The characteristics and variability of intraseasonal coastal Kelvin waves in the Bay of Bengal under hindcast conditions and the RCP8.5 scenario

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ABSTRACT: The characteristics and variability of intraseasonal internal coastal Kelvin waves (CKWs) along the Bay of Bengal (BoB) waveguide are investigated in the context of global warming by employing a regional ocean model. The analyzed period covers 120 years from 1980 to 2099, which includes the historical scenario and the RCP8.5 scenario. CKW information is successfully extracted from the temperature anomalies along the pycnocline by applying a newly developed methodology. The analysis reveals that intraseasonal CKWs in the BoB are highly in accordance with the intraseasonal zonal wind stress in the western equatorial Indian Ocean; the downwelling CKW lags the equatorial intraseasonal westerly winds, and the upwelling CKW lags the equatorial intraseasonal easterly winds. The CKWs significantly affect subsurface characteristics at the eastern BoB boundary; and the weakening of CKWs near the Irrawaddy Delta tip is a general feature occurring in the subsurface. With respect to the long-term scale, the occurrence of significant CKWs is predicted to be more frequent in the future under the high emissions pathway. Remarkably, the monthly climatology of CKWs varies over time; unlike the first two 30-year analyzed periods, significant CKWs are predicted to mainly occur around August during the last two 30-year periods due to the corresponding variabilities in the equatorial wind field, suggesting that the BoB characteristics may greatly deviate from the current climatological state.
1. Introduction

The region encompassing the Bay of Bengal and the Andaman Sea located in the northern Indian Ocean (this region is referred to as the BoB for simplicity in this study) is uniquely bordered to the east by the Sumatra-Java Islands and to the north by a wide continental shelf; as a result of these geographic features, the BoB is susceptible to equatorial signals carried by Kelvin waves and associated Rossby waves. Coastal Kelvin waves (CKWs) and the associated westward-moving Rossby waves appear as eigenmodes of the stratified ocean (Allen 1975; Wang 1975; Wang and Mooers 1976). Theoretically, higher latitudes and frequencies favor trapped motions while lower latitudes and frequencies favor untrapped motions (Grimshaw and Allen 1988; Clarke and Shi 1991). In the BoB, the contributions of these planetary waves to the variability of the thermocline and barrier layer (McCreary et al. 1993; Rao and Sivakumar 2000; Girishkumar et al. 2011, 2013), the variability of sea surface salinity (Sharma et al. 2010; Trott et al. 2019), the variability of currents (Potemra et al. 1991; Yu et al. 1991; McCreary et al. 1996; Durand et al. 2009; Chen et al. 2017; Chatterjee et al. 2017), and the generation of mesoscale eddies (Sreenivas et al. 2012; Cheng et al. 2013; Chen et al. 2018; Cheng et al. 2018; Dey et al. 2018) have been intensively investigated. For instance, the seasonal reversal of the East Indian Coastal Current (EICC) is significantly influenced by remotely forced planetary waves, resulting in possible lags between the EICC and the local monsoon (Eigenheer and Quadfasel 2000). Furthermore, intraseasonal changes observed in the southern BoB are also largely attributed to eddy-like Rossby wave signals (Schott et al. 1994; Sengupta et al. 2001; Chen et al. 2017). A lag Pearson correlation between BoB subsurface salinity anomalies and the Indian Ocean Dipole mode was found in multiple data sets, in which advection processes induced by planetary waves dominate (Zhang et al. 2021). Satellite observations and hydrographic data revealed that eddies are generated near the tip of the Irrawaddy Delta when equatorial wind-driven Kelvin waves reach this area, demonstrating the dominant role that CKWs play in eddy generation within the eastern BoB (Cheng et al. 2018). While, a recent high-frequency coastal radar based study revealed that the annual cycle of surface shelf current off Indian east coast is absent, which is attributed to the finite bottom friction that makes shelf a high pass filter (Paul et al. 2021).

Previous studies have demonstrated the existence of an annually repeating cycle of two pairs of alternating upwelling and downwelling Kelvin waves, which travel along the waveguide of
the equatorial Indian Ocean (EIO) and the BoB coast in the presence of a seasonally reversing monsoon (Rao and Sivakumar 2000; Rao et al. 2010; Nienhaus et al. 2012; Sreenivas et al. 2012). The first (second) upwelling Kelvin wave originates in the western (eastern) EIO, whereas two downwelling Kelvin waves originate in the central EIO (Sreenivas et al. 2012). Moreover, the second downwelling Kelvin wave can propagate into the southeastern Arabian Sea, while the others are geographically limited to the BoB coast (Rao et al. 2010). In situ observations detected significant semiannual and intraseasonal oscillations in the upper eastern EIO in terms of the temperature and salinity structures (Hase et al. 2008) and current field therein (Masumoto et al. 2005; Iskandar et al. 2009). In addition to the signal inherited from equatorial Kelvin waves (EKWs), the local alongshore wind is also able to trigger and modulate CKWs along the BoB coast (McCreary et al. 1993; Rao et al. 2010). A modeling study demonstrated that equatorial wind and the associated CKWs dominate the sea level anomalies near the eastern and northern BoB boundaries, while local winds have a comparable influence near the western boundary (Han and Webster 2002). Furthermore, the oceanic intraseasonal oscillation in the tropical Indian Ocean is highly affected by the Madden-Julian Oscillation (MJO), which is the dominant component of the intraseasonal variability in the tropical atmosphere (Zhang 2005; Vialard et al. 2009; Roman-Stork et al. 2020). In addition, there are atmospheric processes on the intraseasonal time scale that propagate northward in the tropical Indian Ocean, whether it is called the boreal summer intraseasonal oscillation (BSISO) or the monsoon intraseasonal oscillation (MISO), which also contribute to the oceanic intraseasonal variability in the BoB (Wang and Xie 1997; Sharmila et al. 2013; Krishnamurti et al. 2017; Girishkumar et al. 2017; Trott and Subrahmanyam 2019; Chen and Wang 2021). Interestingly, satellite-observed sea level heights have been extensively used to investigate the intraseasonal variability within the Indian Ocean (Somayajulu et al. 2003; Iskandar et al. 2005; Vialard et al. 2009; Durand et al. 2009; Cheng et al. 2013, 2017; Chen et al. 2018; Dey et al. 2018; Trott and Subrahmanyam 2019; Pujiana and McPhaden 2020; Roman-Stork et al. 2020).

However, due to the lack of in situ observational data, the characteristics and variability of the intraseasonal internal CKWs along the BoB coast are still not fully understood. As mentioned above, these CKWs greatly influence the thermocline, barrier layer, eddy fields, and currents throughout the BoB. Notably, the CKWs and associated Rossby waves provide an oceanic link between the
equatorial region and the BoB, transporting equatorial information northward (Somayajulu et al. 2003). Although a number of studies have focused on the sea surface features of CKWs, the CKW propagation processes along the isopycnal interface of the stratified ocean have rarely been discussed. To the best of our knowledge, a related question, that is, whether CKW behavior will change as a result of global warming, which is also within the scope of this paper, has not yet been addressed. To conclude, two scientific questions are addressed herein. First, what are the characteristics of the intraseasonal CKWs in the BoB with respect to their propagation along the isopycnal interface? Second, what is the projected change in the CKW behavior under the high emissions pathway?

To answer the above questions, the characteristics and variability of the intraseasonal CKWs in the BoB are investigated through a numerical downscaling approach. Employing a newly developed methodology for extracting intraseasonal EKW signals (Rydbeck et al. 2019), the intraseasonal CKWs in the BoB can be quantitatively analyzed. Considering that current greenhouse gas emissions are already high and are expected to continue to rise for various reasons, representative concentration pathway 8.5 (RCP8.5) is adopted to represent the high emissions scenario (Moss et al. 2010; van Vuuren et al. 2011). The remainder of this paper is organized as follows. The model configuration and the methodology for extracting CKWs are introduced in section 2. The results, specifically, the stratification variability, CKW characteristics, and CKW variability, are presented and analyzed in section 3. Finally, the summary and discussion are given in section 4.

2. Model and methodology

a. Model description and validation

A dynamic downscaling simulation of the BoB (Figure 1) is performed using the Hamburg Shelf Ocean Model (HAMSOM) (Backhaus 1985; Pohlmann 1996, 2006; Zhang et al. 2021). The horizontal model resolution is 1/12th degree × 1/12th degree, and a total of 58 vertical layers are set, of which 10 layers cover the upper 50 m and 16 layers cover the upper 100 m. The model is driven by Max Planck Institute for Meteorology Earth System Model (MPI-ESM-MR) data (Giorgetta et al. 2013; Jungclaus et al. 2013; Stevens et al. 2013) under the historical scenario (from 1951 to 2005) and the RCP8.5 scenario (from 2006 to 2099). Atmospheric forcings, such as air temperature, cloud cover, precipitation, specific humidity, air pressure, wind stress in the
zonal and meridional directions, and wind speed, are composed of six-hourly data prescribed at the air-sea interface. Oceanic forcings, such as sea level height, salinity, and temperature, are prescribed monthly at the open lateral southern and western boundaries in our case. In addition, six-hourly river discharge data are prescribed by adjusting the sea level heights at river mouth locations. The HAMSOM simulation of the RCP8.5 scenario is a continuation of the historical simulation previously described in our preceding article (Zhang et al. 2021), where all relevant model settings and validations of climatological means and seasonal variabilities are described in full detail. The model results from 1980 to 2099 are analyzed.

![Map of the Indian Ocean and Western Equatorial Indian Ocean (WEIO)](image)

**Fig. 1.** (a) Bathymetry (m) and topography of the Indian Ocean, where the blue box shows the modeling domain (BoB), and the red box shows the Western Equatorial Indian Ocean (WEIO). (b) Waveguide along the 100 m depth contour of the BoB, where gray contours show the bathymetry (m), and blue pentagons are labeled to give their distance from the starting point (D0) along the waveguide with the distance (in $10^3 km$) is labeled in square brackets. The subarea Area-E is marked by the blue box.

In order to further validate the sensitivity of intraseasonal variability for our model simulation, the daily output of the Global Ocean Forecasting System version 3.1 (GOFS 3.1) (Cummings 2005; Metzger et al. 2014) is used. The standard deviation can be used to measure the magnitude of variable changes, therefore, standard deviations of unfiltered temperature and 30-180-day bandpassed temperature from 1995 to 2004 are compared at 100 m depth between GOFS 3.1 and HAMSOM results (Figure 2). It can be seen that in the subsurface, the standard deviations of unfiltered (Figure 2c) and filtered (Figure 2d) temperature calculated by HAMSOM not only agree well with GOFS 3.1 (Figure 2a, 2b) in magnitude, but also in spatial pattern. The above results suggest that, in the western BoB and the area close to the western Sumatra coast, the subsurface
temperature changes are significant on the full-timescale and the intraseasonal timescale. In the region of about 5°N, the subsurface intraseasonal temperature changes simulated by HAMSOM (approximately 1°C) are relatively weaker than that of GOFS 3.1 (approximately 1.5°C), which may be due to the open-boundary scheme of the regional simulation and the complex equatorial flow system here. However, this region is not the focus of this study, and the temperature changes agree well in the waveguide region that we are interested in. Therefore, our model simulation can reproduce the intraseasonal variability in our research area.

![Images of temperature distribution]

Fig. 2. Distributions of standard deviations of (a, c) unfiltered temperature and (b, d) 30-180-day bandpassed temperature at 100 m depth. The first (second) row shows the results of GOFS 3.1 (HAMSOM).

b. Methodology

Rydbeck et al. (2019) developed a methodology to objectively determine the amplitude and phase information of EKWs on the basis of sea surface height anomalies. This method is analogous to approaches utilized for atmospheric intraseasonal variabilities. Therefore, this method should also be appropriate to analyze internal CKWs as well as other intraseasonal wave-type variations.

Figure 1b shows the waveguide at 100 m (D100) depth isobath along the BoB coast. Previous studies have verified that CKW propagates along the eastern boundary of the Andaman Sea (Clarke and Liu 1994; Rao et al. 2010; Nienhaus et al. 2012; Cheng et al. 2013, 2018). This
waveguide start at the western Sumatra coast and run counterclockwise to the east coast of Sri Lanka. In the Northern Hemisphere, CKWs always propagate counterclockwise, so we define the counterclockwise direction as the positive direction of the waveguide. Thus, the CKWs must be positive wavenumber signals along the waveguide.

To determine the information of internal CKWs, the daily anomalies of the one-year running mean of the temperature along the waveguide are analyzed. Because spatial filtering is necessary, temperature anomalies along the waveguide are linearly interpolated to achieve an equidistant distribution in space; then, these temperature anomalies can be filtered using the methodology described by Rydbeck et al. (2019). The first three harmonics of the seasonal cycle are first removed by implementing linear regression. Next, the clockwise-moving waves are eliminated from the anomalies by applying a two-dimensional fast Fourier transform. Finally, the anomalies are bandpass filtered using a 30-180-day Lanczos filter to isolate the time scale of concern.

These filtered anomalies are used to perform an empirical orthogonal function (EOF) analysis. The Kelvin wave index (KWI), which represents the averaged strength of the CKW-associated oscillations over the analyzed area, can be calculated from the first two leading principal components (PCs) using the following formula:

\[ KWI = \sqrt{(PC1^2 + PC2^2)}. \]

The CKW-associated temperature anomalies can be reconstructed from the EOFs as follows:

\[ T_{ckw}' = EOF1 \times PC1 + EOF2 \times PC2. \]

3. Results

a. Stratification variability

Figure 3 offers an overview of the simulated background stratification in the model domain during four climate periods. The depth of the maximum buoyancy frequency indicates the depth of the pycnocline. Under the RCP8.5 scenario, the stratification strength enhances and the pycnocline depth becomes shallower in most areas of the research domain. Although the background stratification is affected by the high greenhouse gas emissions, the overall spatial pattern is primarily
the same throughout the entire simulation. At the northern and western boundaries of the BoB, the stratification is relatively shallow and strong; in contrast, in the eastern, central, and southern regions, the stratification is relatively deep and weak. In the area close to the equator, the pycnocline depth is approximately 100 m, while the area near the coast, especially the northern boundary, is affected by abundant freshwater supplied by rivers, and the pycnocline is shallow with a depth of 20 m. Overall, in the counterclockwise direction of the waveguides, the pycnocline depth changes northward from deep to shallow and southward from shallow to deep.

Fig. 3. Distributions of the maximum buoyancy frequency (s$^{-1}$) and its depth (m) for the climate period (a) 1, (b) 2, (c) 3, and (d) 4 (clim. 1, 2, 3, 4 refer to 1980-2009, 2010-2039, 2040-2069, 2070-2099, respectively). White contours mark the depth of the maximum buoyancy frequency, and depths greater than 100 m are specifically hatched with gray dots.

The subarea (purple box in Figure 1b) representing the eastern boundary are selected to investigate the long-term stratification trend and vertical structure (Figure 4). The results of the 30-year running mean clearly show a temperature increase within the upper ocean under the RCP8.5 scenario. Although the depth of the maximum buoyancy frequency shows an increasing trend, even reaching approximately 30 m at the end of the 21st century, a substantial pycnocline exists at a depth of approximately 100 m.
b. CKW characteristics

Figure 5a shows an example of temperature anomalies at D100 in the year 1993, where the year 1993 was randomly selected. The unfiltered temperature anomalies show a clear semiannual signal that reflects two pairs of upwelling and downwelling Kelvin waves from the equator (Rao and Sivakumar 2000; Hase et al. 2008; Rao et al. 2010; Sreenivas et al. 2012), while intraseasonal CKWs are unrecognizable. The filtered temperature anomalies are shown in Figure 5b, which clearly shows intraseasonal signals propagating counterclockwise along the waveguide. An obvious weakening of these intraseasonal signals is detected near D2, which probably results from the transfer of energy from CKWs to eddies (Sreenivas et al. 2012; Cheng et al. 2013, 2018; Chen et al. 2018; Dey et al. 2018). Equatorial wind-driven Kelvin waves can even propagate into the southeastern Arabian Sea (Rao et al. 2010). The filtered subsurface temperature anomaly also indicates that some extreme values starting from D0 can still propagate along the waveguide after D3 reaching the BoB western boundary. Meanwhile, some stripes of extreme values in the second half of D100 show a greater slope, which means that the propagation speed of these signals is slower. Taking into account the difference in stratification distribution shown in Figure 3, and that the depth of the pycnocline along D100 is shallower from D3 to D6 (approximately 20 m) than from D0 to D3 (approximately 100 m), we think that these signals are modulated by local winds. To focus on the CKW signals originating from the EIO and eliminate as much as possible the influence of local winds and associated shelf waves, the part from D0 to D3, which in the following is called D100-seg, is analyzed hereafter.
Figure 6 shows the EOF results of the filtered temperature anomalies. The variance proportions explained by the first ten modes are shown in Figure 6a for D100 and D100-seg. The first two EOF modes of Rydbeck et al. (2019) explain approximately 70% of the variance. In our case, due to the specific distribution of the background pycnocline depth, the leading two PCs of the segmented waveguides explain a remarkably larger proportion of the variance than compared to the complete waveguide. Therefore, in the following, we analyze only the EOF results of D100-seg. The leading two PCs of D100-seg explain approximately 72% of the variance. Different from the analysis of EKWs (Rydbeck et al. 2019), the variance fractions of the first two modes in our analysis of CKWs are not roughly equal. We think that this is mainly due to the complicated coastal bathymetry, which causes changes in the waveform. The first two coherent EOF modes show a phase shift and are not smooth (Figure 6b). Unlike the smooth EOF modes shown by the EKW signals, the EOF mode associated with the CKWs is often not smooth, which is also shown by a recent analysis of EKWs and CKWs through sea surface height anomalies in the Pacific (Amaya et al. 2022). Here we choose the year 1993 as an example to show variations of the first two PCs (Figure 6c). It is clear that PC1 is lagging behind PC2. The lag Pearson correlation between PC1 and PC2 shows that their phase difference is approximately 15 days with a correlation exceeding 0.6 (Figure 6d). The power spectra for PC1 and PC2 of D100-seg show two peaks near 70 and 90 days (Figure 6e, 6f). Because of the larger proportions of the explained variance and the considerable correlation
between the first two modes, it is reasonable to assume that these two modes can represent the internal CKWs in our study area reasonably well.

![Graphs and figures](image)

Fig. 6. (a) Variance proportions of the first ten EOFs at D100-seg (from D0 to D3) and D100, (b) the first two EOF modes at D100-seg, (c) the first two PCs at D100-seg in the year 1993, (d) lag Pearson correlation between the first two PCs at D100-seg, and (e) power spectrum of PC1 (dashed line shows the 99% significance level); (f) same as in (e) but for PC2.

The time series of the daily KWI and its distribution for D100-seg are shown in Figure 7. By definition, the strength of the KWI indicates the strength of wave-associated anomalies averaged over the analyzed waveguide, which makes it possible to detect significant cases of CKWs. The distribution of the KWI is skewed to the left, and it agrees well with the exponentiated Weibull distribution. To detect significant CKWs, the 90% value of the probability density function for the fitted distribution is defined as the threshold. So, in our case, every peak value of the KWI, which is greater than 15.86 will be labeled as a significant CKW.

The distribution of reconstructed temperature anomalies for D100-seg is depicted in Figure 8a. The reconstructed anomalies are consistent with the Student’s t-distribution since they represent the deviation from the average state. Similar to the selection of the 90% threshold for the KWI, the 5% and 95% values are defined as symmetric thresholds for significant negative and positive anomalies, respectively. Under this definition, temperature changes that represent significant CKWs must exceed 0.87 °C at D100-seg. The large number of events, where this threshold value is exceeded, points out that the temperature changes caused by CKWs are significant in our study area.
Fig. 7. (a) Time series of the KWI at D100-seg; (b) shows the probability density distribution of (a). The solid black line shows the fitted probability density function of the exponentiated Weibull distribution, and the dashed red line indicates the 90% threshold of the KWI. Red parts in (a) indicates a KWI exceeding the 90% threshold value.

The reconstructed temperature anomaly of D100-seg and the corresponding KWI are shown in Figure 8b. The stripes of extreme values shown here represents the temperature fluctuation caused by intraseasonal CKWs. As can be seen, these fluctuation are greatly weakened near D2, where is near the Irrawaddy Delta tip, supporting the finding that eddies generated in the eastern bay are highly correlated to the CKWs (Cheng et al. 2018), when looking at subsurface variations. In 1993, there are three peak values of the KWI, which are greater than the threshold value we defined, and thus three significant CKWs are detected and labeled. Stripes with negative anomalies are recognized as upwelling CKWs, on the contrary, stripes with positive anomalies are recognized as downwelling CKWs. Therefore, in our example of 1993, there are two significant upwelling CKWs and one significant downwelling CKW. The phase speed of each significant CKW can be estimated on the basis of the significant temperature anomalies of the reconstructed fields by calculating a linear regression in the space-time domain. The method for estimating the phase speed may have some errors due to the size and distribution of the analyzed samples. Other processes near the waveguide area, especially those that may lead to changes in stratification, may also change the phase speed of extracted significant CKWs. However, the specific phase speeds and their variations are not in the focus of the current study. Overall, the magnitudes of the estimated phase speeds reasonably agree with the estimates of gravity wave speeds calculated on the basis of the shallow water equation considering the average values for the local upper layer thickness and the local reduced gravity.
Fig. 8. (a) Distribution of the reconstructed temperature anomaly of the first two EOFs for D100-seg from 1980 to 2099, where the solid black line shows the fitted probability density function of the Student’s t-distribution, and the dashed blue (red) line indicates the 5% (95%) threshold of the data. (b) Waveguide-time diagram of the reconstructed temperature anomaly, where solid black line shows the KWI, dashed blue (red) line indicates the -0.87 °C (0.87 °C) contour, and solid magenta lines indicate the phase speed estimated by linear regression of all points within the dashed contour.

A total of 93 significant downwelling CKWs and 127 significant upwelling CKWs are detected from 1980 to 2099 in our simulation. It is well known that oceanic intraseasonal variabilities in the Indian Ocean are modulated by the equatorial wind (Han et al. 2001; Iskandar and McPhaden 2011). By analyzing satellite observations, Pujiana and McPhaden (2020) clearly shows a lagged relationship between the intraseasonal westerly (easterly) wind stress ($\tau_w$) of the western EIO and the EKW signals reflected in the positive (negative) surface height anomalies of the eastern EIO. The detected CKWs in our research area are essentially originating from the EKW signals that are forced by equatorial wind stresses. Therefore, the WEIO-averaged intraseasonal $\tau_w$ and the reconstructed temperature anomaly that attributed to significant CKWs are composited for understanding their general features (Figure 9). To analyze long-term changes, these composites are
divided into four climate periods. \( \tau_x \) is extracted from the MPI-ESM-MR which offers the forcing data for the downscaling simulation in this study. As can be seen, downwelling (upwelling) CKWs with positive (negative) temperature anomalies show a lagged relationship with the equatorial intraseasonal westerly (easterly) \( \tau_x \). For both downwelling and upwelling CKWs, they lag the equatorial intraseasonal \( \tau_x \) by about 25 days. The weakening of CKWs near D2 is a general feature in our research area. By applying a linear regression of significant temperature anomalies, the general phase speed of composited CKWs is estimated to be about 1 \( \text{m/s} \); there is no obvious difference between downwelling and upwelling CKWs. From the perspective of long-term changes, the general pattern and phase speed of significant CKW signals do not show apparent changes (Figure 9b-e, 9g-j); however, the lagged relationship for upwelling CKW signals in the fourth climate period (Figure 9f, dotted line) is not as clear as the other three periods, implying the forcing process may change under the high emissions pathway.

c. CKW variability

The temporal distribution of all detected significant CKWs at D100-seg is shown in Figure 10. The time series of the KWI are also plotted on the same time scale, and they show highly correlated changes with the accumulated occurrence of significant CKWs. The accumulated occurrence for each month (Figure 10b) indicates that the significant CKWs at D100-seg occur throughout the year and show a primary peak in August. Moreover, at D100-seg, significant CKWs do not necessarily occur every year; the statistics show that only one or even no waves may occur in some years, while in other years, up to 5 significant CKWs may appear (Figure 10c). The cumulative results for each 30-year period reveal obvious long-term changes (Figure 10d). The occurrence of significant CKWs at D100-seg will become more frequent; our simulation suggests that the maximum number of occurrence will occur in the second climate period. This finding clearly suggests that under the high emissions pathway, the occurrence of significant CKWs along the BoB coast will be affected.

Figure 10a shows a relatively random distribution, but it also shows that significant CKWs tend not to appear in the first few months in the second half of the 21st century. Further, the monthly distribution of significant CKWs in the four climate periods is statistically analyzed (Figure 11). The monthly climatological KWI is also plotted, and they keep consistent with the accumulated
Fig. 9. (a) Composites of equatorial intraseasonal $\tau_x$ (extracted from the MPI-ESM-MR) averaged over WEIO attributed to significant downwelling CKWs for four climate periods. Lag=0 marks the time when the significant downwelling CKW occurs at D0. (b)-(e) Corresponding composite of the reconstructed temperature anomaly for each period, where dashed red line indicates -0.87 °C contour, and solid magenta line indicates the estimated phase speed. The total number of CKWs used for compositing is labeled in brackets. The second row is as the same as the first row but for significant upwelling CKWs.

occurrence of significant CKWs. Considering the monthly climatology, the KWI at D100-seg shows remarkable changes over time. During the first two climate periods, two comparable peaks are detected in the KWI time series around February and September. In contrast, during the last two climate periods, only one significant peak is found around August. The accumulated occurrence of significant CKWs also indicates this change on the monthly climatological time scale. Hence, our simulation suggests that the monthly variability of the intraseasonal CKWs at the BoB eastern boundary will significant change during the second half of the 21st century under the high emissions pathway.

The equatorial wind field is expected to influence the CKW variability along the BoB coast. The intraseasonal $\tau_x$ in the EIO (Figure 11) shows a clear seasonal cycle reflecting the seasonality of the MJO (Zhang and Dong 2004), and the seasonality of intraseasonal $\tau_x$ varies on the climatological
Fig. 10. (a) Temporal distribution of significant CKWs for D100-seg, where red square represents downwelling CKWs, blue square represents upwelling CKWs, and the total number of downwelling and upwelling CKWs is labeled in brackets, respectively; (b) accumulated occurrence for each month; (c) accumulated occurrence for each year; and (d) accumulated occurrence for each 30-year period. The KWI of the corresponding time scale is marked by a solid orange line in (b), (c), and (d), respectively.

scale, which is probably responsible for shifts in the frequency of CKW occurrence. Different changes of equatorial intraseasonal $\tau_x$ seasonality in different longitude result in changes of its zonal gradient, and in turn probably also affect the generation of EKWs and the subsequent occurrence of significant CKWs. Previous studies suggested that the significant intraseasonal variations in the tropical Indian Ocean are attributed to the complex air-sea coupling processes in this region (Webber et al. 2010, 2012; Sharmila et al. 2013; Krishnamurti et al. 2017). Webber et al. (2010) described a possible dynamic feedback mechanism among the atmospheric progress MJO, the oceanic eastward-moving EKWs, and the westward-moving Rossby waves. The CKW signals analyzed in this study are principally inherited from the EKWs, therefore, changes in the equatorial wind field under the high emissions pathway eventually lead to changes in CKWs along the BoB coast.
Fig. 11. Intraseasonal $\tau_s$ in the EIO (averaged between $2^\circ S-2^\circ N$, shaded) extracted from MPI-ESM-MR, the accumulated occurrence of significant downwelling CKWs (red bar) and significant upwelling CKWs (blue bar), and the monthly KWI (solid orange line) for the climate period (a) 1, (b) 2, (c) 3, and (d) 4, respectively.

4. Summary and discussion

In this study, we investigate the characteristics and variability of intraseasonal CKWs in the BoB by means of a high-resolution regional ocean model. As we presented in our last article, the model we used is able to reveal the distinct contributions of CKWs and associated Rossby waves to the variability of BoB subsurface salinity anomalies (Zhang et al. 2021). The information of BoB CKWs is successfully extracted in this study by using a newly developed methodology (Rydbeck et al. 2019) and is analyzed to investigate the occurrence and strength of these CKWs, as well as their distinct wave characteristics. Because of the specific characteristics of the stratification in the BoB, the CKWs originating from EKWs are discussed at the depth of 100 m at the eastern boundary, which is their entrance area. The entire analyzed period covers 120 years from 1980 to 2099, which includes both the historical scenario and the RCP8.5 scenario. These long-term results allow us to statistically analyze the nature of these CKWs and to predict their possible changes in the context of global warming. Moreover, the influences of equatorial wind on the BoB CKWs are discussed under consideration of climate change.

CKWs are a systematic feature of the stratified ocean. Our analysis reveals distinct internal CKW signals reflected in the subsurface temperature anomaly. Statistically, the top 10% of temperature changes related to CKWs can reach 0.87 °C or more, which demonstrates CKWs can significant
affect subsurface characteristics at the eastern BoB boundary. The KWI effectively scales the strength of CKWs, and hence significant occurrences of CKWs can be detected and discussed. Meanwhile, the sign of CKW-related temperature changes can be used to distinguish between downwelling and upwelling CKWs. During the entire analyzed period, a total of 93 significant downwelling CKWs and 127 significant upwelling CKWs are detected based on the KWI greater than 15.86 in our case. The composite of significant CKWs reveals that the weakening of CKWs near the Irrawaddy Delta tip is a general feature, and the estimated general phase speed of CKWs in our research area is about $1 \text{ m/s}$. The BoB CKWs are highly in accordance with the intraseasonal zonal wind stress in the WEIO with a lag of about 25 days. Our model results demonstrate that the downwelling CKW lags the equatorial intraseasonal westerly winds, and the upwelling CKW lags the equatorial intraseasonal easterly winds.

The occurrence of significant CKWs varies significantly on different time scales. With respect to the long-term scale, the occurrence of significant CKW at the eastern BoB boundary is predicted to be more frequent in the future. Notably, with respect to the monthly climatology, the number of significant CKW events shows remarkable changes over time; unlike in the first two climate periods, significant CKWs mainly occur around August during the last two climate periods. The number of peaks revealed by the corresponding KWI time series also indicates that the seasonal variability of intraseasonal CKWs will be prominently affected under the high emissions pathway. The equatorial wind, which is expected to significantly affect the variability of BoB CKWs, exhibits corresponding monthly climatological variabilities. Therefore, it can be concluded that the changes in equatorial wind field caused by global warming probably affects the generation of EKWs in the EIO, and further influence the variability of CKWs in the BoB.

Previous studies suggest that the CKWs at the eastern boundary play an indispensable role in the generation of mesoscale eddies in the eastern and central basins (Chen et al. 2018; Cheng et al. 2018). Therefore, the predicted changes in CKWs under global warming will cause further changes in mesoscale eddy activities, which in turn will trigger other processes farther offshore and eventually significantly impact the BoB characteristics. The relationship between the intraseasonal/interannual variabilities of the sea level height along the BoB coast and equatorial/local winds has been discussed previously (Han and Webster 2002; Suresh et al. 2013), showing that the importance of equatorial winds is decreasing counterclockwise along the BoB coast, while
the importance of local winds is increasing. Our results regarding the intraseasonal CKWs at the pycnocline corroborate these conclusions and further demonstrate that in addition to these equatorial and local signals, the internal stratification of the BoB exhibits a prominent influence on CKW propagation. Along the counterclockwise direction of the BoB coast, the pycnocline depth generally changes from deep to shallow when going north, and then from shallow to deep when going south. Meanwhile, as the model results indicate, the pycnocline depth shows a more distinct variation along the BoB coast and, moreover, a clear response to climate change. Accordingly, a closer look at the vertical propagation of CKWs (Romea and Allen 1983; Nethery and Shankar 2007) is of definite importance to fully understand the full bandwidth of the underlying processes.

There are some other physical processes related to the BoB CKWs that need further study, especially the generation process of the intraseasonal EKWs (Nagura and McPhaden 2012; Pujiana and McPhaden 2020), the interaction between the Wyrtki Jets and the EKWs (Wyrtki 1973; Nyadjro and McPhaden 2014; McPhaden et al. 2015), and the feedback mechanism between the atmospheric MJO and the oceanic planetary waves (Zhang and Dong 2004; Webber et al. 2012; Sharmila et al. 2013; Krishnamurti et al. 2017). A high-resolution regional air-sea coupled model covering the entire tropical Indian Ocean would be required to perform a comprehensive investigation of these processes, and also of their interaction with CKWs and reflected Rossby waves. To conclude, the primary focus of our current study is the variability and characteristics of the intraseasonal CKW signals reflected at the pycnocline. In subsequent research, we plan to study how these CKWs interact with the current system and how they affect mesoscale eddies, the thermocline and other local features, especially in the context of climate change.
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