limited data from the truth run (the "identical twin" method). The updating runs are started from rest at day 50 of the truth run with the constant temperature and salinity profiles, defined as the mean at each level of the Levitus data for the whole basin. Every 10 days (beginning at day 50) in the updating run the model is updated by replacing some (or no) fields with the observed field from the truth run. The updating at all grid points every 10



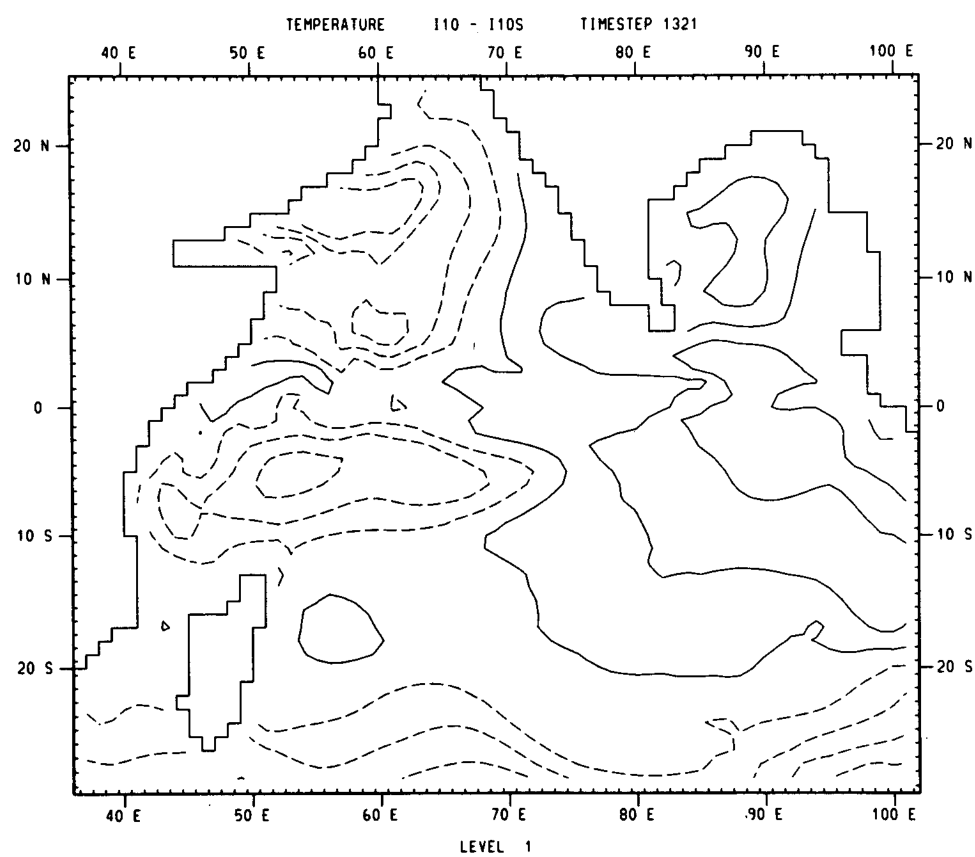


FIG. 6. (*Continued*)

days is obviously an oversimplification of the situation which will be encountered in reality when using observed data, for oceanic data coverage is very sparse and aperiodic. Using this assumption, however, we can investigate how important the different ocean fields are and hence evaluate what types of observation are necessary for running a prognostic Indian Ocean model. Various methods exist for assimilating sparse data (Lorenc 1986) and these will be the subject of future analysis.

Previous work (MCA, Philander et al. 1987) has shown that the density field is much more important than the velocity field because the potential energy is much greater than the kinetic energy. Given the density field, the velocity field is reconstructed within a couple of days by the Rossby adjustment processes. Conversely, velocity data does not produce the appropriate density field, or even density gradient field, because of the smaller amount of energy supplied by the data. Experiments with the model described above confirmed that the same result holds here: density data, comprising both temperature and salinity, will cause the model to reproduce the truth very accurately within 60 days of integration; in contrast, using velocity data gives only a slight improvement compared to having no oceanic data whatsoever. [The rms temperature and salinity fields are calculated with a regional mean removed (see MCA for details) as while density data gives information about the velocity field, the velocity field can only provide information about the density gradients.]

The question of importance here is: Given that salinity has an important contribution to the density field, what happens if we only have temperature data? To answer this question, two further runs were made: one in which only temperature data was updated and the other in which only salinity data was updated. The results are shown in Fig. 5. The surprising result is that providing only one component of density, either temperature or salinity, results in the model being degraded with the rms errors in the velocity field being larger than if no data at all were supplied. This is because the pressure gradients due to salinity or temperature variations alone are greater than their sum, for as shown in section 2 they counteract each other. The morphology of the error field can be seen in Fig. 6, which shows horizontal cross sections of the uppermost model level for the various experiments performed. As in section 3, where the model had the wrong initial salinity field, supplying only temperature data results in errors in the temperature field over much of the basin. Repeated insertion of temperature data reduces the errors over much of the basin, but in the Somali Current region the error is enhanced as large velocity errors are introduced. One cause of this error is the "shock" of data updating for, due to the velocity and density fields no longer balancing, inertial oscillations are produced. While these oscillations decay within a couple of days, they will result in more long term effects particularly where they have large amplitudes. Also, the production of temperature and salinity fields from different sources can result in density induced instabilities (M. MacVean, personal communication). Another cause of these errors is due to the large scale temperature and salinity fields producing opposite pressure (and density) gradients, which can be seen in Figs. 1 and 2.

Hence we see that the lack of complete density data results in the failure to reproduce the "real ocean" accurately. This problem will obviously occur in real time models due to the poor salinity data coverage at present available.

Two further data updating experiments were performed in which, as well as temperature data, surface salinity data was also used. The basis for these updating experiments was that when an XBT sounding is made a surface water sample is also taken, and hence surface salinity data is simultaneously available. In one experiment only the surface salinity value was updated. In the other case, at depth levels 2 and 3 the salinity field was also changed, by an amount proportional to the surface anomaly. In both cases there was no significant reduction in the rms velocity and temperature errors. One further technique could be used, that of Kessler and Taft (1987), in which a local T–S relationship is used below the mixed layer and the surface salinity above. However, this method relies on there having been salinity measurements made in the vicinity at some previous time and this is only true for a small proportion of the tropical ocean. Hence to be able to use this method many more salinity profile measurements need to be made. Operationally this method would also have problems as at present the salinity sample is not analysed until after the ship has reached its destination and thus is not available in real time.

5. Conclusions

It has been shown that salinity gradients have a significant effect on the tropical dynamic height field, and thus to the pressure gradients. Thus, salinity cannot be neglected in tropical ocean modeling when an accurate ocean surface temperature field is required. Therefore both coupled ocean–atmosphere and prognostic ocean models should include salinity actively, although this will require estimation of the surface fresh water flux. Failure to include salinity gradients results in surface temperature differences of up to 2°C and surface velocity differences of up to 40 cm s^{-1}. The temperature differences are due both to the advection of heat being altered by velocity differences and to the vertical stability being changed. The effect of salinity is very important in the determination of the density field and hence also needs to be included in the calculation of dynamic heights. For prognostic ocean models, lack of concurrent salinity data results in temperature data being of little use. This last conclusion will hold in the Indian and Atlantic oceans but not necessarily in the Pacific where, as seen in Figs. 1 and 2, salinity is less important in the climatological dynamic height field.

Having shown that salinity is vital for an accurate reproduction of the tropical ocean circulation, we are faced with the awkward problem of "How do we get the data?". Unfortunately atmospheric models are at present unable to determine the surface fluxes with any accuracy. Also, these fluxes cause the upper ocean to not have a simple *T–S* relationship. Hence we are forced to rely on in situ measurements by conductivity–temperature devices (CTDs) which at present are very sparse. Therefore, we urge the development of a simple technique by which salinity (or any related property such as conductivity or density) could be measured relatively inexpensively and easily by ships of opportunity. Without such data our models will always have an Achilles' heel.

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