Forced resonant undulation in the deep Mascarene Basin

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Abstract

Current meters moored for 19.5 months at Lat. 20°S in the deep water of the western Mascarene Basin recorded a distinct, large-amplitude [O(10 cm s⁻¹)] undulation of bimonthly period, propagating westward at 7 cm s⁻¹. Its characteristics demonstrate that it was a barotropic Rossby wave of relatively large meridional scale. Simple theory accounts for it as having been forced by local wind-stress curl at one of the resonant frequencies of the Mascarene Basin. A sharp bimonthly peak is also prominent in spectra of TOPEX/POSEIDON sea-surface height in the Mascarene Basin, but is not seen to the eastward, as is consistent with the local generation. Fluctuations of 45-day period reported earlier in the upper ocean just northeast of Madagascar might have been generated through a similar process, but with frequency shifted by the South Equatorial Current.

1. Introduction

In the upper kilometer of the ocean just northeast of Madagascar, within the northern part of the South Equatorial Current, Quadfasel and Swallow (1986) and Schott et al. (1988) observed large velocity fluctuations of period 40–55 days. The former interpreted these as barotropic Rossby waves, probably wind-forced, the latter as an internal instability phenomenon. Near the southern edge of this current, in the deep water of the central Mascarene Basin, we recently found a very large, persistent, strikingly distinct undulation of somewhat greater period, around 59 days. In this paper we document the phenomenon, interpret it as a barotropic Rossby wave, discuss its generation, adduce evidence for its sea-surface manifestation, and compare it to the shallower observations off northern Madagascar.

As part of a WOCE project (designated ICM3) to measure the volume transports of the deep western-boundary currents in the South Indian Ocean, current meters were moored at six sites along Lat. 20°S in the western Mascarene Basin (Fig. 1). Attention will be limited here to the three moorings on the abyssal plain, Nos. 4–6, which were deployed from R.V. Knorr on 30–31 May 1995, and recovered by R.V. Melville on 18–19 January 1997. Each mooring carried four current meters, at nominal depths of 2500, 3500, 4000, and 4800 m, except that the deepest instrument on No. 4 was set for 4700 m. We identify each current meter by Mooring No. (4–6) and position in the vertical series (a, b, c, d, running from shallow to deep). Instrument 5b failed after 40 days, and will not be mentioned again. The record from 5c has...
several gaps between mid-September and the end of December 1995, amounting to half that time interval; it is otherwise unimpaired, but will not be used much in the analysis. All other instruments performed satisfactorily for the entire duration of the deployment. The ICM3 records are available from the WOCE Current Meter Data Assembly Center at Oregon State University.

2. Observations

As representative examples of the results from the 10 current meters, the low-passed (72 h) time series for the northward ($v$) and eastward ($u$) velocity components are displayed in Fig. 2 for the records from the top (4a, 5a, 6a) and bottom (4d, 5d, 6d) meters. The other four instruments produced qualitatively similar results. The overwhelmingly dominant feature in all records is the bimonthly oscillation over the entire 19.5 months of deployment, with meridional amplitudes of some 10 cm s$^{-1}$, zonal amplitudes somewhat smaller, and velocity vectors rotating counterclockwise more often than not. On the one hand, its occurrence is a great disappointment because it obliterates a portion of the mean flow that the current-meter array was put out to measure. On the other, it is a spectacular oceanic phenomenon that invites investigation for its own sake.

It takes no elaborate statistical analysis to measure the period of the oscillation. Nevertheless we calculated variance-conserving power spectra for all the low-passed records, subsampled at 6 h
intervals; three illustrative examples for the $v$-component are included in Fig. 3. The highlight is the bimonthly peak, but it is rather broad, and not quite the line spectrum that visual inspection of the time series might have suggested: for example, in the spectrum for record 6a (Fig. 3), the energy density is still half that of the bimonthly peak at periods of about 50 and 75 days (i.e., the “$Q$” of the peak is less than about 3). The period of maximum energy was very much the same in all 10 records, ranging from 59.2 to 59.5 days, for an average of 59.4 days. (The peak period was usually different for the $u$-spectra, but the $u$-series were generally muddier and much less energetic than the $v$-series, so we mainly disregard them.)

Similarly, it is easy to see simply by overlaying records from the different moorings on a light table that the undulation was propagating westward at about 7 cm s$^{-1}$. For a more exact estimate, lagged cross-correlations for the records from horizontally adjacent pairs of current meters yield average lag times for maximum correlation of 14.8 days for pairs 6a–5a, 6c–5c, and 6d–5d, and 14.6 days for pairs 5a–4a, 5c–4c, and 5d–4d. Dividing these times into the corresponding distances between moorings gives estimated westward phase speeds of 7.2 cm s$^{-1}$ from No. 6 to 5, and 6.8 cm s$^{-1}$ from No. 5 to 4, for an average of 7.0 cm s$^{-1}$.

There is no very noticeable vertical variation in amplitude of the undulation (Fig. 2), but our observations are confined to the relatively uniform, lower half of the water column, where vertical shear does not readily distinguish barotropic from baroclinic modes. The issue can be addressed indirectly. If, as we propose, the undulation is a Rossby-wave mode, the dispersion relation is $\omega = -\beta k (k^2 + \ell^2 + R^{-2})^{-1}$, where $\omega$ is
frequency, $k$ and $l$ the zonal and meridional wave numbers, respectively, $\beta$ is the derivative of the Coriolis parameter, and $R$ is the deformation radius for the mode. With respect to $k$ and $l$, the frequency is greatest for $l = 0$, $k = R^{-1}$, where it is $\beta R / 2$. For Coriolis parameter $f$ ($0.50 \times 10^{-4}$ m s$^{-1}$ at Lat. 20°S) and given buoyancy frequency $N(z)$, the deformation radii are the values of $R$ that allow solutions of

$$\frac{d}{dz} \left( \frac{f^2}{N^2} \frac{dF}{dz} \right) + \frac{1}{R^2} F = 0$$

...
exp \(i(kx + ly + \omega t)\) the amplitude ratio of \(u\) and \(v\) is \(l/k\). However, \(u\) is clearly not drastically smaller than \(v\) (Fig. 2). Much of this \(u\)-field may simply be background commotion, but in the power spectra for the 10 records the ratio of \(u\)-energy to \(v\)-energy at the frequency of the peak in \(v\)-energy is about 0.1 overall, which suggests a value of \(\epsilon \beta /\omega^2\) 10% greater than measured. Of course the discrepancy is not large.

We note that since the particle speeds are comparable to the phase speed, the wave is not at all of small amplitude, so the foregoing linear analysis may not be entirely valid. With respect to disregarding advective effects, however, what needs to be compared to the phase speed is the zonal component of particle velocity, which is certainly less than the meridional one.

The bimonthly undulation is less prominent at Mooring 3, and the time series at Nos. 2 and 1 are more confused, than at the eastern three moorings. This different behavior from that found above the abyssal plain seems unsurprising given the likelihood of boundary-layer dynamics close to the Madagascar slope, and, especially, given the properties of barotropic Rossby waves should be governed by the local gradient in \(f/H\), whose magnitude and direction are both quite different from that of \(f\) alone.

3. Resonant forcing

The breadth of the Mascarene Basin at the 4-km depth ranges from 540 km at Reunion to 900 km at Lat. 12°S, with a mean of 750 km. Since that is about an integral number (2) of wavelengths of the bimonthly undulation, it is very tempting to attribute the prominence of the undulation to some sort of a resonant selection of that frequency from among the jumble of forcing frequencies by the geometry of the basin. Phillips (1966) examined forced oscillations in a rectangular basin, and a simplified treatment of his problem is helpful in understanding the Mascarene Basin oscillation.

In a basin of zonal breadth \(L\) consider motion that is barotropic and essentially independent of the meridional coordinate, as suggested by the previous observational analysis, and is forced by a spatially independent wind-stress curl varying periodically in time. The linear vorticity equation is

\[
\left( \frac{\partial}{\partial t} + r \right) \psi_{xx} + \beta \psi_x = \frac{k \cdot \text{curl} \, \tau}{\rho D},
\]

where \(\psi\) is the stream function, \(r\) is a Rayleigh damping coefficient, \(\rho\) is water density, \(D\) the depth of the ocean, and \(\tau\) the wind-stress vector. Let \(k \cdot \text{curl} \, \tau = K(\omega) e^{i\omega t}\), where \(K(\omega)\) depends on the driving frequency but is otherwise a constant, and look for solutions of the form \(\psi(x, t) = \phi(x) e^{i\omega t}\). To avoid water piling up in the northern or southern parts of the basin, require that the basin-wide meridional transport be zero: \(\psi(0, t) = \psi(L, t) = 0\), for definiteness. Then,

\[
(i\omega + r)\phi_{xx} + \beta\phi_x = \frac{K}{\rho D}
\]

subject to \(\phi(0) = \phi(L) = 0\).

The solution satisfying these boundary conditions is

\[
\phi(x) = \frac{K}{\rho D \beta} \left[ x - \frac{L}{1 - e^{-aL}} \epsilon^{ikL} \epsilon^{i\omega t} \right],
\]

where \(a \equiv \beta r/(\omega^2 + \omega^2)\), \(k \equiv \beta r/(\omega^2 + \omega^2)\).

It follows that the complex meridional velocity component

\[
\psi_x = \frac{K}{\rho D \beta} \left[ \epsilon^{i\omega t} - \frac{L}{Q} (1 - e^{-aL}) \epsilon^{ikL} \right. \\
\left. \times (a - ik \epsilon^{i\omega t + kL}) \right],
\]

where \(Q \equiv 1 + e^{-2aL} - 2e^{-aL} \cos kL\). Consider high-frequency oscillations, in the sense that \(r/\omega \ll 1\), so that \(a/k \ll 1\), and therefore omit waves of amplitude \(aL\) with respect to those of amplitude
where \(\tan \gamma = e^{-aL} \sin kL / (1 - e^{-aL} \cos kL)\).

The wave number \(k \approx \beta / \omega\), and the phase of the wave propagates westward at speed \(c = -\omega / k \approx -\omega^2 / \beta\).

The amplitude of the wave in \(v\) varies with \(kQ^{-1/2}\), which, to order \((r / \omega)^2\), is maximized with respect to \(\omega\) when \(kL \sin kL = (2 + kL r / \omega)(1 - \cos kL)\). Thus resonance occurs for \(kL = 2m\pi, m\) being an integer; or when \(\omega = \omega_m \equiv \beta L(2m)^{-1}\), and the wave length is an integral fraction of the basin breadth. In this condition \(Q = (1 - e^{-aL})^2\), and \(\gamma = 0\), so

\[
v = \frac{K}{\rho D^2} \left( \cos \omega_m t - \frac{Lk}{1 - e^{-aL}} e^{-aL} \sin(\omega_m t + kx) \right)\]

Moreover, if \(aL \ll 1\) (as is shown plausible below), (6) reduces to the pleasingly simple result,

\[
v = \frac{K}{\rho D^2} \left[ \cos \omega_m t - \frac{\omega_m}{r} e^{-aL} \sin(\omega_m t + kx) \right],
\]

where now \(a \approx \beta r / \omega_m^2\) and \(k \approx \beta / \omega_m\).

The resonant waves, apart from the damping, are essentially the “normal basin modes” described by Pedlosky (1979, pp. 144–153). While, as he showed, the westward phase speed of the stream function for these modes is \(2\omega_m^2 / \beta\), whatever the combination of zonal and meridional wave numbers, the phase speed of the \(v\)-field, for zero meridional wave number, is just \(\omega_m^2 / \beta\), as above.

Because of the simplicity of (5)–(7) it is easy to see why resonance responses occur in this situation. The first term on the right-hand side is the oscillating Sverdrup transport (per unit depth) driven by the oscillating wind-stress curl. Since net meridional transport is not allowed in the basin, an opposed flow must counter the Sverdrup transport. For very low-frequency motions (e.g., steady ones) a western-boundary current could accomplish the balance; and for \(\omega \ll r, \alpha^{-1} = r / \beta\), the Stommel boundary-layer thickness. But with \(\alpha^{-1}\) comparable to, or much greater than, the basin breadth, this is not possible. Instead, balance is achieved by the net meridional transport in a Rossby wave of the same frequency as the Sverdrup transport. However, if the wave length should be an integral fraction of the basin breadth, the net meridional transport in the pure, unmodulated wave is exactly zero. The spatial variation in the wave amplitude is therefore needed to obtain a net meridional flow (e.g., somewhat greater northward flow in the western part of the wave than southward flow in the eastern part). Since for the high-frequency oscillations the amplitude modulator \((e^{-aL})\) diminishes only gradually across the basin, a very large basic wave amplitude is required in order to get enough net meridional transport in the modulated Rossby wave to balance the directly forced Sverdrup transport. Hence if motion is forced at a frequency corresponding to a Rossby wave length equal to an integral fraction of the basin breadth, a big response is to be expected.

To estimate realistic values for \(r\) (to the extent that \(r\) embodies realistic physics itself), suppose that the Rayleigh damping is a representation of friction (viscosity coefficient \(v\)) in a bottom Ekman layer acting on barotropic flow; then \(r = (\nu / 2D^2)^{1/2}\) (e.g. Gill, 1982, p. 414). From analogy to recent measurements of diffusivities, Johnson (1998) has suggested a plausible range of deep viscosities, \(1 \times 10^{-5}–1 \times 10^{-2} \text{ m}^2 \text{ s}^{-1}\), which, for \(f = 0.50 \times 10^{-4} \text{ s}^{-1}\) at Lat. 20°S, and \(D = 5000 \text{ m}\), imply a corresponding range of \(r, 3 \times 10^{-9} \text{ to } 10^{-7} \text{ s}^{-1}\). The smaller values would probably be more appropriate to the results from Moorings 4 to 6, given their location on an abyssal plain.

[From measurements of geostrophic drag coefficient compiled by Weatherly (1984), Dewar (1998) has estimated a vertical eddy viscosity of \(1.6 \times 10^{-3} \text{ m}^2 \text{ s}^{-1}\), within Johnson’s (1998) range.]

For bimonthly oscillations \(\omega \approx 10^{-6} \text{ s}^{-1}\), so \(r / \omega \ll 1, a / k \ll 1\), as asserted above. The de-colding scale \(\alpha^{-1} \approx 10^2 / \beta r\), which is about 5000 km for \(r = 10^{-8} \text{ s}^{-1}\), considerably greater than the
breadth of the Mascarene Basin, whereby \( aL \ll 1 \), as also asserted above. (But for greater \( r = 10^{-7} \) s\(^{-1} \) that approximation would not be justified.)

With a mean breadth of the Mascarene Basin between the 4-km isobaths, from Lat. 12°S to 22°S, of 750 km, the resonant period for \( m = 2 \) is 57 days, very close to the observed period at the moorings of 59 days. Sverdrup speeds are of size \( K(\rho Db)^{-1} \), or around 0.05 cm s\(^{-1} \) for fluctuations in wind-stress curl \( K \), say, of magnitude \( 0.5 \times 10^{-7} \) N m\(^{-3} \) (Fig. 4). Off resonance, the amplitude of the Rossby wave is several-fold greater, but at resonance it is magnified by a factor \( \omega/r \), and could easily be 100 times greater, as seen in the Mascarene Basin. The phase speed of the predicted wave is also consistent with the observations.

As already remarked, the observed frequency spectra (Fig. 3) are not line spectra, and while one does not expect line spectra in the presence of frictional damping, the peaks seem much too wide to be spread by friction alone. If \( aL \ll 1 \), the frictional width of a peak at the level of half the maximum energy would be \( O(r) \), so the width, in terms of period, would be only a day or two, much less than observed. However, the breadth of the real Mascarene Basin is not uniform: it varies by some 25% from the mean in Lats. 12–22°S. For the extreme values of the breadth the corresponding resonance periods are 47 and 79 days, so this imprecision in basin breadth might be an important circumstance tending to smudge the observed resonance peak.

Underlying the theoretical requirement of zero net meridional flow in the basin is the assumption that the northern boundary of the basin, extending northeastward from the northern tip of Madagascar to the northern Mascarene Plateau and then southeastward along the Mascarene Plateau to Lat. 10°N (Fig. 1), blocks throughflow into the Somali Basin beyond. This boundary does not actually reach the sea surface, and in fact is cut by several deep passages. Apart from narrow gaps, the crest of this ridge is generally shallower than 2000 m (Fisher et al., 1982), and we suppose that such a height would be sufficient to contain the barotropic flow registered by the current meters. The deepest and broadest cut through the ridge is at the Amirante Trench (near 8°S, 53°E); in that its breadth at the 2000-m level is about 110 km (Fisher et al., 1982), or three-tenths the wave length of the inferred Rossby wave, it does not seem large enough to allow much leakage through the boundary. [The problem of modes in a sub-basin bounded by an island or island group resembles those considered by Pedlosky and Spall (1999). A solution is feasible, and involves solving for the circulation around the islands, but is beyond the present scope.]

While solution (5) has zero net meridional flow, it contains alternating bands of opposed meridional flows, which are also supposed blocked by the northern boundary. The northward flows therefore must turn round to feed the southward flows, and so violate (somewhere) the condition of \( \gamma \)-independent \( \psi \). We assume that this happens near
the northern boundary, far from the domain of validity of (5). More complex models (e.g. Pedlosky and Spall, 1999) could be constructed to include \( y \)-dependence and realistically detailed bottom topography, but these seem unnecessary (and perhaps obfuscatory) for the basic features we are trying to highlight.

4. Winds

We inquire next into the availability of wind-stress curl forcing over the Mascarene Basin at the bimonthly period. Intra-seasonal oscillations in circulation and convection features in the tropical atmosphere are now widely recognized. Originally identified by Madden and Julian (1972) as of period 40–50 days, the energy-containing band was expanded at the one end to 40–60 days by Mertz and Mysak (1984), at the other end to 30–50 days by Krishnamurti et al. (1985), and was described as of 30–60 days by Knutson and Weickmann (1987)—although Madden and Julian (1994) themselves adhere to 40–50 days. The oscillation circles the globe (zonal wave number one), propagating eastward at speeds of 3–10 m s\(^{-1}\), and is associated with atmospheric heating anomalies over the Indian Ocean and western Pacific Ocean, perhaps being a forced wave formed over the monsoon regions, but the generating mechanism has not been conclusively identified (Krishnamurti et al., 1985; Knutson and Weickmann, 1987).

Specific reports have focused on near-equatorial fluctuations. To probe those at somewhat higher latitudes, time-series of wind-stress curl at one-day intervals were extracted from the National Centers for Environmental Prediction (NCEP) reanalysis data set for the interval of the mooring deployment, June 1995–January 1997. Variance-conserving spectra were then calculated for the transindian (Long. 39–121°E), zonal averages at Lats. 0, 12, and 20°S, for the zonal average over the longitude interval on Lat. 20°S spanning Moorings 4–6 (50°38′–54°22′E), and for the areal average over that part of the Mascarene Basin bounded by Madagascar on the west and by the Mascarene Plateau (plus Mauritius and Réunion) on the east (Lats. 12–22°S, Longs. 51–58°E). The estimated time series itself for the wind-stress curl averaged over this portion of the Mascarene Basin is given in Fig. 4, and the power spectra for the 12°S average, the local mooring-line average, and the Mascarene Basin average are shown in Fig. 5.

The spectrum for Lat. 12°S (Fig. 5a) has a pronounced annual period (registered here, for the frequency intervals used, at 300 days), and higher peaks at intra-seasonal periods near 65 and 40 days. Prominent peaks occur at similar periods in the spectrum for the equator (not shown), but the annual peak is higher than, and the two intra-seasonal peaks lower than, those at 12°S. The spectrum for the local wind-stress curl over the three moorings (Fig. 5b) also has an annual peak, and a broad, higher intra-seasonal peak in the interval 40–60 days, plus another peak near 30 days. The spectrum for the average curl over the Mascarene Basin (Fig. 5c), as well as for the zonally averaged curl at Lat. 20°S (not shown), is similar in shape, but the energy levels are several times greater in the local spectrum than in the basin-averaged or zonally averaged ones. The annual cycles at the equator and Lat. 12°S are the Asian monsoon, but those for Lat. 20°S and the Mascarene Basin average (Fig. 4) are 6 months out of phase with them; these latter regions lie just outside the traditionally identified area of monsoon influence (Ramage, 1971, pp. 2–5). Presumably the peaks near 40 days in all the spectra described are the Madden–Julian oscillation as narrowly delineated originally. The dominant 60-day period found by Mysak and Mertz (1984) in a 10-month record of zonal stress close to the equator in 1976 also appears in the NCEP curl spectrum for the equator, but energy peaks at other intra-seasonal periods are equally high there.

The local spectrum (Fig. 5b) is the average for time series at three longitudes, 50°38′, 52°30′, and 54°22′E. The peak periods for these individual spectra (not shown) are similar, but the levels and shapes of the peaks differ among the spectra, indicating modest small-scale variability in the NCEP analyses. It is difficult to gauge how realistic that is, especially given other uncertainty in the reliability of the NCEP product (Newman et al., 2000). This variability and the anomalously
high energy levels in the local spectrum lead us to surmise that the basin-averaged spectrum is probably the best estimator of the forcing for the presumably basin-scale oscillations observed at the moorings. This spatial variation in strength of wind-stress curl, however, is not consistent with the assumption of spatially uniform curl in Section 3; it is unclear to us whether the discrepancy, to the extent that it is real, limits use of the analysis at all severely.

Whatever the case, according to both Figs. 5b and c, at the time and location of our current-meter deployment there appears to have been a broad, intra-seasonal frequency band of high wind-stress-curl density suitable for driving oscillations in the Mascarene Basin. While the basin-averaged forcing was maximal at 45 days, that is not a resonance period of the basin [the amplitude factor $LkQ^{-1/2}$ in (5) is about 6 at 45 days for $r = 10^{-8}$ s$^{-1}$, $L = 750$ km], and little response occurred. Instead, the bimonthly forcing, being matched to the geometry of the basin, appears to have generated the great resonant undulation observed.

5. Discussion

As stated in Section 2, the period of the undulation demonstrates indirectly but conclusively that it is a barotropic mode. In addition, the 7.4-year record of the TOPEX/POSEIDON altimeter provides direct evidence of the sea-surface manifestation of the undulation. Variance-preserving spectra of sea-surface height for the full record length are compared in Fig. 6 for four locations: the center of the Mascarene Basin (15°S, 56°E), and a point on a similar latitude east of the Mascarene Plateau in the Central Indian Basin; and a point on Lat. 20°S midway...
between Moorings 5 and 6, as well as a point on the same latitude in the middle of the Central Indian Basin.

At Lat. 15°S a sharp, high, bimonthly peak, with maximum at 59 days (as in the current records), is seen in the Mascarene Basin, but not a hint of one farther east. At Lat. 20°S the bimonthly peak is only half as high, and is of shorter period, maximal at 54 days; again, there is no such peak in the Central Basin. Both spectra at 20°S contain a good deal more energy at periods of 70–200 days than those at 15°S; the opposite holds for the longer periods of 200–1000 days. Nevertheless, at each latitude, the bimonthly peak in the Mascarene Basin is much the most prominent feature.
The full peak is contained within the period interval 57–67 days at 15°S. The rms amplitude of height variability in that interval is 2.4 cm. If that value corresponds to the amplitude of a wave in sea-surface height of zonal wave length 359 km, as inferred from the current-meter records, then the amplitude of the corresponding meridional component of geostrophic surface velocity (for Lat. 15°S) is 11 cm s⁻¹. At 20°S in the Mascarene Basin the bimonthly peak occurs in the interval 48–58 days, with rms amplitude of 2.0 cm, and an accompanying meridional velocity amplitude of 7.0 cm s⁻¹. Both speed estimates match well the amplitudes of the bimonthly undulation recorded at depth by the current meters, and they thus confirm that it is a barotropic phenomenon. (The rms amplitudes of the corresponding meridional component of geostrophic surface velocity (for Lat. 20°S) is 3.0 cm s⁻¹.)

The fact that the bimonthly peak is prominent in the full 7.4-year height record indicates as well that the undulation is commonplace in the Mascarene Basin, and not likely to be a mere chance event during the 19 months of ICM3 observations. The further stark fact that no bimonthly concentration of variance whatsoever is seen at the sites east of the Mascarene Plateau demonstrates that the undulation is not forced somewhere to the east and then radiated or advected westward into the Mascarene Basin. We note in passing that ICM3 deep current-meter arrays also were deployed in Longs. 70–74°E and Longs. 88–93°E on Lat. 20°S: at neither of these were bimonthly oscillations noticeable either.

Because of its orbit inclination and sampling interval the TOPEX/POSEIDON satellite aliases the S2 and M2 tides as periods of 58 and 62 days, and the energy at these periods has therefore been deleted from the available altimetric record. It is unlikely that the bimonthly peak observed in the Mascarene Basin (Fig. 6) could be residual aliased energy because it is much broader than that for any tidal energy, and, as emphasized, no such peak is seen in the Central Indian Basin (C. Wunsch, personal communication).

The reduction in peak period of height variability at Lat. 20°S from that observed at Lat. 15°S, or from the peak period in deep velocity variability at 20°S, might be due to Doppler-shifting by the relatively shallow South Equatorial Current at 20°S. Between Lats. 14 and 17°S the Mascarene Plateau is effectively shallower than 200 m and could shield the Mascarene Basin from the South Equatorial Current. Between 17°S and Mauritius, however, passages roughly 2500 m deep are open to it. Ship-drift compilations (e.g. Rao et al., 1989) are too coarse to verify a branching or diversion of the current by the Mascarene Plateau, but, in conformity with the topography, models (Schott et al., 1988, Fig. 7; Semtner and Chervin, 1992, Plates 3 and 5) calculate much weaker upper-ocean zonal flow in the Mascarene Basin at Lat. 15°S than, say, at Lats. 13 and 19°S.

For a zonal current of uniform constant speed \(U\), the advective term to be added to the left-hand side of (1) is \(U\psi_{xxx}\). For \(a \ll k\), \(U\partial / \partial x \approx iUk\), and so, approximately, (2) is modified simply by replacing \(i\omega\) with \(i(\omega + Uk)\). The resonant frequencies then become \(\omega = \omega_m - Uk\), and for \(U\) westward (<0), as in the South Equatorial Current, they are shifted to higher values. Because \(k = \beta(\omega + Uk)^{-1} = \beta / \omega_m\), \(\omega - \omega_m = - U \beta / \omega_m\), and a mean westward current of only 0.8 cm s⁻¹, present at Lat. 20°S but not at 15°S, would move the peak period from 59 to 54 days. Since the South Equatorial Current is confined to the upper ocean, it would have had no effect on the velocity series recorded by the current meters beneath it. (The mean zonal velocities at Moorings 4–6 are statistically indistinguishable from zero, but, for what it is worth, their average was 0.3 cm s⁻¹ westward—insufficient for a noticeable frequency shift.) This argument is rough at best, and not intended to be much more than suggestive, because the simplicity of the calculation requires that \(U\) be spatially uniform, whereas the South Equatorial Current, as stated, is sharply sheared in the vertical.

In a further search for information on surface currents we scanned drifter trajectories from the WOCE surface-drifter archive. We found 22 drifters that remained between Lats. 12°S and 22°S in the Mascarene Basin for more than 2 months. None of these, however, showed prominent intra-seasonal periodicities, so we did not think it rewarding to examine them statistically.
Both the resonant-response interpretation and the contrasting altimetric records imply that the bimonthly undulation is specific to the Mascarene Basin. As stated at the outset, large intra-seasonal variations also have been reported in the upper ocean near the northern boundary of the basin, just northeast of Madagascar. One asks whether the two phenomena are related.

The amplitude of the velocity variations above 1100 m in the 11-month records from Moorings 190 and 191 (Fig. 1) was 10–20 cm s\(^{-1}\) (Schott et al., 1988) and was similar above 600 m (records < 5 months long) at Mooring 188 (Quadfasel and Swallow, 1986). This is somewhat larger than, but still comparable to, that at Moorings 4–6. Time lags between appearances of signals at Moorings 190 and 191 (Schott et al., 1988) indicated a westward phase speed of, very roughly and imprecisely, 9 cm s\(^{-1}\)—resembling that along the ICM3 line.

The big difference from the ICM3 observations was the frequency range of high energy, in which the energy density peaked at a period of 45 days (Schott et al., 1988, Fig. 5). The moorings northeast of Madagascar were set in the northern branch of the South Equatorial Current, however, where the frequency response may have been influenced by the mean advection, as speculated for the sea-surface height variability at Lat. 20°S. In this case, a westward current of about 2 cm s\(^{-1}\) would be needed to shift a 59-day period to 45 days. Since this current magnitude is not out of line with the mean values estimated by Schott et al. (1988), it is at least plausible that the energetic fluctuations northeast of Madagascar were also a resonant response to the basin geometry, but with frequencies shifted by the South Equatorial Current. [At Mooring 188 the mean zonal velocity was actually eastward, not westward (Quadfasel and Swallow, 1986), but Schott et al. (1988) already have pointed out that this result is statistically insignificant because of the brevity of the records.]

It is true that the curl spectrum for Lat. 12°S (Fig. 5a) has a peak at 40 days, but a coincidence of forcing and response frequencies alone does not easily account for the great strength of the response without the resonance. More unsettling is the location of the Moorings on the flank of Madagascar (like Nos. 1–3), where barotropic Rossby waves should be markedly affected by bottom slope, and where flat-bottom, resonant-response theory may be irrelevant. The average bottom slope between Moorings 190 and 191, for example, is \(4 \times 10^{-2}\), which, times \(f/D\), gives a topographic \(\beta\) 20-fold greater than the planetary value.

If these oscillations northeast of Madagascar were nevertheless flat-bottom modes generated over the abyssal plain to the east, the baroclinic modes are still precluded at the observed frequencies. The deformation radius for the first baroclinic mode was calculated as described in Section 2 for \(N^2(z)\) as determined at WOCE Indian Ocean Stations 720 (12°15′S, 55°00′E) and 721 (11°53′S, 54°33′E). The results were 90 and 88 km, respectively, implying a minimum period for a (flat-bottom) baroclinic Rossby wave of about 73 days.

Mysak and Mertz (1984) discussed near-surface fluctuations in the Somali Current in Lats. 2–4°S that were also prominent at periods of 40 to 60 days. These could hardly have been due to Mascarene Basin resonance. The authors suggested local atmospheric forcing, linked to the Madden–Julian oscillation, but did not propose a quantitatively sufficient mechanism. By analogy with results from one-layer, reduced-gravity models, Kindle and Thompson (1989) and Woodberry et al. (1989) advocated barotropic instability of the basic current system there, as well, perhaps, as just north of Madagascar. Not even in the models, though, was it discovered why periods near 50 days were preferentially selected.

Nevertheless, the large, distinct, bimonthly undulation in the deep Mascarene Basin at Lat. 20°S seems clearly to have been caused by local wind-stress-curl forcing at the mode-2 resonance period of the basin. The lack of a response (Fig. 3) at the mode-3 resonance period, 85 days, is unsurprising, given the gap in the wind-stress-curl spectrum in the neighborhood of that period (Fig. 5), and the decline in expected resonant-response magnitude with decreasing frequency (Eq. (7)). On the other hand, apart from a hint of a shoulder on the current spectra to the right of the 59-day peak, there is little suggestion of a peak
at the mode-1 resonance period, 28 days, either (Fig. 3). The sea-surface height spectrum for 15°S in the Mascarene Basin (Fig. 6) is equally striking for lack of a peak near 28 days. (While the companion spectrum for 20°S does have a subsidiary peak at 30 days, it has others at the unnoteworthy periods of 36, 77, and 142 days; all four may thus be unrelated to resonant forcing.) These omissions are puzzling, because while the curl energy is lower at this period than at the bimonthly one (Fig. 5c), the spectral gap is not so very deep, and the resonant response should be twice as great for the 28-day period as for the bimonthly one. It seems there “ought” to have been a prominent monthly undulation too, if the NCEP reanalysis is reliable; we have no idea why there was not one.

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