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An equatorial transect of the Indian Ocean

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ABSTRACT

A hydrographic survey made from 95E to 50°30'E between 1N and 1S in December 1976-January 1977 indicates zonal pressure gradients as high as $10^{-4}$ dynes/g (pressure increasing to the east) in the near-surface. These are the strongest large-scale zonal pressure gradients measured in any equatorial ocean to date. A large volume of anomalously saline water was found along the equator east of the Maldives Islands. Equatorial jets between monsoons apparently moved this water east after it reached the equator south of its source region in the Arabian Sea. Fluctuations of roughly 2 dyn. cm. in dynamic height with zonal wavelength of 1200-1500 km. are present below the thermocline; these are possibly related to short vertical scale currents measured by Luyten and Swallow (1976).

1. Introduction

The equatorial region of the Indian Ocean is comparatively unexplored primarily because of logistic difficulties in carrying out observational programs there. Yet wind forcing by the monsoons creates large fluctuations in currents and hydrography. We report here the results of a modest survey across the ocean from 95E (near Sumatra) to 50°30'E (near Africa) made within 1° of the equator taken from 16 December 1976 to 3 January 1977. The only other equatorial transects reported were made by Taft and Knauss (1967) in July, 1962 (during the southwest monsoon) and March-April, 1963 (toward the end of the northeast monsoon). Our measurements indicate very different large scale dynamic topography of the upper ocean than those of Taft and Knauss. Taken together these transects give a rough
picture of seasonal pressure field fluctuations along the Indian Ocean equator, assuming that the seasonal changes are stronger than interannual ones.
2. Observational program

The survey was done largely as a ship-of-opportunity observational program. The R.V. *Atlantis II* of the Woods Hole Oceanographic Institution sailed from Singapore to Goa, India and recovered an array of current meter moorings located on the equator centered at 53E. Constraints of time and ship fuel capacity limited the experiment to a series of CTD (conductivity-temperature-depth) casts taken to nominally 1000 m depth followed by current profiles to roughly 500 m at sites along the equator and at 1N and 1S roughly every 7° of longitude. Between each short meridional section a station was occupied on the equator giving an equatorial spacing of roughly 400 km. The station sites are shown in Figure 1.

The CTD instrument used was a Plessey model; the system configuration is described by Scarlet (1975). On each cast at least three sea-water samples were taken with Niskin bottles mounted on a rosette sampler. These samples were used to calibrate salinity measured with the CTD. Duplicate sets of samples were analyzed at W.H.O.I. after shipment: one set travelled by air freight, the other remained on the ship until it arrived in Woods Hole five months after the cruise. Samples contaminated by leakage are clearly distinguishable from others; excluding these contaminated samples, about two thirds were within 0.01‰ of the calculated CTD salinity and the rest were within 0.02‰. A single calibration curve was used to correct CTD salinity for all casts. Instrument drift tending to give a systematic error is less than 0.01‰ over all the casts.

The CTD uses a modified conductivity cell to give digitized salinity which is processed by the system described by Scarlet (1975). A shroud over the thermistor was used for all casts and correcting for a time constant of 0.5 seconds seemed to most consistently reduce sharp salinity fluctuations in regions of high temperature gradient. Although slowing the CTD lowering rate during passage through large gradient regions allows better vertical resolution because of the slow time constant, this increase in resolution may be offset by less effective flushing of the thermistor shroud at slower lowering rates, tending to decrease vertical resolution. For certain casts large spikes to low salinity at the base of a very sharp thermocline (10°C/10 m) were removed manually in the processing in a manner consistent with historical T-S characteristics (using Wyrtki, 1971).

The current profiles were made with instrumentation developed by Düing and Johnson (1972). The profiler consists of a ballasted housing carrying an Aanderaa current meter which descends along the hydrographic wire at a speed determined by its buoyancy defect and the projection of the housing hydrodynamic drag on the wire. The instrument was ill-suited to measurement in situations of strong current which reverses with depth. Under these conditions the instrument descends either slowly or not at all above the depth at which the wire is at rest with respect to the current and descends much faster below this depth. The descent rate may easily exceed the relative horizontal current in this situation, thus speed information re-
corded by the current meter (which uses a Savonius rotor aligned vertically) is sus-
pect. In addition, the unknown shape (catenary) of the hydrographic wire, the align-
ment (pitch) of the instrument during descent, and poor navigation (only occasional
satellite fixes were available) all degrade the quality of the data. As a result the cur-
rent data can be trusted only to give qualitative impressions of shear in the upper
500 m; they will not be discussed here.

### 3. Monsoon winds and equatorial current jets

The Indian Ocean experiences well known strong seasonal fluctuations in wind
speed and direction. These are conveniently described as a northeast monsoon
(strong northeasterly winds over the Arabian Sea and Bay of Bengal peaking in
January) and a southwest monsoon (even stronger southwesterly winds over the
same region peaking in July). Wyrtki (1973) noticed that a major feature of the
winds along the equator was that they became organized between the monsoons into
moderate westerlies almost all across the ocean. This eastward wind stress pattern
exists during April and May and again during September, October and November
according to Nederlands Meteorologisch Institut (1952) charts. Wind stress has a
westward component only weakly and in the western part of the ocean from January
through March. During the two monsoons, the wind stress is primarily meridional
at the equator.

Associated with the intermonsoonal westerlies at the equator are strong surface
currents which may reach the depth of the thermocline. Wyrtki (1973) noted and
documented these eastward jets which are strongest between 2N and 2S. Current
measurements over nearly two years at Gan in the Maldive Islands clearly show the
appearance of the jet (Knox, 1976) in the thermocline as well as at the surface with
speeds exceeding 100 cm/sec. Wyrtki estimated a transport of $22.5 \times 10^6$ m$^3$/sec
assuming an average speed of 75 cm/sec. Associated with the large transport of
water are variations in the depth of a mid-thermocline isotherm (20°C) by as much
as 40 m over large areas. According to Wyrtki, the strong eastward wind stress be-
tween the monsoons drives surface water to the east where it accumulates then re-
turns west via the north and south equatorial currents. O'Brien and Hurlburt (1974)
reproduced the basic features of this flow regime in a nonlinear numerical model.

Further evidence for these jets and ramifications of their existence are drawn from
our transoceanic section on the equator.

### 4. Thermocline structure

The CTD casts taken along the equator show dramatic change of pycnocline
structure across the Indian Ocean. Selected profiles of temperature, salinity, and $\sigma_t$
at equatorial stations are shown in Figures 2a-e. Contours of these quantities in
equatorial sections are drawn in Figures 3a-c.
Figure 2. Profiles of temperature, salinity, and $\sigma_t$ at stations along the equator. From east to west, (a), (b), (c), (d), and (e) are profiles from stations 4, 10, 16, 20, and 25, respectively. Salinity spikes of order 0.1% at depths where there are large temperature gradients may be artifacts of the CTD technique used.
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\[ \sigma_i \]

SALINITY %

\[ 0 \quad 5 \quad 10 \quad 15 \quad 20 \quad 25 \quad 30 \]

TEMPERATURE (°C)

\[ 0 \quad 200 \quad 400 \quad 600 \quad 800 \quad 1000 \]

PRESSURE (DBAR)

\[ 20 \quad 22 \quad 24 \quad 26 \quad 28 \quad 30 \]

SALINITY %

\[ 0 \quad 5 \quad 10 \quad 15 \quad 20 \quad 25 \quad 30 \]

TEMPERATURE (°C)

\[ 0 \quad 200 \quad 400 \quad 600 \quad 800 \quad 1000 \]

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PRESSURE (DBAR)

\[ 20 \quad 22 \quad 24 \quad 26 \quad 28 \quad 30 \]

SALINITY %
East of 73E (where the Maldive Islands cross the equator) the thermocline is very sharp and deepens to the east. Temperature changes by more than 10°C in 10 m depth at some stations. A sharp main pycnocline exists by virtue of the strong thermocline. Salinity reaches a maximum just above the pycnocline (with values exceeding 35.7‰ at some stations) and decreases to roughly 35.1‰ at the base of the pycnocline. The eastern stations (see Figure 2a or 3b) show rather fresh water in the upper 50 m or so. Salinities in this layer are as low as 34.1‰.

West of 73E the thermocline is rather diffuse, extending from near the sea surface to 200 m depth. The pycnocline is correspondingly diffuse and the salinity maximum is at or near the sea surface. Salinity contrasts are much weaker in the western part of the equatorial ocean where surface salinity is less than 35.3‰ west of 53E (see Figure 2e).

Below the pycnocline, salinity changes along the transect by at most 0.2‰. The equatorial section (Figure 3b) indicates a complicated picture of weak maxima and minima. A weak salinity maximum evident in the profiles east of 70E just beneath the pycnocline at roughly 150-200 m depth can be identified as Persian Gulf water. A deeper, thicker salinity maximum centered roughly at 700-800 m depth can likewise be identified with Red Sea water.

Salinities higher than 35.4‰ along the equator east of 65E are rare according to historical sections (Wyrtki, 1971). From their extensive surveys, Taft and Knauss
Figure 3: Equatorial sections: (a) temperature (°C), (b) salinity (ppt), (c) $\sigma_t$. 
(1967) report only a single bottle observation of water more saline than 35.6‰ within 1° of the equator, and that was made west of 49E (west of our transect) at roughly 100 m depth. Thus the isolated plug of water more saline than 35.7‰ extending from at least 74E to 85°30'E along the equator is truly anomalous. Salinities higher than 35.7‰ are found in the Arabian Sea and the southern Indian Ocean at roughly 30S, the two regions where evaporation exceeds precipitation sufficiently to form a saline water mass at the sea surface. The water mass known as Arabian Sea Water is more saline, less dense \((\sigma_t \sim 24)\), and is formed nearer to the equator than the saline water formed roughly at 30S \((\sigma_t \sim 26)\). The temperature-salinity characteristics of the anomalously saline water found on the equator are consistent with those of Arabian Sea Water, so it is plausible that the anomalous water was formed at the surface in the Arabian Sea.

The anomalously saline water on the equator does not appear to be connected directly to Arabian Sea Water; it is not part of a (presumably diffusive) tongue stretching south and east from the Arabian Sea. There is no evidence from meridional sections east of the Maldive Islands (Wyrtki, 1971) for such a tongue which might connect the anomalously saline water to the Arabian Sea source. The limited geographical extent of the saline equatorial plug suggests a primarily advective (rather than diffusive) transport of Arabian Sea Water to the east by equatorial currents.

The intermonsoonal jets noticed by Wyrtki (1973) are a plausible mechanism for transporting the saline water from the western part of the ocean through the Maldives and on to nearly 90E. Travelling at roughly 75 cm/sec (36 miles/day), 50 days are needed to travel the 30° of longitude from 60E (south of the center of the Arabian Sea Water Source) to 90E. The jets exist for about 2 months, making them a plausible way for the anomalously saline water to end up far in the eastern part of the Indian Ocean after having “diffused” to the equator south of the Arabian Sea.

At the onset of the northeast monsoon, when this survey was made, the jets also explain the depression of the pycnocline to the east. The dynamical implications of this depression are discussed in the next section.

Most of the profiles through the thermocline in the western region of the Indian Ocean show one or more mixed layers a few tens of meters thick each (see Figures 2d-e). The thickest of these are homogeneous to within 0.05°C and 0.01‰ over 70 m and are embedded in the thermocline. Their T-S characteristics are not different from those of the surrounding water, which suggests that they are formed \textit{in situ}, not by interleaving of different water masses. None of the layers was observed at more than one station, which indicates that the layers have horizontal scales less than roughly 100 km meridionally and 400 km zonally; this is added evidence that the layers are formed locally, perhaps by an instability involving the considerable vertical shear of swift equatorial currents.
5. Near-surface zonal pressure gradients

Qualitatively the depression of the thermocline toward the east indicates increasing pressure to the east at a given level, assuming some deep pressure surface is level along the equator. This impression is confirmed by computation of dynamic height relative to 800 dbar (the highest pressure common to all 28 profiles). Figures 4a-c are zonal sections of dynamic height at 1N, the equator, and 1S. At pressures
higher than 140 dbar, geopotential changes by less than 5 dynamic centimeters (1 dyn. cm = 10^5 erg/g) from 50°30'E to 95E. The strongest zonal pressure gradients are confined to the thermocline and above and persist over 2000 km or more along the equator. The sections at 1N and 1S are qualitatively the same as the equatorial section.

The most remarkable feature of the measured near-surface pressure field is that dynamic height increases to the east by roughly 8 dyn. cm. over the western region of the equatorial Indian Ocean and an additional 25 dyn. cm. over the eastern region at 20 dbar. This is in marked contrast to the dynamic height sections of Taft and Knauss (see Scripps Institution of Oceanography, 1968); they show an increase of roughly 18 dyn. cm. from 45E to 95E at 20 dbar relative to 400 dbar in July, 1962 and an increase of less than 4 dyn. cm. over the same region in March-April, 1963. A comparison of the Taft and Knauss dynamic height sections computed from hydrographic stations with the results of our CTD survey is given in Figure 5. At 100 dbar, the December-January transect indicates an increase of 16 dyn. cm. to
Figure 6. Horizontal pressure gradient force calculated from pairs of stations in $10^{-4}$ dynes/g. Negative (positive) values of $-p_e/p$ indicate a force to the west (east). Large scale gradients over the western (a) and eastern (b) portions of the survey are given by heavy curves. All
gradients are computed relative to 800 dbar. Note that a sequence of pairs of stations gives gradients which fluctuate smoothly changing the sense of the zonal force over two or more successive stations.
the east, whereas the July transect indicates an increase of less than 4 dyn. cm. and the March-April transect indicates a decrease of the same magnitude. The pressure field both at the surface and in the thermocline changes radically throughout the year, presumably in response to the monsoonal winds. The intermonsoonal jets of surface wind and current noticed by Wyrtki suggest that the near surface pressure field fluctuates at roughly the semi-annual period. At the end of the May eastward jet the zonal pressure field on the thermocline and above might well be qualitatively the same as measured during the December-January transect.

Horizontal pressure gradients computed from pairs of equatorial stations are plotted in Figures 6a and b, again taking 800 dbar as a reference level. The large scale pressure gradient (that is, the gradient calculated from stations 20° or more apart) is as high as $10.0 \times 10^{-5}$ dynes/g east of 74°E (i.e. east of the Maldives Islands). The large scale pressure gradient is weaker than about $5.0 \times 10^{-5}$ dynes/g west of the Maldives Islands. The large scale gradient becomes negligibly small below the thermocline as shown by the heavy solid curves in Figure 6a and 6b. The geopotential gradients measured by Taft and Knauss (1967) at 50 dbar relative to 400 dbar correspond to a pressure gradient of $2.9 \times 10^{-5}$ dynes/g in the July 1962 section and $-1.7 \times 10^{-5}$ dynes/g in the March-April section (where a positive pressure gradient indicates a westward force). The corresponding pressure gradient from our December-January section at 50 dbar is $9.2 \times 10^{-5}$ dynes/g over the eastern region.

Not only are the primarily westward pressure gradient accelerations found in the Indian Ocean in the opposite sense of those in the other equatorial oceans, but they also are the steepest gradients measured. Knauss (1963) reported gradients as strong as $-6.5 \times 10^{-5}$ dyne/g (at the sea surface) in the central Pacific Ocean with smaller values farther to the east and west. In the core of the equatorial undercurrent relative to 1000 dbar, Knauss (1966) reports a gradient of $-2.6 \times 10^{-5}$ dynes/g. Lemasson and Piton (1968) report a stronger gradient ($-4.8 \times 10^{-5}$ dynes/g) for 50 dbar relative to 700 dbar farther to the west in the Pacific. Katz et al. (1977) reports pressure gradients ranging from $-0.4 \times 10^{-5}$ to $-5.3 \times 10^{-5}$ dynes/g for 50 dbar relative to 500 dbar in the Atlantic west of 10W. He attributes the pressure gradient fluctuation to large scale wind fluctuations. The negative pressure gradient (eastward force) in the Atlantic and Pacific is due to the easterly winds found near the equator in both oceans.

6. Deep zonal pressure gradients

Beneath the thermocline, the large scale zonal pressure gradients virtually vanish (the solid curves in Figures 6a and 6b), but significant gradients appear over shorter scales. The small dynamic height fluctuations apparent in Figure 4b are expanded in Figure 7 for closed examination. Dynamic height at 200 dbar computed relative
to 800 dbar changes by roughly 2 dynamic centimeters from peak to trough along the equator. Similar fluctuations are found deeper as well. The pressure gradient plots (Figures 6a and b) indicate how smooth the zonal change in dynamic height is; differencing a sequence of adjacent stations gives successive curves which gently oscillate about zero pressure gradient. The fluctuations are not undersampled in longitude because one or more stations is between peaks and troughs in dynamic height. Their zonal wavelength appears to be roughly 1200-1500 km. Further, the amplitude of pressure gradient fluctuations appears roughly equal in the regions on opposite sides of the Maldive Islands.
These fluctuations are not attributable to measurement errors. To change dynamic height by 1 dyn. cm. at 300 dbar relative to 800 dbar would require a systematic error in salinity of 0.03%o, or three times the estimated drift given above. In order to produce the observed fluctuations in dynamic height, salinity errors would need to vary smoothly over a range of 0.06%o every several stations. This sort of instrument drift is unexpected, and there is no evidence from the bottle calibrations for such a drift, so the dynamic height variations in Figure 7 appear to be a real oceanic signal.

Are these fluctuations primarily zonal or temporal or an even mixture of both kinds of variability? These data do not answer this question, but the measurements reported by Luyten and Swallow (1976) indicate persistence of features with little discernable change in the current field over a month at 50E in May-June 1976. These current features are short vertically (vertical wavelengths of roughly 100-200 m) and extend virtually from surface to ocean bottom. However, that experiment yielded little zonal information of the current/density structure. It is plausible to assume that the current fluctuations would have a pressure signal associated with them, and this signal would be of magnitude \( \frac{P_s}{\rho} \sim u_o 2\pi / T \) where \( u_o \) is the magnitude of the current and \( T \) the period of the fluctuations, assuming that local acceleration and the zonal pressure gradient are of the same magnitude. A period of order a few months implies pressure gradients of about the magnitude observed in the zonal section according to this extremely simple model. It is plausible to regard the short vertical scale currents of Luyten and Swallow and the zonal dynamic height fluctuations as part of the same physical phenomenon.

If the dynamic height variability is primarily spatial (and not temporal, since the survey was completed in less than 3 weeks), then, assuming plane waves, phase changes found by tracing peaks (troughs) in depth and longitude indicate both a vertical and a horizontal wavelength. If we assume the horizontal wavelength to be roughly 1000-1500 km. and note that a peak in dynamic height at 300 dbar occurs near 71E and perhaps the same peak occurs at 50 dbar at 74E, we compute a vertical wavelength of roughly 600-800 m. This estimate, albeit crude, is consistent with the vertical scales Luyten and Swallow observed in the current field. Clearly a broad spectrum of vertical and horizontal wavenumbers (and probably frequencies) is present, since the sense phase propagation is not unambiguous from these data.

### 7. Discussion

Given the nature of the monsoons, strong temporal variation of dynamic height above the pycnocline is undoubtedly part of a forced dynamical response to wind stress. The strong equatorial jets are a component of this response and they, in turn, appear responsible for the large redistribution of density (as indicated by changing thermocline depth and structure) and water types (e.g. the appearance of Arabian Sea Water from 70E to 90E along the equator). The dynamics of this response are
far from understood; indeed the phenomena are not even adequately described by measurements.

The westward pressure gradient force usually found in the Indian Ocean is due to the winds, apparently especially the intermonsoonal westerly winds at the equator. These intermonsoonal winds are probably responsible for the strong zonal pressure gradient observed to occur at the beginning of the northeast monsoon. That these pressure gradients are roughly two to three times as strong as any observed over large scales in other equatorial oceans with comparable wind stress values is noteworthy, possibly indicating a dynamical asymmetry between eastward and westward wind forcing of equatorial oceans or a difference between steady and unsteady response. Gill (1972) pointed out that eastward momentum is generated at the equator by the Reynolds stress $\langle uv\rangle$, in response to either an eastward or westward wind. The tendency is for undercurrent flow to be weaker for an eastward wind than for a westward one, so that for a given wind stress magnitude the zonal pressure gradient will tend to be larger for an eastward wind than for a westward one.

It is yet to be demonstrated whether the deep fluctuations indicated here by zonal dynamic height changes are a forced or free response to wind forcing. The problem of transmitting momentum from the wind to an equatorial ocean has not yet been solved, but one can imagine some analog of the usual Ekman pumping at mid-latitudes or coupling with swift currents near the surface providing forcing beneath the surface layers. Wunsch (1977) used the former mechanism (postulating some vertical velocity beneath the surface) to model equatorial response to periodic forcing at fixed zonal wavenumber of an ocean with exponential stratification. The model gives deep flows which are qualitatively similar to the Luyten and Swallow (1976) observations. Vertical phase propagation is a feature of the model so that horizontal phase differences should also be evident if the ocean is forced and responds to a travelling periodic monsoon. If a variety of frequencies and wavenumbers are forced, a variety of vertical wavenumbers will be present.

Ideally long time series (a few years) of profiles of current and density would be useful in discerning just what the deep fluctuations are and what their role in equatorial circulation is. It is clear that the near-surface currents are nonlinear, however the deep fluctuations may be governed by nearly linear dynamics. Theoretically and observationally the problem of near-surface equatorial currents is more difficult. Valid time series of profiles near the surface would help determine what the relation between wind stress and current is as well as describe the complicated evolution of shallow equatorial currents in the Indian Ocean. The usefulness of single vessel transects is limited where the temporal and spatial scales of equatorial ocean fluctuations are mismatched to slow ship operation.

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