

1 Yanai Waves in the Western Equatorial Indian Ocean

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2 **Abstract.** Observations and models have shown the presence of intrasea-
3 sonal fluctuations in 20–30-day and 10–20-day bands in the equatorial In-
4 dian Ocean west of 60°E (WEIO). Their spatial and temporal structures char-
5 acterize them as Yanai waves, which we label low-frequency (LFYW) and
6 high-frequency (HFYW) Yanai waves, respectively. We explore the dynam-
7 ics of these intraseasonal signals, using an ocean general circulation model
8 (MOM) and a linear, continuously stratified (LCS) model. Yanai waves are
9 forced by the meridional wind τ^y everywhere in the WEIO most strongly dur-
10 ing the monsoon seasons. They are forced both directly in the interior ocean
11 and by reflection of the interior response from the western boundary; inter-
12 ference between the interior and boundary responses results in a complex sur-
13 face pattern that propagates eastward and has nodes. Yanai waves are also
14 forced by instabilities primarily during June/July in a region offshore from
15 the western boundary (52–55°E). At that time, eddies, generated by barotropic
16 instability of the Southern Gyre, are advected southward to the equator. There,
17 they generate a westward-propagating, across-equatorial flow field, v_{eq} , with
18 a wavenumber/frequency spectrum that fits the dispersion relation of a num-
19 ber of Yanai waves, and these waves are efficiently excited. Typically, Yanai
20 waves associated with several baroclinic modes are excited by both wind and
21 eddy forcing, and typically they superpose to create beams that carry en-
22 ergy vertically and eastward along ray paths. The same processes generate
23 LFYWs and HFYWS, and hence their responses are similar; differences are

²⁴ traceable to the property that HFYWs have longer wavelengths than LFYWs
²⁵ for each baroclinic mode.

1. Introduction

26 Yanai waves are equatorially trapped signals characterized by a meridional-velocity
 27 field that is symmetric about the equator with a single equatorial maximum, and by anti-
 28 symmetric zonal-velocity and sea-level fields with extrema located one equatorial Rossby
 29 radius from the equator. In the equatorial Indian Ocean (EIO), they are a prominent part
 30 of intraseasonal variability across the basin. Although first detected in the western EIO
 31 (WEIO; box in Figure 1a), recent research has focused on Yanai waves in the central and
 32 eastern regions, a consequence of the availability of direct current measurements there.
 33 In this paper, we further investigate WEIO Yanai waves, focusing on the processes that
 34 generate them.

1.1. Background

35 One measure of Yanai-wave activity is the part of anomalous sea level that is anti-
 36 symmetric about the equator (ASLA), as defined in Eq. (1) below. Figure 2 plots a
 37 variance-preserving power spectral map of ASLA from satellite observations across the
 38 WEIO during 2001, a year when intraseasonal variability at period less than 40 days is
 39 significantly strong. The spectra show a strong peak in the 20–30-day band that is local-
 40 ized near 50–55°E. There is also a weak peak at 10–20 days. Since the temporal resolution
 41 of the altimeter data is 7 days, the biweekly band is poorly resolved and so likely to be
 42 underestimated. In support of this idea, the biweekly band is much stronger relative to
 43 the 20–30 day band in our solution to an ocean general circulation model (OGCM; mid-
 44 dle and right panels of Figure 2), and it appears prominently in spectra of the surface,
 45 equatorial, meridional current field (v_{eq}) in the central and eastern EIO [Sengupta et al.,

2004; Masumoto et al., 2005]. Hereafter, we label the Yanai waves in these lower- and higher-frequency bands as LFYWs and HFYWs, respectively.

1.1.1. Low-frequency Yanai waves:

LFYWs were first observed in the Indian Ocean in the early 1980s by equatorial current-meter moorings deployed in the WEIO from 47–62°E: A 20–30-day signal was prominent in v_{eq} at 200 m during the summer monsoon [Luyten and Roemmich, 1982]. Subsequently, Yanai waves with a period of ~ 26 days were shown to exist in sea-surface temperature (SST) data by Tsai et al. [1992]. LFYWs have since been revealed in current observations from Acoustic Doppler Current Profilers (ADCPs) near the surface in the central EIO [David et al., 2011] and from 100–250 m at 90°E in the eastern EIO [Masumoto et al., 2005].

The LFYWs exhibit a prominent seasonality in the WEIO. Figure 1c is a longitude-time plot of altimeter ASLAs band-passed at 20–30 days. The signal begins in June, reaches a maximum during August to September between 50–55°E, and essentially is gone by November. We note that these LFYWs are separate from the waves observed farther east during the winter monsoon [David et al., 2011], suggesting that LFYWs west and east of the Maldives (73°E) are dynamically distinct.

Modeling studies suggest that LFYWs can be generated by both current instabilities and wind stress. Kindle and Thompson [1989] simulated 26-day Yanai waves in a non-linear, reduced-gravity model forced by a monthly wind-stress climatology (see Schott and McCreary [2001] for a review). They clearly cannot be forced by monthly winds; the authors attributed them to instabilities associated with the Southern Gyre, but did not determine precisely how they were generated. The equatorial τ^y and $(\sigma/\beta)\tau_y^x$ fields

69 also force Yanai waves [Miyama et al., 2006], where τ^y and τ^x are the meridional and
70 zonal components of wind stress, σ is frequency, β is the meridional derivative of the
71 Coriolis parameter, and the subscript y represents the meridional derivative. Figure 1b
72 shows spectra of the two quantities averaged from 45–60°E; they both have a peak in the
73 20–30-day band, indicating that wind might also force LFYWs in the WEIO. Moore and
74 McCreary [1990] explored this idea using a linear, continuously stratified (LCS) model
75 (see Section 2.2); among other things, they showed that, because the western boundary of
76 the WEIO is slanted, antisymmetric Yanai waves with a 30-day period can be generated
77 when forced by a *symmetric* τ^x , as well as by τ^y . Kelly et al. [1995] also explored the
78 forcing of 20–30-day Yanai waves by τ^y .

79 1.1.2. High-frequency Yanai waves:

80 In the central and eastern EIO, the dominant Yanai-wave signal in v_{eq} lies in the 10–20-
81 day (biweekly) band [Schott et al., 1994; Reppin et al., 1999; Sengupta et al., 2001, 2004].
82 Masumoto et al. [2005] also found strong biweekly signals at 90°E, located in the near-
83 surface ocean above the band of 20–30-day oscillations. Unfortunately, there are no data
84 from the WEIO that can resolve biweekly variability. As discussed below, our control-run
85 solution using an OGCM suggests that HFYWs are also present in the WEIO throughout
86 the year (Figures 2).

87 Sengupta et al. [2004] attributed the biweekly Yanai wave in the central EIO to direct
88 forcing by τ^y . Using both an OGCM and an LCS model, Miyama et al. [2006] con-
89 sidered the generation of HFYWs in detail, among other things considering the relative
90 importance of forcing by τ^y and τ_y^x .

1.2. Present research

91 While model simulations suggest that WEIO Yanai waves can be forced both by in-
92 stabilities [Kindle and Thompson, 1989] and the wind [Moore and McCreary, 1990; Kelly
93 et al., 1995], the relative importance of these mechanisms has not been determined; more-
94 over, the processes by which instabilities excite Yanai waves are unclear. Our goal in
95 this paper is to investigate these issues. Questions that we address include the follow-
96 ing. What instability processes are involved in forcing LFYWs in the WEIO? Does wind
97 stress also force them? What processes generate HFYWs in the WEIO? In particular, do
98 instabilities, as well as the wind, generate HFYWs there?

99 To address these questions, we obtain solutions to two types of models: a state-of-the-
100 art OGCM, the Modular Ocean Model (MOM) and a dynamically simpler, LCS system.
101 Each type has its own advantages and limitations. The MOM solutions simulate the
102 ocean’s response most realistically, including instability-driven flows, but it is difficult to
103 separate wind- and instability-driven signals. The LCS solutions, which are only wind-
104 driven, identify the basic properties of the wind-driven response, and help to identify that
105 response in MOM.

106 Key results are the following. Yanai waves in the WEIO are forced by both τ^y and
107 eddies, the latter generated by instabilities of the Southern Gyre that advect to the equator
108 from 52–55°E. Typically, Yanai waves associated with several baroclinic modes are excited
109 by both forcings, and they superpose to create beams that carry energy into the deep ocean
110 as well as eastward. Finally, differences between the LFYW and HFYW responses occur
111 primarily because the wavelengths of HFYWs are longer than those of LFYWs for all
112 baroclinic modes.

113 The paper is organized as follows. Section 2 reviews both models and the experiments
114 that are carried out with each of them. Section 3 describes our control runs and compares
115 them to available observations. Section 4 discusses the dynamics of the LFYWs and
116 HFYWs in the control runs, among other things isolating their wind- and instability-
117 driven flows. Section 5 provides a summary and discussion of results.

2. The ocean models

118 In this section, we first review our two ocean models (Sections 2.1 and 2.2), and then
119 report the solutions obtained for each of them (Section 2.3).

2.1. Modular Ocean Model

120 Our OGCM is version MOM4p1 of MOM [Griffies, 2009], configured for the Indian
121 Ocean as in Kurian and Vinayachandran [2007] and Vinayachandran et al. [2007]. MOM
122 has a free surface and uses the hydrostatic approximation. The primitive equations are
123 volume-conserving, employ the Boussinesq approximation, and the equation of state fol-
124 lows the nonlinear formulation of Jackett et al. [2006]. Tracer advection uses the multi-
125 dimensional, piece-wise parabolic method (MDPPM), with a flux limiter based on the
126 MITgcm (<http://mitgcm.org>). The model uses generalized orthogonal coordinates hori-
127 zontally and a quasi-horizontal, rescaled height coordinate vertically [Stacey et al., 1995;
128 Adcroft and Campin, 2004]. In our grid, the horizontal resolution is $0.25^\circ \times 0.25^\circ$, and
129 there are 40 vertical levels with 25 levels in the top 200 m and a 5-m resolution in the top
130 60 m.

131 The model domain extends over the Indian Ocean from $30\text{--}120^\circ\text{E}$ and $30^\circ\text{S}\text{--}30^\circ\text{N}$ (Fig-
132 ure 1). Bottom topography is based on a modified ETOPO2 bathymetry [Sindhu et al.,

133 2007], with the minimum ocean depth set to 15 m and isolated land points (i. e., single
134 cells) removed. The Palk Strait between India and Sri Lanka is closed, and the outflow
135 regions of the Red Sea and Persian Gulf are broadened and deepened to ensure smooth
136 flow between these narrow channels and the Arabian Sea (Figure 1a). At land bound-
137 aries, a no-flux condition for tracers and no-slip and no-normal-flow conditions for velocity
138 are imposed. Sponge layers of width 3.5° are applied to the open southern and eastern
139 boundaries, where tracer fields are relaxed to climatological values [Locarnini et al., 2010;
140 Antonov et al., 2010] with a time scale of 30 days; no restoration is applied to the tracer
141 fields elsewhere.

142 The K-profile parametrization (KPP) scheme [Large et al., 1994] is used for vertical
143 viscosity and diffusivity with a pre-specified vertical-diffusivity profile [Bryan and Lewis,
144 1979] and a critical bulk Richardson number of 0.3. Local mixing due to double diffusion
145 is not included to avoid enhanced mixing within the stratified water column. Horizontal
146 friction is a combination of biharmonic and Laplacian mixing, with Smagorinsky mixing
147 coefficients of 0.01 (velocity scale 0.04 m s^{-1}) and 0.1 (velocity scale 0.005 m s^{-1}), respec-
148 tively [Griffies and Hallberg, 2000]. Penetrative shortwave heating in the upper 100 m is
149 calculated using a Chl-*a*-based scheme [Morel and Antoine, 1994].

150 Sources for the model forcing are listed in Table 1. River discharge is specified in the
151 top two layers of the vertical grid. Wind forcing is from the QuikSCAT data set for
152 2000–2008. For future reference, Figure 3 presents longitude-time plots of τ^y band-passed
153 from 20–30 days (left) and 10–20 days (right). Both bands are similar in that they have
154 strong variability during the summer monsoon. They differ in that the higher-frequency

band is present more uniformly throughout the year, with strong variability also during
the winter monsoon as well as in the eastern WEIO during the intermonsoons.

2.2. LCS Model

The LCS model is described in detail elsewhere [McCreary, 1980; McCreary et al., 1996; Miyama et al., 2006; Shankar et al., 1996, 2010; Aparna et al., 2012]. Its equations of motion are linearized about a state of rest with a background Brünt-Väisälä frequency [Moore and McCreary, 1990], and the ocean bottom is assumed to be flat. With these restrictions, solutions can be represented as expansions in the vertical modes (barotropic and baroclinic) of the system. Solutions are found for a realistic basin geometry, in which the 200-m isobath of the MOM topography forms the coastline, and the horizontal resolution is the same as that used for MOM. The ocean bottom is at 4000 m and 25 vertical modes are summed to obtain the sea-level and velocity fields. Following Moore and McCreary [1990], wind stress is introduced as a body force that is constant from the surface to 50 m, ramps linearly to zero from 50–100 m, and is zero at greater depths. Vertical viscosity and diffusion are included with the coefficients, $\nu = \kappa = A/N_b^2$, where $A = 1.3 \times 10^{-4} \text{ cm}^2/\text{s}^3$. This choice for A ensures that vertical mixing impacts only high-order baroclinic modes, and so does not have a strong impact on our solutions. For example, the resulting damping time scale for the n th baroclinic mode is c_n^2/A , where c_n is the characteristic speed; using the values of c_n Table 3, c_n^2/A varies from 6215 days to 79 days as n increases from 1 to 10.

2.3. Numerical experiments

174 Solutions to both models are initialized from a state of rest, with temperature and
 175 salinity fields in MOM set using the climatology of Chatterjee et al. [2012]. A background
 176 solution is then spun up for seven years using climatological forcing data for 2000–2008.
 177 All runs are then started using the final state of the background run as initial conditions.

178 The MOM and LCS control runs (labeled MOMCR and LCSCR, respectively) are the
 179 interannual simulations forced by the full QuikSCAT winds from 2000–2008. To isolate the
 180 role of instabilities, we also forced MOM with 40-day low-passed wind stress (Experiment
 181 MOMLP τ), the assumption being that any high-frequency signals in this experiment must
 182 result from a nonlinear process [Kindle and Thompson, 1989]. In Sections 3 and 4, we
 183 show these solutions during 2001.

184 We also carried out additional test experiments to isolate key processes. Most of the
 185 tests are defined and discussed in Section 4, and all of them are listed in Table 2. Here,
 186 we note four test experiments that are *not* discussed later. Specifically, to identify the
 187 individual effects of forcing by intraseasonal meridional and zonal winds, we forced both
 188 models by wind fields that separately low-pass τ^x or τ^y , and so retain intraseasonal forc-
 189 ing only for τ^y (Solutions MOM τ^y and LCS τ^y) or τ^x (Solutions MOM τ^x and LCS τ^x),
 190 respectively. We find that in the WEIO, Solutions MOM τ^y and LCS τ^y are almost the
 191 same as in their respective control runs, whereas the responses of MOM τ^x and LCS τ^x are
 192 weak there. We conclude that forcing by τ^y dominates that by $(\sigma/\beta)\tau_y^x$ in the WEIO.
 193 In fact, $(\sigma/\beta)\tau_y^x$ is significant only in the eastern part of the WEIO, where its magnitude
 194 reaches a third of that of τ^y .

3. The MOM control run

195 Here, we describe basic properties of our MOM control run, MOMCR, and compare it
 196 with available observations. A detailed validation of a similar MOM solution was done by
 197 Kurian and Vinayachandran [2007]. We therefore restrict our validation to Yanai waves
 198 and other antisymmetric equatorial signals in the WEIO. We note that the LCSCR also
 199 has a realistic Yanai-wave response, but without a contribution from instability forcing
 200 (Section 4).

201 One measure of antisymmetric equatorial signals is

$$202 \quad \text{ASLA} = \frac{1}{2} (\text{SLA}_n - \text{SLA}_s), \quad (1)$$

203 where SLA_n and SLA_s are SLAs at $\pm 3^\circ$, roughly an equatorial Rossby radius north and
 204 south of the equator [Tsai et al., 1992]. Another is the meridional current at the equator,
 205 $v_{eq} \equiv v(x, 0, t)$. Both measures are consistent with Yanai-wave properties, but neither
 206 rules out the possibility of detecting other types of antisymmetric signals (other free
 207 waves, eddies, and a directly wind-forced response). At the intraseasonal frequencies of
 208 interest to us, the only possible, antisymmetric, equatorially trapped wave is the Yanai
 209 wave, so detection of other waves is not a concern. Both eddies and the directly-forced
 210 response, however, are as large as the Yanai waves they excite. Thus, since the group
 211 velocity of Yanai waves is eastward, both measures detect pure Yanai waves only east of
 212 any forcing region.

213 Figure 2 plots a variance-preserving spectral map of ASLA across the WEIO from
 214 MOMCR (middle panel). Consistent with the observed spectral map (Figure 2, left
 215 panel), the modeled spectra have prominent peaks in the 20–30-day band and a weaker
 216 one in the 14–17-day band, the latter corresponding to the biweekly peaks in the central

217 and eastern EIO [Sengupta et al., 2001; Masumoto et al., 2005]. Figure 2 also plots a
218 spectral map of v_{eq} (right panel). The 20–30-day band shows an energy maximum from
219 52–55°E, which is due primarily to the cross-equatorial movement of the unstable eddies
220 (see the discussion in Section 4.2). In contrast to the modeled ASLA map (middle panel),
221 the v_{eq} variability in the 10–20-day band extends all the way to the western boundary.

222 Unfortunately, current observations have not been taken in the WEIO since the mea-
223 surements reported by Luyten and Roemmich [1982]. The westernmost mooring deployed
224 in the EIO under the RAMA (Research Moored Array for African-Asian-Australian Mon-
225 soon Analysis and Prediction) program is at 80.5°E [Iskandar and McPhaden, 2011], and
226 data are available only from November 2004 onwards. A comparison for 2005 shows that
227 the model v_{eq} matches the observations fairly well (Figure 4), the correlation being 0.54
228 between observed and model raw meridional current and 0.58 between the 40-day high-
229 passed observed and model current. A comparison for other years also shows similar
230 correlations. Correlations between 40-day high-passed observed and modeled ASLAs are
231 not as high, with values of 0.43, 0.33, and 0.16 at 55°E, 80°E and 95°E, respectively. One
232 possible reason for the weaker ASLA correlations is that the 20–30-day band is so weak
233 in the observations.

234 Figure 5 presents longitude-time plots of 20–30-day band-passed ASLA (left panel) and
235 v_{eq} (middle-left panel) from MOMCR during 2001. During the summer monsoon, the
236 MOMCR ASLA response is similar to that of the altimeter (Figure 1c). Later in the year,
237 however, MOMCR is weaker than the observations near the coast, perhaps because its
238 wind-driven response is too weak there. There is variability in locations and at times with

239 little or no forcing (*e. g.*, after the first week of September when τ^y is weak), and hence
 240 these signals can only be LFYWs.

241 A striking aspect of the plots in Figure 5 is the slope of phase lines. They are similar in
 242 the observed and modeled fields. Moreover, they are consistent with those of theoretical
 243 Yanai waves. The Yanai-wave dispersion relation is

$$244 \quad k_n = -\frac{\beta}{\sigma} + \frac{\sigma}{c_n}, \quad (2)$$

245 where k_n and c_n are the wavenumber and characteristic (Kelvin-wave) speed of the n th
 246 baroclinic mode. Table 3 lists values of c_n and $\lambda_n = 2\pi/k_n$ for modes 1–10 and for several
 247 values of $P = 2\pi/\sigma$. The wavelength of the signals in Figures 1c and 5 are typically 600–
 248 1000 km (Table 3 uses negative values for the wavelength of westward-propagating waves.
 249 The sign is dropped in the text because we consider only these westward-propagating
 250 waves).

251 For $P = 20$ days, there are no values of λ_n that lie within this range, whereas for $P = 25$
 252 and 30 days there are several such values for $n \leq 2$ and $n \leq 4$, respectively. These Yanai
 253 waves are the ones that are most visible in the responses, but waves with large λ_n are also
 254 present (see Section 4).

255 Figure 6 (left panel) shows a zonal section of v_{eq} from MOMCR band-passed at 20–30
 256 days on 5 September, 2001, a time when the subsurface response is particularly large. The
 257 surface-forced Yanai waves associated with several baroclinic modes superpose to carry
 258 energy into the deep ocean along ray paths at the slope $dz/dx = -\sigma/N_b$, where $N_b(z)$ is
 259 the background Vaisala frequency [e.g., McCreary, 1984; Miyama et al., 2006]. The ray
 260 path obtained by integrating dz/dx with $\sigma = 2\pi/(25 \text{ days})$ from a point at the surface
 261 and the coast is plotted in the figure, and it closely follows the modeled beam.

262 Figure 7 and the right panel of Figure 6 show that similar features exist for HFYWs,
 263 and that they are prominent over most of the year. During January–February and June–
 264 July, they extend to the western boundary, whereas at other times they are confined
 265 to the eastern WEIO or are weak. Note that phase lines slope much less (*i. e.*, have
 266 longer wavelengths) than they do for LFYWs. This property is consistent with dispersion
 267 relation (2): Provided that σ is low enough for k_n to be negative (*i. e.*, to have westward
 268 phase velocity), it follows from Eq. (2) that the wavelengths of Yanai waves increase with
 269 σ (decrease with P) for all baroclinic modes. A comparison of λ_n values listed in Table 3
 270 for the two period bands illustrates this property.

4. Dynamics

271 Figures 5 and 7 present longitude-time plots of v_{eq} band-passed at 20–30 days and
 272 10–20 days, respectively, for Solutions MOMCR (panels b), MOMLP τ (panels c), and
 273 LCSCR (panels d). The three solutions illustrate the total, instability-driven, and wind-
 274 driven parts of the response, respectively. Solution MOMLP τ (panels c) has a patch
 275 of variability from 52–58°E mostly during June–July. This patch is absent in the linear
 276 LCSCR solution (panels d) and we conclude that it is due to instabilities associated with
 277 the Southern Gyre 4.2). This western-boundary regime off Somalia has been shown earlier
 278 to be the one part of the Indian Ocean where even the seasonal cycle of the circulation is
 279 significantly affected by nonlinearity during the summer monsoon [Shankar and Shetye,
 280 1997]. It is absent in Solution LCSCR (panels d), which has no instabilities. Very roughly,
 281 since MOM is nonlinear, panels c and d superpose to give panel b (Solution MOMCR).
 282 Note that throughout most of the year and most of the WEIO (with the exception of the
 283 longitude-time patch noted above), the features of the MOMCR solution are reproduced

284 reasonably well by the strictly wind-forced LCSCR. In this section, we report test solutions
 285 that examine the wind and instability forcing mechanisms of Yanai waves in the control
 286 runs in detail.

4.1. Wind forcing and boundary effects

287 To better understand the impacts of wind forcing and boundary effects in solution
 288 LCSCR (Figures 5d and 7d), we obtained a suite of test solutions forced by the idealized
 289 τ^y field,

$$290 \quad \tau^y = \tau_o X(x) Y(y) T(t), \quad (3a)$$

291 where

$$292 \quad X(x) = \cos\left(\pi \frac{x'}{L_x}\right) \theta\left(\frac{L_x^2}{4} - x'^2\right), \quad Y(y) = \frac{1}{2} \left[1 + \cos\left(\pi \frac{y}{L_y}\right)\right] \theta\left(\frac{L_y^2}{4} - y^2\right), \quad (3b)$$

$$293 \quad T(t') = \sin(\sigma_w t') \theta\left[\frac{(\Delta t)^2}{4} - t'^2\right], \quad (3c)$$

295 $\tau_o = 1.5 \text{ dyn/cm}^2$, $P_w \equiv 2\pi/\sigma_w = 25 \text{ days}$ or 15 days , $t' = t - t_0$, $x' = x - x_o$, $L_x = 70^\circ$,
 296 $L_y = 5^\circ$. Variable θ is a step function, which ensures that τ^y is confined to the region
 297 $x_o - L_x/2 < x < x_o + L_x/2$, $-L_y/2 < y < L_y/2$, and is switched on at $t = t_0 - \Delta t/2$
 298 and off at $t = t_0 + \Delta t/2$. Solutions are obtained in a rectangular basin with a western
 299 boundary at $x = 20^\circ\text{E}$. The zonal scale L_x was chosen to mimic the scale of the observed
 300 winds (Figure 3).

301 In the interior ocean, the Fourier transform in space and time of $F(x, t) = X(x)T(t)$,
 302 $\mathcal{F}(k, \sigma) = \tilde{X}(k)\tilde{T}(\sigma)$, determines which of the Yanai waves in Table 3 can be excited by
 303 the wind: Only those waves with σ and k values that lie within spectral peaks of \mathcal{F} will be
 304 significantly excited. Because $X(x)$ and $T(t)$ are limited in time and space, \mathcal{F} has power
 305 in spectral peaks at periods and wavelengths other than P and $2L_x$. As a consequence, a

306 variety of Yanai waves can appear in the response. (See Kelly et al. [1995] for a discussion
 307 of the Yanai waves excited by a switched-on wind.)

308 At the western boundary, Yanai waves are generated because τ^y forces zonal flow into
 309 (or out of) the boundary: The flow is eliminated by the reflection of a chain of short-
 310 wavelength Rossby waves that culminates in a Yanai wave [*e. g.*, Moore, 1968; McCreary,
 311 1986]. Because boundary response is forced only by τ^y right at the coast, there is no
 312 constraint on the zonal wavenumber of the Yanai waves. As a result, more Yanai waves
 313 are present in the boundary response than in the interior response, typically with shorter
 314 wavelengths (see below).

315 4.1.1. Low-frequency Yanai waves:

316 Figure 8 shows longitude-time plots and vertical sections of v_{eq} from three test solutions
 317 with $P_w = 25$ days. Solutions LCSIN25 (left panels) and LCSWB25 (right panels) are
 318 forced by (Eq 3a) with $\Delta t = 75$ days and $t_0 = 7$ July so that the wind switches on on 1
 319 June and off on 14 August; they differ in that $x_o = 60^\circ\text{E}$ or 20°E , respectively, so that
 320 τ^y is either confined to the interior ocean or centered on the western boundary. Solution
 321 LCSDIF25 is the difference between Solution LCSWB25 and Solution LCSIN'25 (Solution
 322 LCSIN25 shifted westward by 40°); as such, it provides a measure of the boundary-
 323 generated response.

324 In Solution LCSIN25 (Figure 8, top-left panel), the Yanai waves that have a period
 325 close to $P = 25$ days (the primary response) have long wavelengths, which best fit the
 326 large zonal scale of the wind; this property is evident in that the response consists of
 327 zonally-oriented bands. For $P = 20, 25,$ and 30 days, Yanai waves with long wavelengths
 328 correspond to the $n = 4-6, n = 7-9,$ and $n \geq 10$ baroclinic modes, respectively (Table 3).

329 These modes can superpose to form a beam that carries energy eastward and downward
 330 along ray paths (Figure 8, bottom-left panel) and, since the beam is composed of long-
 331 wavelength Yanai waves, phase lines are oriented largely horizontally. Note that the
 332 number of zero crossings in the beam is 6, consistent with the above inferences from the
 333 dispersion relation.

334 When τ^y switches off on 14 August, the LFYWs propagate from the forcing region along
 335 ray paths as a packet of free waves. Because the wave packet is composed of a number
 336 of vertical modes, energy still propagates vertically as well as eastward. As a result, the
 337 surface response in the forcing region weakens rapidly after 14 August. Interestingly, there
 338 is a patch of variability from 90–120°E and from January–March: It is the primary wave
 339 packet, which has bounced off the ocean bottom and to return to the surface.

340 In addition to the primary response, signals with considerably shorter wavelengths and
 341 periods other than $P = 25$ days are visible in Solution LCSIN25, a consequence of $T(x, t)$
 342 and $X(x)$ being limited in time and space. In this regard, note that waves with short wave-
 343 lengths emanate from the edges of the wind region where the slope of the forcing changes
 344 rapidly in x . Yanai waves with long wavelengths and a period $\lesssim 20$ days are also visible
 345 after 14 August; the sporadic nature of this response indicates that two or more waves
 346 with somewhat different periods are present, interfering with each other constructively
 347 and destructively.

348 When there is a western boundary, the total response is a superposition of directly wind-
 349 driven and boundary-generated responses, as measured by Solutions LCSIN25 (shifted
 350 westward by 40°) and LCSDIF25. In contrast to Solution LCSIN25, Solution LCSDIF25
 351 (which has no long-wavelength constraint) consists of zonally tilted bands with a zonal

352 wavelength of $\lesssim 1000$ km (Figure 8, top-middle panel). The boundary-generated Yanai
 353 waves also superpose to form a beam. In this case, because low-order baroclinic modes
 354 with shorter wavelengths dominate the beam, phase lines are oriented much more vertically
 355 (Figure 8, bottom-middle panel). Note that there is also a subsurface region west of the
 356 beam where phase lines are nearly horizontal (Region A). It exists only because Solution
 357 LCSDIF25 is not a perfect measure of the boundary-generated response: Part of the
 358 response in Solution LCSIN'25 is driven by winds west of 20°W (i. e., essentially coming
 359 from outside the basin), and Region A results from this erroneous contribution.

360 As for the interior response (LCSIN25), when τ^y is switched off, the response propagates
 361 eastward and downward as a packet of free Yanai waves. The patch of variability after
 362 October and east of $\sim 60^\circ\text{E}$ is the return of this wave packet to the ocean surface after it
 363 has reflected from the bottom. In addition, secondary waves, excited because $T(t)$ and
 364 $X(x)$ are not perfect sinusoids, are visible as well. They form bands of variability with
 365 shorter periods and longer wavelengths than the primary band, and they appear right of
 366 it (are less tilted vertically) because they have a faster group velocity. Conversely and
 367 less clear, bands with longer periods and shorter wavelengths appear left of the primary
 368 beam.

369 Solution LCSWB25 (Figure 8, right panels) is a superposition of Solutions LCSIN'25
 370 and LCSDIF25, and its overall structure is similar to those of its two components. A
 371 striking aspect of the total response is that it exhibits nodes where its two parts interfere
 372 destructively (Figure 8, marked by black arrows), most prominently in the region of the
 373 primary forced response (west of 50°E and prior to 15 August).

374 4.1.2. High-frequency Yanai waves:

Figure 9 illustrates the response when $P = 15$ days. It is similar to the response at $P = 25$ days with some notable differences. In Solution LCSIN15 (left panels), the primary (period of 15 days) response is now dominated by the $n = 3$ baroclinic mode, since it has the longest wavelength (Table 3). There are also secondary Yanai waves, which are more strongly excited than in Solution LCSIN25 because they have longer wavelengths that fit better with the wind. They also have a slightly smaller period, thereby generating an interference pattern with the primary response. The primary waves also form a beam dominated by the $n = 3$ baroclinic mode (bottom-left panel). It descends into the deep ocean at a slope that is steeper than that for 25-day waves by the factor $25/15 = 1.7$; the region of intensified flow during September and October is caused by the return of the primary beam to the ocean surface.

Similar differences can be seen between Solutions LCSWB15 and LCSDIBF25 and their counterparts at $P = 25$ days. For example, one difference is that the nodes in Solution LCSWB15 are separated by a greater distance than in Solution LCSWB25, a consequence of 15-day Yanai waves having longer wavelengths than 25-day waves. Another is that there is no tilt of phase lines across the 15-day beam, a result of so few vertical modes contributing to it.

4.1.3. Comparison to LCSCR and MOMCR:

Aspects of Solutions LCSWB25 and LCSWB15 can be identified in Solutions LCSCR and MOMCR, including the existence of offshore nodes (compare the top-right panels of Figures 8 and 9 with the right and middle-left panels of Figures 5 and 7). Differences between the idealized and realistically-forced solutions result from differences in the wind forcings. For LCSCR and MOMCR band-passed at 20–30 days, a single forced

398 response like LCSWB25 dominates the response beginning about mid-May. In LCSCR
 399 and MOMCR band-passed at 10–20 days, there are three responses like LCSWB15, be-
 400 ginning about 1 January, mid-May, and 1 September. The vertical structure of the beams
 401 is also consistent between the realistic and idealized solutions, for example, with beams in
 402 MOMCR having a smaller vertical scale when band-passed at 20–30 days than at 10–20
 403 days (compare the bottom-right panels of Figures 8 and 9 with Figure 6).

4.2. Eddy forcing

404 At the onset of the summer monsoon, the Somali Current begins to flow northward
 405 across the equator. Potential-vorticity conservation prevents it from continuing to flow
 406 along the coast, and so it bends offshore somewhat north of the equator to form the
 407 Southern Gyre (*e. g.*, Schott and McCreary, 2001). In MOMCR, the swift currents of the
 408 Southern Gyre become unstable and generate eddies. These eddies, in turn, are advected
 409 southward to the equator between 50–55°E, where they excite Yanai waves.

4.2.1. Eddy generation

411 To identify important nonlinear processes, we analyze the eddy-kinetic-energy (EKE)
 412 budget. Following Masina et al. [1999], we divide the flow field into mean and perturbation
 413 (eddy) parts, $p = \bar{p} + p'$, $\mathbf{V} = \bar{\mathbf{V}} + \mathbf{V}'$ and $w = \bar{w} + w'$, where $\mathbf{V} = (u, v)$ is the total
 414 horizontal-velocity vector, $\bar{\mathbf{V}}$ and \mathbf{V}' are defined similarly, w is the total vertical velocity,
 415 and \bar{p} , $\bar{\mathbf{V}}$ and \bar{w} are the means of their respective fields. To estimate the mean fields,
 416 we have used a 31-day running-mean Parzen window. With this definition, an “eddy”
 417 is simply the deviation from the mean field. We note that the mean field itself evolves
 418 over the averaging period (Figure 10, right panels), but these changes in the mean field
 419 are much slower than the changes in the deviations from the mean. The mean current

420 associated with the Southern Gyre, for example, shows that the gyre retains its identity
 421 through June–August even as it moves northward. This slow change in the mean field
 422 permits the use of a 31-day window; with this choice, the filter is short enough to allow
 423 for the slowly changing mean flow and long enough not to disturb our signal of interest
 424 (periods less than 30 days). The basic processes demonstrated here are, however, robust
 425 as they remain unchanged even if we use monthly means, as done by Masina et al. [1999],
 426 instead of the running mean (computed using the Parzen window) to define the mean
 427 fields.

428 It follows that

$$\begin{aligned}
 & K_{et} + \underbrace{\bar{\mathbf{V}} \cdot \nabla K_e + \bar{w} K_{ez}}_A + \underbrace{\mathbf{V}' \cdot \nabla K_e + w' K_{ez}}_B = \\
 & \underbrace{-\nabla \cdot (\mathbf{V}' p') - (w' p')_z - g \rho' w'}_C + \rho_0 \underbrace{[-\mathbf{V}' \cdot (\mathbf{V}' \cdot \nabla \bar{\mathbf{V}}) - \mathbf{V}' \cdot (w' \bar{\mathbf{V}}_z)]}_D \\
 & + \rho_0 \underbrace{[\mathbf{V}' \cdot (\mathbf{V}' \cdot \nabla \mathbf{V}')] + \mathbf{V}' \cdot (w' \mathbf{V}'_z)]}_E + \underbrace{[A_{MH}(\mathbf{V}' \cdot \nabla^2 \mathbf{V}') + \mathbf{V}' \cdot (A_{MV} \mathbf{V}'_z)]}_F, \tag{4}
 \end{aligned}$$

430 where K_e is EKE, ρ_0 is a mean density, A_{MH} and A_{MV} are horizontal and vertical viscosity
 431 coefficients, subscripts t and z denote partial differentiation with respect to time and the
 432 vertical coordinate, and ∇ is the horizontal gradient operator. Terms A and B represent
 433 the horizontal and vertical advection of EKE by mean and perturbation flows. C is the
 434 pressure work due to eddies (with the aid of the continuity and hydrostatic equations, it
 435 can be rewritten $-\mathbf{V}' \cdot \nabla p'$). The first two terms of C sum to a total divergence, and
 436 so do not provide a net source of EKE. The last term of C represents the conversion of
 437 available eddy potential energy into EKE, a measure of baroclinic instability. D is the
 438 horizontal and vertical deformation work due to the mean velocity fields. The first term of
 439 D, $S_{bt} = -\rho_0 \mathbf{V}' \cdot (\mathbf{V}' \cdot \nabla \bar{\mathbf{V}})$, represents energy conversion of mean kinetic energy (MKE)
 440 to EKE by horizontal deformation work, that is, via barotropic instability. The second

441 term of D describes the conversion of MKE to EKE by vertical deformation work, that is,
 442 by Kelvin-Helmholtz instability. E is the horizontal and vertical deformation work due to
 443 the eddy velocity fields, and term F is the dissipation of EKE due to viscosity (For more
 444 details, see Masina et al. [1999]). We note that horizontal advection due to mean flow
 445 and barotropic energy conversion are two most dominant term in this energy budget.

446 Figure 10 illustrates snapshots of the dominant terms in MOMCR on 15 June, 15
 447 July, and 15 August. The EKE field (left panels) is highly asymmetric with respect to
 448 the equator, with high values north of the equator in the neighborhood of the Southern
 449 Gyre. By far, the largest energy conversion term is S_{bt} (middle-left panels); its values are
 450 positive almost everywhere, indicating the generation of EKE. The eddy advection terms
 451 S_{adv} (middle-right panels) follow the mean path of the Southern Gyre (last column). In
 452 July, S_{adv} has negative values that extend across the equator, suggesting that EKE is
 453 being advected either southward or westward. In August, however, the Southern Gyre
 454 has shifted northward somewhat, and the negative patch no longer crosses the equator.
 455 This advection of the eddies by the Southern Gyre is shown more clearly in Figure 11,
 456 which shows S_{bt} following the trajectory of the mean current associated with the gyre.

457 4.2.2. Excitation of Yanai waves:

458 Yanai waves are generated when the eddies advect to the equator during June and July
 459 (Figure 11). When they do so, they generate a v_{eq} field that has periods and wavelengths
 460 (Figure 12) similar to those of Yanai waves associated with several low-order baroclinic
 461 modes, and they are the waves that are efficiently excited by the eddy forcing. The
 462 reason why the advection of eddies to the equator excites Yanai waves is not clear: A
 463 likely cause is that the quasi-geostrophic eddies become dynamically unbalanced when

464 they are advected onto the equator, and radiate waves to try to regain balance. (An
 465 analogous eddy-generation process is known to occur in the eastern, equatorial Pacific
 466 Ocean, where current instabilities generate tropical instability waves, TIWs, with a period
 467 and wavelength similar to several Yanai waves. Similar to our case, the TIWs excite Yanai
 468 waves that also superpose to form a beam that carries energy downward. See, for example,
 469 Cox [1980] and McCreary [1984]).

470 To explore the properties of eddy-driven LFYW, we obtained an analog test solution
 471 to the LCS model forced by a τ^y field with spatial and temporal scales similar to that of
 472 the eddy forcing in MOM (Solutions LCSED25 and LCSED15). Specifically, τ^y has the
 473 form (3a) with the same $X(x)$ and $Y(y)$, except that $L_x = 10^\circ$ and $x_o = 40^\circ$, and with
 474 $T(t)$ replaced by

$$475 \quad T(x, t) = \sin\left(\pi \frac{t'}{\Delta t}\right) \sin(k_w x' - \sigma_w t') \theta \left[\frac{(\Delta t)^2}{4} - t'^2 \right], \quad (5)$$

476 where $t' = t - t_0$, $t_0 = 1$ June, $\Delta t = 75$ days, $P_w \equiv 2\pi/\sigma_w = 25$ or 15 days, and
 477 $\lambda_w \equiv 2\pi/k_w = -6.7^\circ$. Thus, τ^y is located in a zonally narrow region only 10° wide, lasts
 478 for 75 days from 1 June to 14 August, and propagates westward across the region with a
 479 period P_w and wavelength λ_w . As for the previous tests, because $X(x)T(x, t)$ is limited in
 480 x and t , its Fourier transform has power for wavenumbers and frequencies other than k_w
 481 and σ_w , so that Yanai waves with periods and wavelengths other than P_w and λ_w appear
 482 in the response.

483 Figure 13 (left panels) shows longitude-time plots of surface v_{eq} (top) and vertical sec-
 484 tions of v_{eq} on 15 December (bottom) for Solution LCSED25. In this case, the response
 485 separates into two distinct patterns, suggesting the predominance of two signal types: a
 486 primary response with a period close to $P_w = 25$ days, a wavelength of ~ 1000 km, and

487 faster group velocity (less vertically tilted band), and a secondary one with a wavelength
 488 near $\lambda_w = 6.7^\circ$, a period of ~ 45 days, and slower group velocity (more vertically tilted
 489 band). The division into two clear signals results from the nature of the forcing, which
 490 has its strongest spectral peaks near P_w and λ_w . The primary response consists mostly of
 491 $n = 1$ and 2 baroclinic modes. Because only two modes dominate the response, there is
 492 not a well-formed beam (Figure 13, bottom); rather, the response appears as the super-
 493 position of the two modes. The secondary response is composed of a number of baroclinic
 494 modes and forms a beam with slanted phase lines.

495 The structure of Solution LCSED15 is similar to that of Solution LCSED25, with the
 496 presence of two distinct patterns. In this case, the primary response (period close to
 497 $P_w = 15$ days, and wavelength of ~ 2000 km) is mostly the $n = 1$ mode. The secondary
 498 response (wavelength close to $\lambda_w = 6.7^\circ$) is very weak, because such small wavelengths
 499 occur only for periods $\gtrsim 25$ days, and the forcing has very little power at those periods
 500 when $P_w = 15$ days.

501 Both Solutions LCSED25 and LCSED15 appear to extend much farther than the eddy-
 502 driven v_{eq} field in Solution MOMLP τ . In fact, that extension is an illusion in the plots, a
 503 result of the contour intervals being chosen to emphasize the remote fields. For example,
 504 within the forcing region v_{eq} in Solutions LCSED25 and LCSED15 has peak values of 190
 505 cm/s and 70 cm/s, respectively. If the solutions are scaled so that their peak values are
 506 25 cm/s, as for v_{eq} in Solution MOMLP τ (middle panels of Figures 5 and 7), then the
 507 response outside the forcing region is weak.

5. Summary and discussion

508 Observations and models have shown the presence of Yanai waves in the WEIO in the
 509 20–30-day (LFYWs) and 10–20-day (HFYWs) bands (Figures 1 and 2). We explore the
 510 dynamics of these intraseasonal signals using an ocean general circulation model (MOM)
 511 and a linear, continuously stratified (LCS) model.

512 The MOM control run (Solution MOMCR) is able to simulate the observed WEIO
 513 ASLA field in the 20–30-day band reasonably well (compare Figure 1c with the left panel
 514 of Figure 5). It also simulates well the observed v_{eq} field in the central EIO (Figure 4).
 515 The WEIO response in the two period bands is driven by both winds and instabilities.
 516 The middle-right panels of Figures 5 and 7 show the MOM solution forced by 40-day
 517 low-passed winds (Solution MOMLP τ), so that its response in both period bands is nec-
 518 essarily instability-driven. In contrast, the right panels show the response of the LCS
 519 model (Solution LCSCR), which is completely wind-driven since the LCS model allows
 520 no instabilities. Roughly (since MOM is nonlinear), Solutions MOMLP τ and LCSCR sum
 521 to give Solution MOMCR.

522 Wind-forced Yanai waves are driven almost entirely by equatorial τ^y in the WEIO, as
 523 the other wind forcing, equatorial $(\sigma/\beta)\tau_y^x$, is weak there (Figure 1b; discussion at the end
 524 of Section 2). To investigate wind-driven dynamics, we obtain a suite of test solutions
 525 forced by idealized τ^y fields of the form, $\tau^y = \tau_o X(x)T(t)$, defined in (Eq 3a). In the
 526 interior ocean, the Fourier transform of $X(x)T(t)$, $\mathcal{F}(k, \sigma)$, determines the Yanai waves
 527 that can be excited by τ^y : They are the waves for which the Yanai-wave dispersion curve
 528 (2) intersects the spectral peaks of \mathcal{F} . Because the wind forcing has a period P_w and a large
 529 zonal scale (Figure 3), Yanai waves with periods close to P_w and large wavelengths (the

530 primary response) dominate the interior response (left and middle-right panels of Figures
 531 5 and 7); in addition, there are “secondary” waves with different periods and wavelengths
 532 visible in the plots. At the western boundary, Yanai waves are generated because τ^y forces
 533 flows into (or out of) the boundary. Boundary-generated Yanai waves have no wavelength
 534 constraint, and hence shorter wavelengths can be excited (Figure 5 and 7; right panels).
 535 The interior and boundary responses superpose to produce a complicated pattern with
 536 nodes where they interfere destructively (Figure 5 and 7; middle-left panels).

537 Yanai waves are also forced by instabilities, principally during June/July in a region
 538 offshore from the western boundary (52–55°E). At that time, eddies (defined here as the
 539 deviation from the mean field), caused by barotropic instability of the Southern Gyre,
 540 advect southward to the equator along the eastern flank of the gyre (Figures 10 and 11).
 541 At such times, they generate a westward-propagating, across-equatorial flow field, v_{eq} , that
 542 propagates westward with a wavelength set by the width scale of the eddies (Figure 12).
 543 To investigate the Yanai waves excited by v_{eq} , we force the LCS model with a τ^y field
 544 that has a structure similar to v_{eq} : one that propagates westward with wavelength of
 545 $\lambda_w = 6.7^\circ$, has period of $P_w = 25$ or 15 days, and is confined to a narrow region only
 546 10° wide (see the discussion of Eq. 5). In this case, τ^y excites predominately two packets
 547 of Yanai waves: one with wavelengths close to λ_w , and another with periods close to P_w
 548 (Figure 13). Furthermore, the Yanai-wave response is weak because there are no waves
 549 that fit well with both λ_w and P_w , and as a result the locally-forced response is dominant.

550 LFYWs and HFYWs are generated by the same processes, and hence have similar
 551 structures. Differences in the responses for each period band are due primarily to HFYWs
 552 having longer wavelengths for each baroclinic mode than LFYWs do (Table 3). One result

553 of this difference is that the boundary response in Solution LCSDIF15 is broader than in
554 Solution LCSDIF25, so that the nodes in Solution LCSWB15 are shifted farther offshore
555 than in Solution LCSWB25 (compare middle and right panels of Figures 8 and 9). Another
556 is that the secondary waves in Solution LCSIN15 are longer and more strongly excited
557 than in Solution LCSIN25 (left panels of Figures 8 and 9).

558 For both LFYWs and HFYWs, waves associated with several baroclinic modes can
559 superpose to create beams that carry energy along ray paths both downward as well as
560 eastward (Figures 6, 8, 9, and 13). When the forcing shuts off, the response propagates as
561 a packet of free Yanai waves. Wherever waves associated with different baroclinic modes
562 continue to overlap, the packet continues to propagate vertically, and can reflect off the
563 bottom to return to the surface. Eventually, though, since the waves have different group
564 speeds, they will tend to disperse, showing up as individual modes.

565 There are noteworthy differences between our study and that of Kindle and Thompson
566 [1989]. For one thing, all the variability in their model developed from the instability
567 associated with the Southern Gyre, whereas in our model a significant amount of it is
568 wind-driven. For another, they attributed the eddy forcing of Yanai waves by the Southern
569 Gyre to its northward movement: When the westward current on the southern flank of
570 Southern Gyre in their model crosses the equator, an intense anticyclonic circulation
571 forms to its southeast; they associated this disturbance with the generation of a Yanai
572 waves, a process that continued until the Southern Gyre completely crossed the equator.
573 In contrast, we associate the generation of the Yanai waves with v_{eq} , which is generated
574 when the eddies north of the equator are advected southward across the equator by the
575 mean flow associated with the Southern Gyre; in this case, the generation of Yanai waves

576 stops once the eddies cease to cross the equator. Finally, the longitude-time plots of Kindle
577 and Thompson [1989, see their Plate 1] show that, although the Yanai waves weaken east
578 of the Maldives, they exist almost all the way to the eastern boundary of the basin. This
579 difference is due to their using a $1\frac{1}{2}$ -layer model, which consists of only a single baroclinic
580 mode; therefore, their model does not permit the existence of beams that can carry energy
581 into the deep ocean.

582 In conclusion, we have looked in detail into the processes that excite Yanai waves in
583 the WEIO, finding that they are generated by both wind and eddy forcing. All our
584 results, however, are for 2001 when the Yanai-wave signals are strongest. Furthermore,
585 there is considerable interannual variability in the strength and timing of the monsoons,
586 and this variability can be seen in the oceanic response as well [Meyers and O'Brien,
587 1994]. Nevertheless, monsoon winds blow across the Arabian Sea every year, and they
588 typically generate a Southern Gyre during the summer monsoon[see, for example, Schott
589 and McCreary, 2001]. We expect therefore that the processes identified in this paper are
590 active every year, albeit with different strengths and timings.

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Table 1. List of data sets used to force the MOM4p1 model.

Field	Data source	Frequency	Reference
Wind stress (N/m ²)	QuikSCAT	Daily	www.ifremer.fr/cersat/en/data/data.html
Wind speed (m/s)	NCEP	Daily	Kalnay et al. [1996]
SW rad. down (W/m ²)	NCEP	Daily	Kalnay et al. [1996]
LW rad. down (W/m ²)	NCEP	Daily	Kalnay et al. [1996]
Air temperature (K)	NCEP	Daily	Kalnay et al. [1996]
Spec. humidity (kg/kg)	NCEP	Daily	Kalnay et al. [1996]
Precipitation (m/s)	TRMM	Daily	http://daac.gsfc.nasa.gov/precipitation
River runoff (m/s)	SAGE (UNESCO)	Monthly	Vörösmarty et al. [1996], Papa et al. [2010]
Chlorophyll (mg/m ³)	SeaWiFS	Monthly	http://nomads.gfdl.noaa.gov

From left to right, the table columns list the data type, source, frequency, and a reference or website that describes the data.

Table 2. List of experiments.

Solution	Forcing
MOMCR	MOM control run
MOMLP τ	40-day low-passed τ^x and τ^y
MOM τ^y	full τ^y , 40-day low-passed τ^x
MOM τ^x	full τ^x , 40-day low-passed τ^y
LCSCR	LCS control run
LCS τ^y	full τ^y , 40-day low-passed τ^x
LCS τ^x	full τ^x , 40-day low-passed τ^y
LCSIN q	idealized τ^y in (3a), $x_o = 60^\circ\text{E}$
LCSIN' q	Solution LCSIN q with $x \rightarrow x - 40^\circ$
LCSWB q	idealized τ^y in (3a), $x_o = 20^\circ\text{E}$
LCSDF q	Solution LCSWB q – Solution LCSIN' q
LCSED q	idealized τ^y as discussed near Eq. (5)

For each solution, the first column lists its name and the second column describes how each test differs from the control run. The test runs are separated into groups for MOM runs (top) and LCS runs (bottom). The ending label q is either 25 or 15 days, indicating the period P of the forcing.

Table 3. Values of c_n and λ_n for $n = 1-10$.

Parameter	n									
	1	2	3	4	5	6	7	8	9	10
c_n	264.2	166.9	104.5	74.7	59.7	49.3	41.5	36.8	32.8	29.7
λ_n (10 days)	-16243	5156	1645	953	695	541	437	378	330	294
λ_n (15 days)	-2186	-3484	-89575	3532	1842	1226	901	743	624	541
λ_n (20 days)	-1282	-1533	-2244	-4448	-33380	5730	2530	1749	1309	1053
λ_n (25 days)	-931	-1029	-1241	-1588	-2111	-3225	-7491	129202	6194	3228
λ_n (30 days)	-739	-789	-885	-1018	-1173	-1396	-1757	-2215	-3092	-5004
λ_n (35 days)	-616	-645	-698	-766	-837	-928	-1051	-1175	-1349	-1574
λ_n (40 days)	-529	-548	-581	-621	-661	-708	-768	-824	-895	-976
λ_n (45 days)	-465	-478	-499	-526	-556	-579	-614	-645	-683	-723

Values of c_n (cm/s) correspond to the $N_b(z)$ profile noted in Section 2.2. Wavelengths λ_n (km)

are given by (2) for the periods P listed in parentheses. Negative (positive) values of λ_n indicate that the Yanai wave has westward (eastward) phase velocity.

Figure 1. (a) Plot of the model domain and study region (WEIO). Shading shows the model bathymetry derived from Sindhu et al. [2007], with depth increasing from dark to light grey, and the blue box defines the boundaries of the WEIO. (b) Variance-preserving power spectra ($10^{-4}\text{dyne}^2\text{cm}^{-4}\text{cpd}^{-1}$) of τ^y (black curve) and $5(\sigma/\beta)\tau_y^x$ (red curve) averaged over $45\text{--}60^\circ\text{E}$ along the equator in the WEIO. The spectra are computed for year 2001 and smoothed using a 5-point Parzen window. (c) Longitude-time plot of altimeter ASLA (cm) during 2001 band-passed at 20–30 days. The signals are excited during June–October within the western and central WEIO. Note that We have used fourth order Butterworth filter to filter data from the total signal in all our analysis except stated otherwise.

Figure 2. Variance-preserving spectral map across the WEIO of ASLA ($10^{-2}\text{cm}^2\text{cpd}^{-1}$) from altimeter observations (left) and MOMCR (middle), and of v_{eq} ($10^{-2}\text{cm}^2\text{s}^{-2}\text{cpd}^{-1}$) from MOMCR during 2001 (right). Note that sampling interval for the altimeter is 7 days, whereas for MOM it is every day.

Figure 3. Longitude-time plots of τ^y (dyn cm^{-2}) from QuikSCAT data during 2001 band-passed at 20–30 days (left) and 10–20 days (right).

Figure 4. Near-surface v_{eq} (cm s^{-1}) at 80°E from the RAMA mooring (black curve) and MOMCR (red curve).

Figure 5. Longitude-time plots of fields during 2001 band-passed at 20–30 days, showing ASLA (cm) from MOMCR (a) and surface v_{eq} (cm s^{-1}) from MOMCR (b), MOMLP τ (c), and LCSCR (d).

Figure 6. Vertical sections of v_{eq} (cm s^{-1}) from MOMCR band-passed at 20–30 days on 5 September (left) and band-passed at 10–20 days on July 8 (right). The black curves represent the ray paths for 25 days (left) and 15 days (right).

Figure 7. Longitude-time plots of fields during 2001 band-passed at 10–20 days, showing ASLA (cm) from MOMCR (a) and surface v_{eq} (cm s^{-1}) from MOMCR (b), MOMLP τ (c), and LCSCR (d).

Figure 8. Longitude-time plots of surface v_{eq} (top panels) and vertical sections of v_{eq} on 15 August (bottom panels) for Solutions LCSIN25 (left), LCSDIF25 (middle), and LCSWB25 (right). In each of these solutions secondary responses (S) are visible along with the primary response (P). Note that in LCSIN25, waves with short wavelengths emanate from the edge of the wind forced region (marked by a red arrow). LCSWB25 exhibits nodes (marked by black arrows) due to destructive interference of the boundary and interior response. There is a shadow zone near the western boundary in Solution LCSWB25 (white region in the bottom-right panel; Region A), which cannot be reached by a Yanai beam. A similar shadow zone should exist in Solution LCSDIF25, but does not because that solution is not a perfect measure of the boundary-generated response.

Figure 9. Longitude-time plots of surface v_{eq} (top panels) and vertical sections of v_{eq} on 15 August (bottom panels) for Solutions LCSIN15 (left), LCSDIF15 (middle), and LCSWB15 (right). LCSWB15 exhibits both primary (P) and secondary (S) responses and also shows nodes (marked by black arrows) due to destructive interference of the boundary and interior response. The secondary response is also seen in LCSIN15 and LCSDIF15, but is weak and the tilt is small. There is a shadow zone near the western boundary in Solution LCSWB15 (white region in the bottom right panel; Region A), which cannot be reached by a Yanai beam. A similar shadow zone should exist in Solution LCSDIF15, but does not because that solution is not a perfect measure of the boundary-generated response.

Figure 10. Snapshots of terms on 15 June (top), 15 July (middle), and 15 August (bottom) from the eddy-kinetic-energy budget from MOMLP τ , showing surface eddy kinetic energy (EKE, $\text{gm cm}^{-1} \text{s}^{-2}$; left), barotropic conversion terms (S_{bt} , $10^{-5} \text{ gm s}^{-3}$; middle-left), eddy horizontal advection (S_{adv} , $10^{-5} \text{ gm s}^{-3}$; middle-right), and the surface mean current computed using a 31-day running Parzen window (cm s^{-1} ; right). S_{bt} and S_{adv} are integrated over the top 100 m of the water column.

Figure 11. Snapshots of daily total barotropic energy conversion (S_{bt} , color; $10^{-5} \text{ gm s}^{-3}$) integrated over top 100 m of the water column and the mean surface current (vectors; cm s^{-1}) from Solution MOMLP τ . The mean fields have been computed using a 31-day Parzen window centered on the day for which the mean is shown; the eddy fields were computed as deviations from these means. S_{bt} is advected along the trajectory of the Southern Gyre. Therefore, during June–July, when the Southern Gyre cuts across the equator, the southward advection along its eastern flank brings the eddy field and associated energy to the equator, thereby generating a v_{eq} there.

Figure 12. Snapshots of 30-day high-passed surface current (vectors; cm s^{-1}) and SLA (color; cm) from Solution MOMLP τ . Only current vectors with a magnitude greater than 2.5 cm s^{-1} are plotted. The anomalous flow field along the equator is a mixture of eddies ($x \lesssim 55^\circ\text{W}$) and LFYWs and has a wavelength of $\sim 6^\circ$ (left and right panels) and $\sim 7^\circ$ (middle panel).

Figure 13. Longitude-time plots of surface v_{eq} (top panels) and vertical sections of v_{eq} (bottom panels) for Solutions LCSED25 (left) and LCSED15 (right). The vertical sections are snapshots on 15 December and 15 September for Solutions LCSED25 and LCSED15, respectively. The forcing region is located from $35\text{--}45^\circ\text{E}$ and from 1 June to 14 August, only a small portion of the plot. In both solutions, secondary responses (S) are visible along with the primary response (P).

























