Sources and pathways of intraseasonal meridional kinetic energy in the equatorial Atlantic Ocean

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ABSTRACT

In the equatorial Atlantic Ocean, meridional velocity variability exhibits a pronounced peak on intraseasonal timescales whereas zonal velocity dominantly varies on seasonal to interannual timescales. We focus on the intraseasonal meridional velocity variability away from the near-surface layer, its source regions and its pathways into the deep ocean. This deep intraseasonal velocity variability plays a key role in equatorial dynamics as it is an important energy source for the deep equatorial circulation. The results are based on the output of a high-resolution ocean model revealing intraseasonal energy levels along the equator at all depths that are in good agreement with shipboard and moored velocity data. Spectral analyses reveal a pronounced signal of intraseasonal Yanai waves with westward phase velocities and zonal wavelengths longer than 450 km. Different sources and characteristics of these Yanai waves are identified: near the surface between 40°W and 10°W low-baroclinic-mode Yanai waves with periods of around 30 days are exited. These waves have a strong seasonal cycle with a maximum in August. High-frequency Yanai waves (10–20-day period) are excited at the surface east of 10°W. In the region between the North Brazil Current and the Equatorial Undercurrent high-baroclinic-mode Yanai waves with periods between 30 and 40 days are generated. Yanai waves with longer periods (40-80 days) are shed from the Deep Western Boundary Current. The Yanai wave energy is carried along beams east- and downward thus explaining differences in strength, structure and periodicity of the meridional intraseasonal variability in the equatorial Atlantic Ocean.

SIGNIFICANCE STATEMENT

Past studies show that intraseasonal meridional kinetic energy is important for the deep equatorial circulation. However, numerical studies use intraseasonal variability with varying characteristics to investigate the formation and maintenance of the deep equatorial circulation. This is partly due to sparse observations at depth that are limited to single locations. The present study investigates intraseasonal meridional kinetic energy in the equatorial Atlantic Ocean in a high-resolution ocean model that is tested against available shipboard and moored observations. We analyze the spatial and temporal distribution as well as the baroclinic structure of intraseasonal variability. Using the model, we identify different sources and pathways of intraseasonal energy in the deep equatorial Atlantic. With this we offer a groundwork for further studies on the formation and maintenance of the deep equatorial circulation.
1. Introduction

The equatorial circulation of the Atlantic Ocean is characterized by a complex set of energetic zonal currents which vary on seasonal to interannual timescales (Brandt et al., 2016). However, meridional velocity variability exceeds zonal velocity variability at intraseasonal timescales with enhanced variability in the period range from 10 to 50 days (Athie & Marin, 2008; Bunge et al., 2008; Tuchen et al., 2018).

In the upper ocean, tropical instability waves (TIWs) are the prevalent source of intraseasonal variability (Weisberg & Weingartner, 1988; Jochum et al., 2004; Athie & Marin, 2008; von Schuckmann et al., 2008; Perez et al., 2012; Tuchen et al., 2018). TIWs can be detected from satellite due to their signature in sea surface temperature (SST), sea surface salinity, and sea level anomaly (e.g., Olivier et al., 2020). In the Atlantic, TIWs typically have periods between 20 and 60 days, their zonal wavelengths range from 600-1200 km and they propagate westward with a phase speed of 20-60 cm/s (Weisberg & Weingartner, 1988; Athie & Marin, 2008; von Schuckmann et al., 2008). Additional to the TIWs at the surface, sea surface height variability exists in the equatorial Pacific on intraseasonal timescales that can be explained by low-baroclinic-mode/barotropic Rossby waves (Farrar and Weller, 2006; Farrar, 2011). These waves propagate away from the equator and can be traced to at least 30°N (Farrar et al., 2021).

Baroclinic and barotropic instability is responsible for the formation of TIWs (Grodsky et al., 2005; von Schuckmann et al., 2008). North of the equator, it is the barotropic instability due to horizontal shear between the Equatorial Undercurrent (EUC) and the northern branch of the South Equatorial Current (nSEC) as well as between the nSEC and the North Equatorial Countercurrent. Additionally, baroclinic instability due to vertical shear within the nSEC contributes. South of the equator, baroclinic instability originating from the vertical shear of the central branch of the South Equatorial Current generate TIWs as well (von Schuckmann et al., 2008).

TIWs undergo a strong seasonal modulation that originates from the seasonal cycle of equatorial winds. In June/July the southeasterly trade winds intensify at the equator. With it, the near-surface circulation, as well as the instability of the zonal currents, intensifies accordingly, resulting in a peak of TIW energy in boreal summer (Weisberg & Weingartner, 1988; Athie & Marin, 2008; von Schuckmann et al., 2008; Tuchen et al., 2018).
On shorter timescales than those of TIWs, another source of intraseasonal variability is biweekly wind fluctuations that directly excite biweekly Yanai waves with long zonal wavelengths (Athie and Marin, 2008). The signal of the meridional wind fluctuations is present all year long. However, during boreal spring and summer, when a strong SST front develops in the eastern equatorial Atlantic, the air-sea interactions between atmospheric wind forcing and the feedback due to advection of warm and cold SST anomalies supports a substantial part of the biweekly variability (de Coëtlogon et al. 2010).

Enhanced intraseasonal variability is also observed at greater depths (Bunge et al., 2008; Tuchen et al., 2018; Zhang et al., 2020)). Tuchen et al. (2018) analyzed meridional velocity data from a mooring site at the equator at 23°W. Their results suggest that TIWs excite Yanai waves of mode 2-5 with 30 to 40-day period, which propagate their energy down- and eastward. This intraseasonal energy can be traced from the surface down to about 2000 m. Nevertheless, the study also found signals of intraseasonal Yanai waves that cannot be explained by sources of wave energy at the surface. Instead, the Deep Western Boundary Current (DWBC) was suggested as another source region for the generation of intraseasonal Yanai waves. From moored observations in the equatorial Pacific Ocean, Zhang et al. (2020) found intraseasonal variability that can be associated with short Rossby waves with westward phase speed which propagate their energy southeastward and downward.

The energy of intraseasonal variability is thought to play an important role in equatorial dynamics. At depth, the so-called deep equatorial circulation (DEC) exists. The DEC consists of high-baroclinic-mode equatorial deep jets (EDJs) and low-baroclinic-mode or barotropic equatorial intermediate current system (EICS) also called extra-equatorial jets (EEJs) (Ascani et al. 2010, 2015; Cravatte et al. 2012, 2017; Ménèsguen et al. 2019; Delpech et al. 2020a, 2020b). EDJs are vertically stacked jets with a periodicity of about 4.5 years (Johnson & Zhang, 2003; Youngs & Johnson, 2015; Claus et al., 2016). EDJs are the dominant source of interannual variability at intermediate depths in the equatorial Atlantic and they are suggested to affect climate variability (Brandt et al., 2011). Model studies suggest that the deep equatorial intraseasonal variability (DEIV) plays an important role in the formation of the DEC (d’Orgeville et al., 2007; Hua et al., 2008; Ascani et al., 2010, 2015).

Based on the results of numerical simulations, Ascani et al. (2010) found that the EICS can be generated by strong intraseasonal Yanai waves through self-advection and dissipation west of the Yanai beam that carries energy from the surface east- and downward. The shear...
instability of short, low-baroclinic-mode Yanai waves generated near the western boundary of the equatorial ocean was found to be responsible for the formation of EDJs (d’Orgeville et al., 2007; Hua et al., 2008). In contrast, in the simulations of Ascani et al. (2015) DEIV is generated in the upper ocean, propagates downward and rectifies into the low-frequency EDJs. Their solution largely depends on the structure and amplitude of the DEIV. Greatbatch et al. (2018) showed how intraseasonal waves can maintain EDJs against dissipation: a zonal jet of smaller meridional scale distorts an intraseasonal wave of larger meridional scale such that it leads to a convergence of the meridional flux of zonal momentum. Recently, Bastin et al. (2020) showed in an idealized model setup that EDJs are generated when the model is driven only by the convergence of the meridional flux of intraseasonal zonal momentum. This underlines the importance of understanding the DEIV in order to realistically model the EDJs. Furthermore, Delpech et al. (2020b) proposed that intraseasonal barotropic Rossby waves are important for the generation of the EEJs.

Intraseasonal variability in the deep equatorial Atlantic Ocean has been analyzed from moored velocity observations which are limited to single locations along the equator (Bunge et al., 2008; Tuchen et al., 2018). We want to extend these analyses by making use of a high-resolution ocean model that is tested against an extended dataset of shipboard and moored observations. The goal of this study is to characterize the deep intraseasonal variability along the equator in the Atlantic Ocean.

This study is structured as follows: Section 2 introduces the model as well as the observational data used to validate the model. Section 3 presents the methods. Similarities and differences between observations and model results are described in section 4. In section 5 we describe the characteristics of intraseasonal variability. We then discuss energy sources and pathways of intraseasonal variability in section 6. Section 7 provides a summary and a discussion.

2. Data

a. Model

To analyze intraseasonal variability in the equatorial Atlantic Ocean we use the output of a numerical ocean model. We utilize the VIKING20X model which consists of a global host model with a high-resolution nest of the Atlantic Ocean. For a detailed description of the
model the reader is referred to Biastoch et al. (2021) where the configuration we use is called VIKING20X-JRA-short.

The simulation is based on the “Nucleus for European Modelling of the Ocean” (NEMO, Madec, 2016) code version 3.6. The global host model ORCA025 has a horizontal resolution of 0.25°. It has 46 depth layers whose vertical extents range from 6 m at the surface to ~250 m at depth. The high-resolution nest covers the Atlantic Ocean from 33.5°S to ~65°N. Its horizontal resolution is 1/20° whereas the vertical resolution is the same as the global ORCA025 configuration.

The atmospheric forcing for the ocean-only model is provided by the JRA-55-do dataset version 1.4 (Tsujino et al., 2018). The atmospheric forcing field is available on a 0.5° horizontal grid at 3-hourly temporal resolution. It is interpolated to the horizontal resolution of the host and the nest grid. The model configuration we use is integrated between the years 1980 and 2018 (38 years). The output is saved at a daily resolution.

b. Observation

The model output is compared to observations to validate the representation of intraseasonal variability. To do so we use the moored velocity observations at six different locations along the equator. An overview of the data availability as a function of time and depth at each mooring location can be found in the appendix (Fig. A.1). Note that except for the mooring at 0°N, 23°W the measurements at depth are relatively sparse.

Single-point current meters installed in two moorings located at the continental slope of the western boundary at 0.4°N, 44.25°W and 0.2°N, 44.3°W collected data from October 1990 to September 1991 and October 1992 to March 1994 respectively. In Fischer & Schott (1997) a detailed description of the dataset can be found. Note that additionally an acoustic Doppler current profiler (ADCP) was installed at the mooring at 0.2°N, 44.3°W. However, as it only recorded data for 110 days, we did not use this data for the present study.

Four moored single-point current meters collected data at 0°N, 36°W between October 1992 and May 1994 (Send et al., 2002). At 0°N, 35°W velocity data were recorded from August 2004 to June 2006 using an ADCP and single-point current meters (Hormann & Brandt, 2009).

The mooring dataset at 0°N, 23°W we use is an update of Tuchen et al. (2018) extended by three mooring periods. The updated moored velocity dataset additionally includes near-
surface velocity observations from a single-point current meter installed at 10 to 12 m as part of the PIRATA buoy at 0°N, 23°W (Tuchen et al., 2022). The mooring provides velocity measurements from December 2001 to June 2021 (apart from a period from December 2002 to February 2004 when no mooring was in place).

From May 2003 to March 2019 velocity was measured at 0°N, 10°W using ADCPs and single-point current meter. However, the mooring was not consecutively installed leading to data gaps of up to two years. For the data measured prior to June 2005 the reader is referred to Bunge et al. (2008). The data afterwards are described in Brandt et al. (2021). At 0°N, 0°E velocity data were recorded by ADCPs between October 2007 and May 2011 (Johns et al., 2014).

We also compare snapshots of meridional velocity from the model to an observed equatorial section of meridional velocity. This section was measured along the equator between September 29, 2019 and October 22, 2019 from east to west in the course of the research cruise M158 conducted by the research vessel Meteor (Brandt et al., 2019). The velocity profiles were obtained from ADCP measurements with two instruments attached to the CTD rosette and lowered from the surface to 6000m or close to the bottom. The horizontal resolution of the velocity profiles is 1° from 5°E to 35°W. West of 35°W the resolution increases towards the western boundary.

3. Methods

a) Lanczos filter

In the course of this study time series of velocity are filtered for variability on intraseasonal timescales which we define as variability with periods shorter than 80 days. For that purpose, a cosine-Lanczos filter as formulated by Thomson & Emery (2014) with a cutoff frequency of 1/80 days⁻¹ and a window length of 185 days is used.

b) Modal decomposition

The concept of standing vertical normal modes is an appealing way to describe the vertical structure of the ocean (Gill, 1982). It is especially useful when describing linear wave dynamics. Wave characteristics derived for the homogeneous ocean can be transferred to the stratified ocean using normal modes. For that, we fit vertical profiles of horizontal velocity to a set of vertical normal modes. Note that for a wave to consist of one baroclinic mode only the wave has to reflect from the bottom and return to the surface within the forcing region.
thus creating a vertically standing wave. The waves discussed in this study mostly do not fulfill this requirement. However, the velocity field can still be decomposed into vertical normal modes. To fully describe the waves, a spectrum of vertical normal modes is needed. The modal decomposition then helps in characterizing these waves as it reveals what part of the vertical mode spectrum is elevated for the respective waves (Philander, 1978).

The normal modes are based on a mean buoyancy frequency profile \( N^2 \) acquired by averaging \( N^2 \) between 5°N and 5°S. The eigenvalue problem

\[
\frac{d}{dz} \left( \frac{1}{N^2} \frac{d\hat{\rho}}{dz} \right) + \frac{1}{c_n^2} = 0
\]  

(1)

can be derived from the governing equations for an incompressible, inviscid fluid in a non-rotating system linearized about a state of rest. Solving the eigenvalue problem for a vertical mode \( n \) gives the structure function \( \hat{\rho} \) (Gill, 1982) where \( c_n \) is the gravity wave speed for the vertical mode \( n \). The eigenvalue problem has two boundary conditions:

\[
\hat{\rho} + \frac{g}{N^2} \frac{d\hat{\rho}}{dz} = 0 \text{ at } z = 0 \text{ and } \frac{1}{N^2} \frac{d\hat{\rho}}{dz} = 0 \text{ at } z = -H
\]

where \( H \) is the full water depth of the buoyancy frequency profile (here we use \( H = 5875 \) m). We normalize the amplitude of each vertical structure function such that:

\[
\int_{-H}^{0} \hat{\rho}_n \hat{\rho}_m \, dz = \delta_{nm} H
\]

(2)

where \( \delta_{nm} \) is the Kronecker delta.

In a next step, the horizontal velocities are fit to the structure functions. The horizontal velocities at one location can be expressed as \( \vec{u}(z, t) = A \vec{\hat{u}}(n, t) \) where

\[
A_{ij} = \hat{\rho}_j \left( z_i \right) / \left( g \rho_0 \right) \in \mathbb{R}^{Z \times N}
\]

with \( Z \) being the number of vertical levels and \( N \) the number of normal modes. In our case \( N \) is 16 as we consider the first 16 vertical normal modes (including the barotropic mode). The modal velocities are then \( \vec{u}(n, t) = A^{-1} \vec{\hat{u}}(z, t) \) with \( A^{-1} \) being the inverse of \( A \). One issue of this fitting is the topography. The modal structures are calculated using a full depth \( N^2 \) profile. However, the velocity profiles reach different depths depending on their location. Thus, we have to perform the fitting in presence of topography. To resolve this issue, we cut the structure function to the depth of the respective locations. Consequently, the matrix \( A \) is not invertible as there is no unique solution to the linear

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system. To find the best solution for our application we calculate the solution using a ridge regression. The ridge regression works by minimizing a penalized residual sum of squares:

$$\min_{\hat{u}} \left\| A\hat{u} - u \right\|^2 + \alpha \left\| \hat{u} \right\|^2$$  \hspace{1cm} (3)

$\alpha > 0$ is a hyperparameter for tuning the solution. Note that this system can also be solved using an ordinary least square method. However, as the system is highly underconstrained in shallow locations, compensating effects lead to unrealistic high modal amplitudes. By using the ridge regression, we can prohibit this behavior by choosing $\alpha$ accordingly. Choosing $\alpha$ is an optimization problem: For $\alpha \to 0$ we have little energy loss when transforming the modal velocities back to physical space but at the same time we have unrealistic high modal velocities. For $\alpha \to \infty$ the ridge regression dampens the system and we have high energy loss when transforming the modal velocities back to space. To find the optimal $\alpha$ we use leave-one-out cross-validation, that is we separate the data into training and test data by marking a single sample as test data, leading to as many test sets as samples. The ridge regression is performed on all of these training sets and the $\alpha$ with the highest average score $R^2 = \left( 1 - \frac{x}{y} \right)$, computed on the corresponding test set, is chosen where $x$ is the residual sum of squares and $y$ the total sum of squares. We find the optimal $\alpha$ for three locations with different depths: 29 (shallowest location away from the boundaries), 36 and 46 depth levels. The optimal $\alpha$ are 24.6, 23.1, and 21.5 with a score of 0.79, 0.8, and 0.81 respectively. As $\alpha$ does not vary much between the different depths and as the compensation effects are largest for shallow profiles, we choose $\alpha$ to be 24.6 for the fitting at all locations. Note that the fitting procedure using the ridge regression still leads to an energy loss of $\sim$10% when transforming the velocities back to physical space.

c) Spectral Analysis

To calculate spectral estimate of velocity time series, we use the method of Welch (1972). This method improves the spectral estimator by sectioning the time series using a Hanning window. The spectrum is then calculated for each section before averaging them. For the spectra of modeled meridional velocity at each grid point along the equator we use a window length of 5 years with a window overlap of 50%.

To calculate the spectrum of gappy moored velocity data at 0°N, 23°W we apply the method of Lomb-Scargle. For details about this method see Tuchen et al. (2018).
To analyze the dominant periodicity of intraseasonal meridional velocity variability of observed moored velocity data we calculate spectral estimates and determine the period with maximum energy in the spectra. To calculate the spectra of meridional velocity at all mooring sites introduced in section 2b except at 0°N, 23°W, we use the Welch method with a window length of 1 year and an overlap of 50%. Note that the moored ADCP data at 0°N, 10°W have gaps. Here we calculated the spectra for each mooring period individually. Afterwards, we determine the period with maximum energy in the spectra on intraseasonal timescales. We average the periods over all mooring periods with a weighted mean based on the length of the mooring periods.

1) Frequency-wavenumber analysis

We analyze the equatorial wave field by calculating frequency-wavenumber spectra of modal velocity. To do so, we calculate spectra as a function of zonal wavenumber and frequency using a two-dimensional Fourier transform. We compare the spectra to theoretical dispersion relations of free equatorial waves. The wave ansatz of the theoretical dispersion relations has an opposing sign ($e^{i(kx-\omega t)}$). Thus, the two-dimensional Fourier transform is defined accordingly:

$$X_{kl} = \sum_{m=0}^{M-1} \sum_{n=0}^{N-1} x_{mn} \exp\left(-2\pi i \left(\frac{ml}{M} - \frac{nk}{N}\right)\right) \text{ for } k = 0, ..., N - 1, \ l = 0, ..., M - 1$$ (4)

where $m$ is the spatial dimension and $n$ the temporal dimension.

Considering the conservation of energy while transforming the signal and the symmetry of real-valued Fourier transforms, we derive

$$\hat{S} = \frac{2}{NM} |X_{kl}|^2$$ (5)

for the frequency-wavenumber spectral estimate.

Frequency-wavenumber spectra are calculated for modal velocities at the equator (longitude-time sections). To reduce the variance of the spectral estimator, the spectra are calculated on the 1/20° grid between 0.25°S and 0.25°N individually and averaged afterward.

2) Yanai wave filter

The frequency-wavenumber spectra of the meridional modal velocity of all modes show a dominant signal of Yanai waves with westward phase velocity (see section 5a). To analyze the Yanai waves in greater detail we filter the modal velocities for their signal. This filtering...
strategy follows the method of Farrar (2011). The basic idea of the filter is to isolate frequencies and wavenumbers of interest and transform them back into the longitude-time domain. Thus, frequency-wavenumber spectra pose as a basis for the Yanai-wave filter. For our case we define a box around the theoretical dispersion relation of Yanai waves of the respective mode. All frequencies in a distance of 2.5 cycles per year of the theoretical dispersion relation are selected. Additionally, only wavenumbers between -0.25 and -0.01 cycles per degree are chosen. After defining the box, the spectra outside of it is set to zero. Subsequently, the inverse discrete Fourier transform is calculated to transform the spectra back to physical space. In analogy to calculating the frequency-wavenumber spectra, the Yanai-wave filtered modal velocities are calculated from spectra between 0.25°S and 0.25°N separately and averaged afterwards to increase statistical robustness. Farrar (2011) applies a two-dimensional Tukey window to the spectrum inside the box to reduce filter sidelobes. For our results the usage of the window function did not alter the results. We therefore abstain from using the Tukey window.

4. Model validation

The moored observations at 0°N, 23°W show meridional velocity signals with upward phase propagation (Tuchen et al., 2018), which imply, following linear wave theory, a downward energy propagation (Gill, 1982). These signals have been associated with intraseasonal Yanai waves (Ascani et al., 2010; Tuchen et al., 2018). The theoretical work of McCreary (1984) derived the slope of vertically propagating equatorial waves. The angle of propagation, \( \gamma(z) \), depends only on the angular wave frequency, \( \omega = 2\pi f \), the buoyancy frequency profile, \( N(z) \), and on the meridional mode, \( l \):

\[
\tan(\gamma) = \frac{c_{gz}}{c_{gx}} = \pm \frac{(2l+1)\omega}{N} \tag{6}
\]

with \( c_{gx} \) and \( c_{gz} \) being the zonal and vertical group velocities, \( l = -1 \) represents Kelvin waves, \( l = 0 \) Yanai waves and \( l = 1, 2, 3, \ldots \) Rossby and gravity waves. Note that within the framework of vertical normal mode decomposition, vertical wave propagation can only exist by a superposition of multiple vertical modes. The group velocity of such waves depends on the dominant mode contributing to its formation (Tuchen et al., 2018).
We evaluate how well intraseasonal variability is simulated by comparing the model output to observations. Previous studies have shown that, at the equator, the kinetic energy in the intraseasonal frequency band is predominantly due to the meridional velocity variability whereas zonal velocity variability is found mostly on longer, seasonal to interannual timescales (Athie & Marin, 2008; Bunge et al., 2008; Ascani et al., 2015; Tuchen et al., 2018). We therefore focus on the meridional velocity component to validate the modeled intraseasonal variability.

In a first step, we qualitatively compare an observed equatorial section of meridional velocity to snapshots of meridional velocity from the model output (Fig. 1). Note that the shipboard velocity section was measured in the course of 23 days. Measured velocity structures can thus appear distorted. In the observed velocity field, waves are identifiable that propagate their energy from the upper ocean at the western boundary east- and downward (indicated by black arrows). Below 1000-m depth the signals lose coherence and become weaker. The three snapshots of meridional velocity from the model output taken in the same season as the observed section exhibit similar structures in the upper ocean (Fig. 1 b-d). In 2000 and 1982 the amplitude of the wave structure is strong and compares well with the amplitude in the observations. In 2007, the meridional velocity field in the upper ocean is, instead, much weaker and noisier. In the western part of the basin, observed velocity structures of large vertical extent are recognizable below 1000 m. Such structures are also visible in the snapshot of modeled meridional velocity in 1982. In 2000 and 2007 the amplitude of meridional velocity is weaker and no clear structures are recognizable. Overall, the observed meridional velocity section and three snapshots of the modeled meridional velocity compare well in structure and amplitude. The three snapshots furthermore suggest a strong year-to-year variability of meridional velocity fluctuations.
Fig 1: (a) Meridional velocity in [m s\(^{-1}\)] along the equator, measured during research cruise M158 of RV Meteor between September 29, 2019 and October 22, 2019 from east to west. Snapshots of modeled meridional velocity from October 10 of the years (b) 1982, (c) 2000, and (d) 2007. Dashed black lines mark wave fronts. Solid black arrows show the proposed energy propagation path.

In a next step, we use moored velocity data at six different locations to further validate the model’s representation of intraseasonal meridional velocity variability. Note that the data coverage is mostly sparse and that the amount of data varies strongly between the different mooring locations (see Fig. A.1 in the Appendix). Fig. 2a shows the modeled equatorial section of mean intraseasonal meridional kinetic energy averaged over the whole model run. The observed mean intraseasonal meridional kinetic energy at the individual mooring locations is superimposed. Overall, the modeled amplitudes fit well to the observed amplitudes. The upper ocean exhibits high intraseasonal energy levels declining with depth. Elevated intraseasonal energy levels are also found in both model and observations around the DWBC. Comparing the profiles of modeled and observed mean intraseasonal meridional kinetic energy at the mooring location individually allows for a closer inspection of the amplitude (Fig. B.1, Appendix). These reveal shortcomings of the model which include missing subsurface energy maxima at 0°N, 35°W; 0°N, 10°W; and 0°N, 0°E. Additionally, the intraseasonal energy levels tend to be stronger in the observation.
The periods with maximum energy in the meridional velocity spectra on intraseasonal timescales both in the model and in moored velocity data is presented in Fig 2b. In general, the periodicity of modeled and observed meridional velocity variability agree well. The model shows that periods above 50 days dominate the variability in the western part at depth. This is also visible from the deep moored current meters at the two moorings at ~44°W and at 0°N, 36°W. Meridional velocity variability with periods between 30-40 days dominate in the moored velocity data at 0°N, 35°W; 0°N, 23°W; and 0°N, 10°W. The modeled meridional velocity variability shows similar periodicity at these locations. At 0°N, 23°W below 3000 m observed meridional velocity varies dominantly on periods between 10-20 days. A similar signal is also visible in the model. However, here it is restricted to areas closer to the topography. Low vertical resolution in the model could be a reason for these differences. At the surface east of 10°W the modeled meridional velocity varies predominantly on short periods (10-20 days). This signal is also visible in the moored velocity data. However, in the observations the signal is restricted to the uppermost ADCP bins.

![Diagram](image)

Fig. 2: (a) Equatorial section of mean intraseasonal kinetic energy of meridional velocity in [m² s⁻²]. Intraseasonal is defined as variability with periods shorter than 80 days. Meridional kinetic energy is averaged over the whole model run. Grey line shows the minimum water depth between 1°S and 1°N in the longitude range of the Mid-Atlantic Ridge. (b) Period at the maximum energy of meridional velocity spectra in the intraseasonal frequency band. In (a) and (b) colored area within black boxes show results of observed moored velocity data.

As the mooring at 0°N, 23°W provides the most extensive moored dataset in the equatorial Atlantic we use these data to further evaluate the model’s ability to simulate...
intraseasonal meridional velocity variability at the equator. To analyze the representation of TIWs in the model, the meridional velocities between 0 and 50 m are filtered for variability on intraseasonal timescales with periods shorter than 80 days. Afterwards, the kinetic energy is calculated and averaged over depth to derive the seasonal cycle of intraseasonal meridional kinetic energy (Fig. 3). In both model and observations, a maximum in August and a minimum in April is visible. In the observations, a secondary maximum is also found in January which is absent in the model. Additionally, the amplitude of the seasonal cycle is slightly weaker in the model than in the observations (about 12% in August). The distribution around the mean is comparable. In August, the 90th percentile in both model and observation is about two times the mean value.

![Figure 3: The seasonal cycle of intraseasonal meridional kinetic energy in [m² s⁻²] at 0°N, 23°W in (a) the model and (b) observations averaged between 0 and 50 m. The red line shows the monthly mean, the blue line the monthly median. The light (dark) grey colored area shows the 10th-90th (25th-75th) percentile range derived from the distribution of monthly means.](image)

The representation of intraseasonal variability at depth is investigated by calculating a spectrum of meridional velocity at 0°N, 23°W. The periodogram of meridional velocity observed at 0°N, 23°W reveals variance at the surface at all frequencies (Fig. 4a). There is also a distinct signal between the surface and 3000 m with frequencies between 6 and 15 cycles per year (periods of 25-60 days). Additionally, at depths between 3000 and 3500 m a low frequency signal (1-3 cycles per year) as well as a high-frequency signal (25 cycles per...
year) is visible. The spectrum of the modeled meridional velocity is comparable in structure to the observations (Fig. 4b). The signal between the surface and 3000 m with frequencies between 6 and 15 cycles per year is present. The low-frequency signal between 3000 and 3500 m is also visible, however, the high-frequency signal around 25 cycles per year in this depth range is missing. The depth averages of the spectra between 100 and 3000 m show that the strength of the variance is comparable between model and observations on intraseasonal timescales (Fig. 4c). However, the model shows more variance on lower frequencies (~8 cycles per year) whereas the variance in the observations show more variability on higher frequencies (~12 cycles per year). Additionally, the model shows a very strong peak at 11 cycles per year which is not present in the observations. On longer timescales than intraseasonal periods, the observations exhibit stronger variance. Here, a distinct peak at a period of 90 days is visible which is absent in the model. In contrast, the annual and semiannual peaks that are found in the model results are not present in the observations.

Fig 4: Lomb-Scargle periodogram of meridional velocities at 0°N, 23°W in (a) the model and (b) observations. Colors represent variance in [m² s⁻²]. Too sparsely sampled depths are marked grey. (c) Depth average between 100 and 3000 m of model (red) and observation (blue).

5. Characteristics of intraseasonal variability

a) Spectral structure
We calculate a frequency-wavenumber spectrum of the meridional velocity component of vertical modes to analyze the structure of intraseasonal variability along the equator. Here, we discuss the spectrum of mode-2 meridional velocity in detail as mode 2 exhibits the strongest Yanai-wave signal (see section 5b). The frequency-wavenumber spectrum shows elevated energy levels around the theoretical dispersion relation of Yanai waves (Fig. 5). The strongest signal is found between zonal wavenumbers of -0.25 and 0 cycles per degree (wavelengths larger than 450 km). Yanai waves with negative wavenumbers exhibit westward phase velocity, their group velocity and thus also their energy propagation is always eastward. Intraseasonal Yanai waves with small negative wavenumbers represent the dominant variability not only for mode 2, but in the frequency-wavenumber spectra of the barotropic mode and the first 15 vertical normal modes (not shown). There are also elevated levels of energy visible between the theoretical dispersion relations of Yanai waves and Rossby waves at around 10 cycles per year. These signals can be associated with TIWs (Farrar, 2011).

Fig. 5: The frequency-wavenumber spectra of modal amplitudes of meridional velocity of vertical mode 2 along the equator in [m$^2$ s$^{-2}$]. Note the logarithmic color scale. Black lines display the theoretical dispersion relations of equatorial waves. The lines show the dispersion relation of Yanai waves and meridional mode 2 Rossby waves (from top to bottom). Note that only asymmetric ($l = 2,4,6,...$) Rossby waves have a meridional velocity component at the equator.
b) Spatial distribution

In the upper ocean, the highest intraseasonal energy levels are visible near the western boundary (Fig. 2a). The maximum here is found at 42°W at 135 m depth. This region is located in the shear region between the North Brazil Current (NBC) and the Equatorial Undercurrent (EUC). From this region, a beam of enhanced energy is detectable reaching from the upper ocean to the ocean bottom at about 20°W. Additionally, high energy levels are also found at the western boundary between 1100 and 2500 m depth in the vicinity of the DWBC. In the eastern part of the basin, intraseasonal energy levels are elevated at the surface and decrease with depth. Additionally, high energy levels are found close to the topography at the eastern boundary. In contrast, low energy levels are found close to the topography between 15 and 25°W. Analyzing the topography in this region reveals shallow topographic features just north and south of the equator (Fig. 2a, grey line). These features are an obstacle for downward propagating intraseasonal equatorial waves and can thus explain the low energy levels here.

To analyze the dominant intraseasonal period along the equator we consider, at each grid point along the equator, the period with maximum energy in the spectra on intraseasonal timescales (Fig. 2b). Periods from 30-50 days dominate over large areas along the equator. In the region of the energetic beam from the upper ocean near the western boundary to the bottom at around 20°W and east of it, periods between 30 and 40 days dominate. In the deep western part of the basin, variability with periods between 50 and 60 days is prevailing. At the surface east of 10°W, variability with periods between 10 and 20 days have the strongest signal. Close to the topography at the western boundary as well as at the Mid-Atlantic Ridge short periods (10-20 days) dominate. These signals likely can be explained by wave-topography interaction that leads to frequency shifts to higher frequencies.

The frequency-wavenumber analysis shows a clear signal of Yanai waves with negative wavenumbers along the equator (Fig. 4). To analyze the spatial distribution of the Yanai waves, we filter the modal velocities for the velocity field of these waves in space and time (see section 3c2). Note again that the waves discussed in this study do not consist of a single baroclinic mode. Discussing the baroclinicity of these waves focuses on discussing what part of the vertical mode spectrum is enhanced and thus gives information on the vertical structure of the waves.
The mean kinetic energy of the Yanai wave velocities reveals differences in strength and spatial distribution between the modes (Fig. 6). On average, mode-2 Yanai waves are the most energetic ones (Fig. 6b). They are predominantly active between the western boundary and about 10°W with the maximum kinetic energy found at 26°W. Mode-3 Yanai waves also have their maximum in kinetic energy located at about 26°W. However, they are much weaker than mode-2 Yanai waves. Mode-4, mode-6, mode-7 and mode-9 Yanai waves exhibit elevated activity close to the western boundary. Here, mode-7 and mode 9 Yanai waves are especially energetic, their energy levels rapidly drop east of 20°W. For higher modes the kinetic energy levels are comparably low. Overall, the spatial distribution of the kinetic energy of Yanai waves suggests different sources for the Yanai waves of different baroclinic structure.

![Fig. 6: The mean meridional kinetic energy of filtered modal velocity in [m² s⁻²] as a function of (a) longitude for vertical normal modes 0 to 12. Mode 0 represents the barotropic mode. (b) shows the mean meridional kinetic energy from (a) zonally averaged over all longitudes. Velocities are filtered for Yanai waves with zonal wavenumbers between -0.25 and -0.01 cycles per degree.](image)

The spatial structure of kinetic energy of Yanai waves suggests different characteristics between the modes. Low-mode Yanai waves (mode 2-3) are most active in other regions than high-mode Yanai waves (mode 6-9). We further discuss these differences by considering Hovmoeller plots of the velocity fields of the most energetic low-mode Yanai wave (mode 2) and the most energetic high-mode Yanai wave (mode 9) (Fig. 7). Mode-2 Yanai wave activity between 1995 and 2000 is found predominantly between the western boundary and

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10°W. The wave packets propagate eastward over long distances before dissipating. Especially energetic Yanai waves are present in 1996. Here, Yanai wave energy propagates across the whole width of the Atlantic. The velocity field of mode-9 Yanai waves appears to be noisier than that of mode-2 Yanai waves. In general, packets of mode-9 Yanai waves have a smaller zonal coherence. They also do not propagate zonally as far as mode-2 Yanai waves. Energetic mode-9 Yanai waves are found between the western boundary and 20°W.

Fig. 7: Hovmoeller plot of filtered meridional velocity of (a) mode 2 and (b) mode 9. Velocities are filtered for Yanai waves with zonal wavenumbers between -0.25 and -0.01 cycles per degree. Black dashed line illustrates westward phase velocities and black solid line eastward group velocities of a Yanai wave with a period of (a) 30 days and (b) 36 days.

c) Temporal variability

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Fig. 8: The seasonal cycle of meridional kinetic energy of modal velocity that is filtered for Yanai waves with zonal wavenumbers between -0.25 and -0.01 cycles per degree. The modal velocity is zonally averaged along the equator.

To analyze the temporal variability of Yanai waves we calculate the seasonal cycle of kinetic energy of the velocity field of Yanai waves with zonal wavenumbers between -0.25 and -0.01 cycles per degree. Averaging over all modes, a minimum in wave activity in February and a maximum in August are found. However, the seasonal cycles of the individual modes differ (Fig. 8). The seasonal cycles of low-mode Yanai waves (mode 1-3) show a clear maximum in August and a minimum in March/April. In contrast, the maximum of mode-7 Yanai waves is found in May. Note that the amplitude of the seasonal cycle of mode-7 Yanai waves is much smaller compared to the seasonal cycles of mode 1-3. The seasonal maximum of mode-9 Yanai wave activity is found in September and a secondary maximum is present in May. The analysis of the seasonal cycle of Yanai wave activity further supports the hypothesis of different sources for Yanai waves of the different vertical structure. Especially, lower-mode Yanai waves (mode 1-3) show a different behavior than higher-mode Yanai waves (mode 7-9).

Despite the seasonal preference described above, the Hovmoeller plots of the velocity fields of mode-2 and mode-9 Yanai waves reveal interannual variability of the wave activity (Fig. 7). The strong year-to-year variability becomes also evident when comparing the three snapshots of meridional velocity (Fig. 1b-d).
6. Sources and pathways of intraseasonal variability

The analyses of spatial and temporal characteristics of Yanai waves show distinct differences with respect to their baroclinic modes. This suggests that there are likely different sources of intraseasonal Yanai waves. Here, we analyze three possible source regions which we identify by considering high energy levels in the equatorial section of mean meridional intraseasonal kinetic energy (Fig. 2a):

1. The first source region with high meridional intraseasonal kinetic energy levels is the Near Surface Intraseasonal Variability (NSIV) region that is defined as the upper 50 m along the equator across the basin.
2. The second region is located between the North Brazil Current (NBC) and the Equatorial Undercurrent (EUC). The NBC/EUC shear region spans from 44°W to 36°W between 60 and 260m depth (Fig. 9).
3. The third region is also located at the western boundary but at mid-depth. This region is in the vicinity of the DWBC from 44°W to 38°W and between 1100 and 2500 m depth (Fig. 9).

![Mean meridional intraseasonal kinetic energy along the equator](image)

**Fig. 9:** Mean meridional intraseasonal kinetic energy along the equator in [m² s⁻²] (color shading). The gray boxes display the extent of the NBC/EUC (upper box) and DWBC (lower box) source region. Lines represent theoretically constructed Yanai beams. White solid lines show waves with a period of 30 days excited at 20m depth at 44°W and 8°W. The grey dashed-dotted line represents a Yanai beam with a period of 16 days excited at the surface at 10°W. Red solid lines are Yanai waves with periods of 30 and 40 days excited at 42°W at

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260 and 60m, respectively. Red dashed-dotted lines show waves with periods of 80 and 40
days excited at 40°W at 1100 and 2500m depth, respectively.

To analyze the vertical structure of the waves that are excited in the different source
regions, we calculate the wavenumber-frequency spectra within the regions (Fig. 10). We
compare these spectra with the theoretical dispersion relation of Yanai waves. Note that
the wavenumber-frequency structure in the source region does not necessarily agree with the
structure of free waves eventually found away from the source region. Additionally, the
vertical structure of the forcing has to be considered as a vertically narrower forcing projects
on a broader spectrum of vertical normal modes than a vertically wider forcing that favors
low baroclinic modes.

The frequency-wavenumber spectrum of the NSIV exhibits the highest variance on
intraseasonal timescales, with negative wavenumbers between -0.2 and 0 cycles per degree
(Fig. 10a). Comparing the spectra with the theoretical dispersion relations of Yanai waves
suggests that enhanced variance levels of NSIV projects rather on low-mode Yanai waves.
This is further supported by the similarity of the seasonal cycle of intraseasonal kinetic
energy of the NSIV (not shown) and the corresponding seasonal cycle of Yanai waves of
modes 1-3 (Fig. 8). Note that similar but weaker seasonal cycles are also visible in higher-
mode Yanai waves (e.g., mode 6). This suggests that waves excited in NSIV region consists
of a superposition of different vertical modes. However, the main signal corresponds to a
low-mode structure. Additionally, we analyze the NSIV as a function of longitude by
calculating spectral analysis in time at each grid point along the equator (not shown). The
NSIV consists mainly of two signals. Between the western boundary and 10°W a strong
signal with a period of about 30 days is found. East of 10°W a weaker high-frequency signal
is visible (~16 days). The longitudinal range of the 30-day signal fit very well to the
longitudinal range of mode-2 Yanai wave activity (Fig. 6).

We further investigate which part of the subsurface intraseasonal variability can be
explained by downward propagating Yanai waves from the NSIV region by constructing
theoretical Yanai beams. We construct theoretical Yanai beams with a period of 30 days
starting at 20 m depth at the western boundary (44°W) and at 10°W (Fig. 9). It becomes
evident that Yanai waves excited by NSIV can reach a vast subsurface region of the
equatorial Atlantic. However, high intraseasonal energy levels at larger depths in the west
cannot be explained by Yanai waves propagating downward from the surface. Additionally,
we construct a Yanai beam with a period of 16 days starting at the surface at 10°W. Note
that, in general, the higher the wave frequency, the steeper the slope of the Yanai beam (see Eq. 6). This beam shows that high-frequency Yanai waves excited east of 10°W are able to explain elevated energy levels near the eastern boundary.

In summary, the analysis above shows that NSIV is a viable source of intraseasonal Yanai waves. Analysis in the frequency-wavenumber space reveals that waves excited by the NSIV primarily are of low-baroclinic-mode structure. This hypothesis is supported by the similarity of the seasonal cycle of NSIV and low-mode Yanai waves. Yanai waves with a period of 30 days excited at the surface can be made responsible for a large fraction of the intraseasonal variability below the surface. However, a large part of the wave energy found in the west located at larger depths cannot be explained by it. Elevated intraseasonal energy levels near the eastern boundary can be explained by Yanai beams with shorter periods.

The NBC/EUC region is a region of strong meridional and zonal shear. The NBC is flowing northwestward along the boundary of Brazil. After crossing the equator, a part of the NBC retroreflects into the EUC. Thus, the EUC is coming from the north and turns eastward at the equator at about 40°W (Schott et al., 2002).

Frequency-wavenumber analyses of the NBC/EUC shear region show variability with negative wavenumbers on intraseasonal to interannual timescales (Fig. 10b). The highest variance is found at periods between 30 and 60 days with wavenumbers of about -0.125 cycles per degree. Note that because of the smaller horizontal extent the wavenumber resolution is lower than in the frequency-wavenumber spectrum of NSIV. Because of the low frequency resolution of the spectrum, an attribution of the derived spectrum to theoretical dispersion relations is ambitious. Nevertheless, the longitudinal distribution of kinetic energy of Yanai waves suggests that the waves excited here are of high-mode structure (Fig. 6). Especially, mode-7 and mode-9 Yanai waves have high energy levels close to the western boundary.

The seasonal cycle of intraseasonal variability in the NBC/EUC region shows elevated activity between March and August with a maximum in July (not shown). The amplitude of the seasonal cycle is much smaller than the seasonal cycle of NSIV. The seasonal cycle of mode-7 Yanai waves has similar characteristics (Fig. 8). It has a maximum in May and elevated activity is found between February and July. The amplitude of the seasonal cycle of mode-7 Yanai waves is also considerably smaller than that of mode 1-3 Yanai waves. The seasonal cycle of mode-9 Yanai has two local maxima. One in May and one in August which

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suggest that mode-9 Yanai waves are probably excited both in the NBC/EUC and the NSIV region. In general, the analysis of the seasonal cycle of intraseasonal variability in the NBC/EUC region shows that waves excited here are of higher baroclinic mode structure.

We construct theoretical Yanai beams from the NBC/EUC region to discuss the intraseasonal variability that can be explained by downward propagation of Yanai waves. Mode-7 and mode-9 Yanai waves with zonal wavenumbers of about -0.125 cycles per degree have periods between 30 and 40 days (Fig 10b). Thus, we analyze a Yanai beam with a period of 30 days which is excited at the lower boundary of the NBC/EUC shear region (260m) and a Yanai beam with a period of 40 days excited at the upper boundary (60m). The beams are excited at 42°W which is the longitude with the maximum intraseasonal energy in the NBC/EUC region (Fig. 9). The Yanai beams form an ‘envelope’ for the downward propagating signal of the mean meridional intraseasonal kinetic energy. The analysis shows that Yanai waves with periods between 30 and 40 days propagating east- and downward encounter topographic features at ~20°W (see also Fig. 2a for the features in the vicinity of the equator). Note that the energy levels of mode-7 and mode-9 Yanai waves greatly decrease east of 20°W (Fig. 6). This supports the hypothesis that intraseasonal variability in the NBC/EUC region excites Yanai waves of high-baroclinic-mode structure.

In summary, a downward propagating signal from the NBC/EUC source region at about 20°W is visible in the mean intraseasonal meridional kinetic energy and can be best explained by high-baroclinic-mode intraseasonal Yanai waves. Those waves have zonal wavenumbers of about -0.125 cycles per degree (wavelengths of ~900 km) and periods between 30 and 40 days. The dominant signals here are high-baroclinic-mode Yanai waves.

The region at the western boundary at mid-depth, close to the DWBC, shows enhanced levels of intraseasonal energy as well (Fig. 9). The energy levels are lower than those for the NSIV or in the NBC/EUC regions. Looking at the mean velocity section shows that the highest intraseasonal variability is found just east of the DWBC which flows southward along the continental slope.

The frequency-wavenumber spectrum calculated for this region reveals variability with negative wavenumbers on intraseasonal to longer periods (Fig. 10c). Elevated levels centered around -0.25 cycles per degree are evident. The variability in the DWBC region exhibits larger negative zonal wavenumbers than in the other source regions. Comparing to theoretical dispersion relations of Yanai waves, we suggest that it might project on Yanai waves with

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longer periods (between 40 and 80 days) and shorter wavelength (between 250 and 500 km). Little can be said about the vertical structure on which this signal projects, as the theoretical dispersion relations are similar in this wavenumber range. However, the snapshot of modeled meridional velocity reveals velocity structures of large vertical extent which imply barotropic and low-baroclinic-mode activity (Fig. 1b). This is in agreement with the elevated barotropic and mode-1 Yanai wave energy near the western boundary (Fig. 6). Note that the dispersion relation of asymmetric \( (l = 2,4,6, \ldots) \), low-baroclinic-mode Rossby waves would also fit to the elevated meridional kinetic energy levels in the frequency-wavenumber spectra in the DWBC. Short Rossby waves would also allow eastward energy propagation away from the western boundary. However, equatorial beams of such Rossby waves \( (l = 2,4,6, \ldots) \) are much steeper than the ones of Yanai waves \( (l = 0) \) of the same periodicity (Eq. 6). As intraseasonal energy levels are low in the corresponding regions intraseasonal Rossby waves seem to not play an important role here.

Beams of Yanai waves with periods of 80 (40) days originating from the upper (lower) boundary of the DWBC region demonstrate that Yanai waves excited here can explain elevated intraseasonal energy levels in the deep western part of the basin (Fig. 9). Waves excited in the upper part of the DWBC region are able to pass the Mid-Atlantic Ridge and reach the eastern basin at depth. Note that Yanai waves excited in the DWBC region can also propagate their energy upwards. However, upward propagating Yanai waves are masked by more energetic Yanai waves from the upper ocean and are hardly detectable.

In summary, the DWBC region seems to be a source region for intraseasonal variability along the equator. However, the energy levels associated with the DWBC are lower than in the other source regions. Yanai wave energy that are shed from the DWBC region propagates eastward with flatter downward propagation as the periods of the waves excited here are longer (see Eq. 6).
Fig. 10: The frequency-wavenumber spectra of meridional velocity in the proposed sources regions (see text and also Fig. 9 for a definition of these regions). Spectra are shown for (a) Near Surface Intraseasonal Variability (NSIV), and for the intraseasonal variability of (b) the North Brazil Current/Equatorial Undercurrent (NBC/EUC), and (c) the Deep Western Boundary Current (DWBC). Spectra are calculated for each depth level of the source region at the equator in its longitude range, and averaged over depth afterwards. For the definition of the depth and longitude range of each source region see text. Black lines represent the theoretical dispersion relations of Yanai waves of mode 2 (solid), mode 7 (dashed-dotted) and mode 9 (dotted).

7. Summary & Discussion

In this study, intraseasonal variability of the meridional velocity in the equatorial Atlantic Ocean is analyzed using the output of a high-resolution ocean general circulation model. The model is able to reproduce the structure and strength of observed meridional intraseasonal...
variability at the equator and is therefore a useful tool to investigate possible sources and pathways of the intraseasonal variability.

The main findings of the study are:

i. The intraseasonal meridional velocity variability in the equatorial Atlantic can mostly be explained by Yanai waves of the first 12 vertical normal modes with zonal wavenumbers between -0.25 and -0.01 cycles per degree (wavelength larger than 450 km). Yanai waves of mode 2 exhibit the strongest signal.

ii. The baroclinic structure of Yanai waves and their spatial distribution varies along the equator. Low-baroclinic-mode Yanai waves are mostly active between the western boundary and 10°W with a maximum at ~26°W. High-baroclinic-mode Yanai waves show highest activity west of 20°W.

iii. The Yanai wave activity along the equator exhibits a large interannual variability. However, a seasonal preference is still visible. Yanai waves of mode 1-3 have a distinct seasonal cycle with a maximum in August. The seasonal cycle of mode-7 Yanai waves has a smaller seasonal amplitude and a maximum in May.

iv. Three regions are identified and analyzed as likely generation sites of intraseasonal Yanai waves:

   a. NSIV is primarily associated with the excitation of low-baroclinic-mode Yanai waves (mode 1-3). Waves excited near the surface dominate the intraseasonal variability of vast subsurface regions along the equator.

   b. High-baroclinic-mode Yanai waves are excited in the shear region between the NBC/EUC. Corresponding wave energy propagates down- and eastward to reach the bottom at about 20°W. The downward propagation is dominated by high-baroclinic-mode Yanai waves.

   c. A region close to the DWBC exhibits elevated intraseasonal energy levels as well. Yanai waves excited here have shorter wavelengths and longer periods than in the other regions. The waves excited near the DWBC are less energetic than the waves from the other two source regions.

Our results show that the NSIV region is an important source region of intraseasonal variability at depth. The NSIV consists mainly of two signals: Between the western boundary and 10°W a strong signal centered around 30-day period is found. East of 10°W a weaker signal with periods of ~16 days is present. The periodicity and seasonality of the 30-day
signal fits well to the characteristics of TIWs (Weisberg & Weingartner, 1988; Athie & Marin, 2008; Tuchen et al., 2018; Specht et al., 2021). This suggests that TIWs excite low-baroclinic-mode Yanai waves, their energy is propagating down- and eastward from the surface. Note that Tuchen et al. (2018) identified in the mooring data taken at 0°N, 23°W Yanai waves dominantly of mode 2-5 with periods of 30-40 days. These finding agree well with the low-baroclinic-mode signal discussed in this study. The longitude range and periodicity of the high-frequency signal east of 10°W agree with observed Yanai waves forced directly by meridional wind fluctuations (Athie and Marin 2008; de Coëtlogon et al. 2010). Their wave energy propagates down- and eastward. Upon reaching the eastern boundary they may propagate poleward as coastally trapped waves (Guiavarch et al., 2008).

The model results suggest that high-baroclinic-mode Yanai waves with periods of 30-40 days are excited in a region between the NBC and the EUC. These waves are most likely generated by flow instabilities associated with the meridional and zonal shear between the two currents. However, further analyses of the exact mechanism have to be conducted. This is of interest as the waves are energetic and play a key role in transporting intraseasonal wave energy into the deep ocean. Note that the NBC/EUC shear region has so far been overlooked as a source region for deep intraseasonal variability although indicated in the maps of barotropic instability shown for the near surface layer (upper 50m) by von Schuckmann et al. (2008). Tuchen et al. (2018) found a signal of intraseasonal energy at depth that cannot be explained by downward propagating Yanai beams originating near the surface. Waves excited in the lower part of the NBC/EUC region possibly represent the source for the deep energy found in the mooring data at 23°W.

Yanai waves excited near the DWBC have longer periods and shorter wavelengths and flatter downward beams than Yanai waves excited near the surface. The waves are most likely excited by instabilities of the DWBC although a closer analysis of generation mechanism has to be conducted as well. Analyses of snapshots of meridional velocity as well as analyses of kinetic energy of Yanai waves suggest that the Yanai waves excited in the DWBC region are of low-baroclinic or barotropic mode. This conclusion can also be drawn from the observed equatorial section of meridional velocity (Fig. 1a). Observed moored velocities support the discussed difference in periodicity between intraseasonal variability near the surface and at depth of the DWBC (Fig. 2b).
The importance of intraseasonal variability for the dynamics of the equatorial Atlantic has been underlined by several studies (Hua et al., 2008; Ascani et al., 2015; Greatbatch et al., 2018; Bastin et al., 2020). Especially, the role of DEIV for EDJs and the EICs also called EEJs was pointed out (Cravatte et al. 2012, 2017; Ménesguen et al. 2019; Delpech et al. 2020b), which in turn influences the distribution of tracer (Brandt et al. 2012, Delpech et al. 2020a) and possibly regional climate variability (Brandt et al., 2011). However, modeling the intermediate and deep circulation still remains a challenge. In this context the need for the knowledge of structure and strength of DEIV has been pointed out (Ménesguen et al. 2019). We show that intraseasonal waves of different strength, frequency and vertical structure are active in different parts of the equatorial Atlantic. The consequences of this for the forcing and maintenance of the DEC has to be closer examined in the future as previous model studies focused typically on one single intraseasonal wave type. For example, d’Orgeville et al. (2007), Hua et al. (2008) and Ménesguen et al. (2009) used short Yanai waves destabilizing in the western equatorial basin to generate EDJ-like structures. These waves resemble the waves excited in the DWBC region discussed in the present study. In contrast, in the simulations of Ascani et al. (2015) and Bastin et al. (2020) DEIV is associated with downward propagating Yanai waves excited from TIWs. The present study shows that such waves are important for elevated intraseasonal energy levels in the deep ocean. Furthermore, we also show that the energy of high-baroclinic-mode Yanai waves propagate down- and eastward from the NBC/EUC region. However, open questions still remain such as: What are the exciting mechanisms of the intraseasonal wave in the different source regions? What are deterministic mechanisms for the year-to-year variability of DEIV? Do we expect long-term changes in the appearance of DEIV due to changes of the upper ocean stratification or of the wind-driven and thermohaline circulation? And what could be the consequences for the DEC and the distribution of tracers? These questions have to be addressed in the future to deepen the understanding of the DEIV as an energy source for the DEC.

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Data Availability Statement.

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The observational data used in this study are publicly available. The mooring data is referenced in Table 1 of the Appendix. The near surface PIRATA current meter data at 0°N, 23°W can be accessed via https://www.pmel.noaa.gov/tao/drupal/disdel/.

APPENDIX

Appendix A

Moored velocity data from the equator

<table>
<thead>
<tr>
<th>Mooring Period</th>
<th>Latitude/Longitude</th>
<th>Data availability</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oct. 1992-May 1994</td>
<td>0°N/36°W</td>
<td></td>
</tr>
<tr>
<td>Aug. 2004- June 2006</td>
<td>0°N/35°W</td>
<td></td>
</tr>
<tr>
<td>Dec. 2001- July 2021</td>
<td>0°N/23°W</td>
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<td></td>
</tr>
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<td>Jun. 2006 – Mar. 2019</td>
<td>0°N/10°W</td>
<td><a href="https://doi.org/10.17882/51557">https://doi.org/10.17882/51557</a></td>
</tr>
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</tr>
<tr>
<td>Jun. 2009- Jun. 2011</td>
<td>0°N/0°E</td>
<td></td>
</tr>
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</table>

Table A1: Overview of moored velocity data used in the study.
Figure A.1: Colored area gives an overview of available velocity data from equatorial moorings at different locations given in the title of the respective subplots.

Appendix B

Comparison of modeled and observed intraseasonal meridional kinetic energy
Figure B.1: Profiles of mean intraseasonal meridional kinetic energy at different mooring locations given in the title of the respective plots. Observations (model) are presented in red (black). Shaded area/horizontal lines give the 10th to 90th percentile of the distribution of intraseasonal meridional kinetic energy. Note the different extent of the horizontal axis.

REFERENCES


Athie, G., & Marin, F., 2008: Cross-equatorial structure and temporal modulation of


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Accepted for publication in *Journal of Physical Oceanography*. DOI 10.1175/JPO-D-21-0315.1.


d’Orgeville, M., Hua, B. L., & Sasaki, H., 2007: Equatorial deep jets triggered by a large


Jochum, M., Malanotte-Rizzoli, P., & Busalacchi, A., 2004: Tropical instability waves in the


WOCE Data Products Committee (2002). NODC Standard Product: World Ocean Circulation Experiment (WOCE) Global Data Resource (GDR), versions 1-3, on CD-ROM and
