ABSTRACT: This study presents the characteristics and spatiotemporal structure of near-inertial waves and their interaction with Leeuwin Current eddies in the eastern south Indian Ocean as observed by Electromagnetic Autonomous Profiling Explorer (EM-APEX) floats. The floats sampled the upper ocean during July–October 2013 with a frequency of eight profiles per day down to 1200 m. Near-inertial waves (NIWs) are found to be the dominant signal in the frequency spectra. Complex demodulation is used to estimate the amplitude and phase of the NIWs from the velocity profiles. The NIW energy propagated from the base of the mixed layer downward into the ocean interior, following beam characteristics of linear wave theory. We visually identified a total of 15 near-inertial internal wave packets from the wave amplitudes and phases with a mean vertical wavelength of 89 ± 63 m, a mean horizontal wavelength of 69 ± 85 km, a mean horizontal group velocity of 3 ± 2 cm s⁻¹, and a mean vertical group velocity of 9 ± 7 m day⁻¹. A strong near-inertial packet with a kinetic energy of 20–30 J m⁻² found propagating below 700 m suggests that the NIWs can contribute to deep ocean mixing. A blue shift of 10%–15% in the energy spectrum of the NIWs is observed in the upper 1200 m as the floats move toward the equator. The impacts of mesoscale eddies on the characteristics and propagation of the observed NIWs are also investigated. The elevated near-inertial shear variance in anticyclonic eddies suggests trapping of NIWs near the surface. Cyclonic eddies, in contrast, were associated with weak near-inertial shear variance in the upper 400 m.

KEYWORDS: Currents; Diapycnal mixing; Eddies; Internal waves

1. Introduction

Internal waves are ubiquitous in the ocean with a vertical length scale of 10–100 m and time scales from minutes to hours. They occur at all frequencies between the inertial frequency (f) and the buoyancy frequency (N). The main sources of internal waves in the ocean are wind (D’Asaro 1985), barotropic tides over rough topography (Egbert and Ray 2000; Rudnick et al. 2003), and interaction of geostrophic flow with bottom topography (Nikurashin and Ferrari 2010). The internal waves with frequency close to f stand apart from the rest of the spectrum. These bands of internal waves, called near-inertial waves (NIW), are mostly generated from winds and have almost circular particle motions (anticlockwise in the Southern Hemisphere). They can also be generated by geostrophic adjustment at fronts and eddies (spontaneous generation; Gill 1984; Silverthorne and Toole 2009; Alford et al. 2013; Nagai and Hibiya 2015), nonlinear wave–wave interactions such as Parametric Subharmonic Instability (MacKinnon and Winters 2005), and lee wave formation over bottom topography by geostrophic flow (Nikurashin and Ferrari 2010). Most of the energy (Leaman and Sanford 1975; Garrett 2001; Wunsch and Ferrari 2004) and vertical shear (Alford et al. 2016) of the internal wave spectrum is in the near-inertial band. Since NIWs are ubiquitous in the global ocean, they are suggested to play a crucial role in mixing the deep ocean (Alford et al. 2012). Despite their intermittent nature, NIWs can propagate far from their source regions before breaking (Alford 2003; Zhao and Alford 2009) and potentially contribute to ocean variability and turbulent mixing.

Observations suggest that wind-generated NIWs can propagate into the ocean interior and contribute to deep ocean mixing (e.g., Silverthorne and Toole 2009). A total of 2 TW of energy is required to maintain the global overturning circulation and abyssal stratification in the ocean (Munk and Wunsch 1998). Tides contribute about 0.9 TW and wind contributes about 1.2 TW of energy, of which 0.2 TW is from wind-generated internal waves that radiate from the sea surface into the abyssal ocean (Munk and Wunsch 1998). Using simple mixed layer slab models, various studies have estimated the contribution of wind flux into near-inertial motions...
blue shift of the near-inertial peak exist in the World Ocean (mons and Alford 2012). A small number of observations of resulting in a blue shift of their frequency (Alford 2003; Simulation, which will be higher than that at the lower latitudes low latitudes where the local inertial frequency is lower than the NIW frequency. As they move equatorward, the NIWs reach a latitude where the local inertial frequency is equal to the NIW frequency. Then the NIWs propagate back toward the hemispheres, propagate downward and equatorward currents in Furuichi et al. (2008) estimated an annual wind energy input of 0.4 TW into near-inertial motions of which ~85% is dissipated in the upper 150 m. On the other hand, Alford et al. (2012) shows substantial deep penetration of NIWs with potential for mixing down to 800 m in mooring data. One reason for this discrepancy between this model estimates and observations could be the absence of mesoscale currents in Furuichi et al.’s model. Global studies using Argo floats suggest that regions of strong wind energy input into near-inertial motions are often associated with strong mesoscale activity (Whalen et al. 2018) and high parameterized turbulent diffusivity (Whalen et al. 2012).

The low-mode NIWs generated at a particular latitude, in both hemispheres, propagate downward and equatorward (Simmons and Alford 2012), and thereby contribute to mixing away from their generation sites (e.g., Meyer et al. 2016). The NIWs with frequency slightly higher than the local inertial frequency tend to propagate toward the poles and will quickly reach a latitude where the local inertial frequency is equal to the NIW frequency. Then the NIWs propagate back toward low latitudes where the local inertial frequency is lower than the NIW frequency. As they move equatorward, the NIWs tend to keep the local inertial frequency of their origin location, which will be higher than that at the lower latitudes resulting in a blue shift of their frequency (Alford 2003; Simmons and Alford 2012). A small number of observations of blue shift of the near-inertial peak exist in the World Ocean (e.g., Fu 1981; Simmons and Alford 2012).

Higher-mode NIWs interact with other oceanic phenomena of different scales such as fronts (e.g., Alford et al. 2013) and mesoscale eddies (e.g., Kunze 1985; Zhai et al. 2005), which can potentially lead to turbulence and mixing. The comparatively slower vertical and horizontal group velocities of these waves allow them to interact more strongly with the mesoscale which can affect their propagation pathways (Alford et al. 2016). Using a primitive equation numerical model, Lee and Niler (1998) found that anticyclonic eddies are efficient in draining near-inertial energy from the surface to below the thermocline whereas in a cyclonic eddy, the near-inertial waves can freely propagate out leaving energy only in the surface layer. Kunze (1985) proposed that when NIWs are generated inside an anticyclonic eddy, the relative vorticity of the eddy can reduce the frequency of the waves and thereby trap them. The anticyclonic vorticity of the eddies reduces the lower bound of the wave frequency to an effective frequency, $f_{\text{eff}} = f + \zeta/2$, where $f$ is the local Coriolis frequency and $\zeta$ is the background relative vorticity. As a result, the waves cannot freely propagate away from the eddy core. As the waves propagate downward from the surface, they reflect off the sides of the anticyclonic core and stall in a critical layer at the base of the eddy core. When the NIWs reach the critical layer, their group velocity and vertical wavelength reduce, and the wave energy density increases. Eventually, a part of the wave’s energy will be dissipated, enabling turbulence production and vertical mixing at the base of the eddy core (Kunze 1995). The approach of Kunze (1985) has been supported by both observational and modeling studies which show that mesoscale eddies can efficiently transfer near-inertial energy generated at the surface into the ocean interior through wave–mean flow interactions (e.g., Zhai et al. 2005; Elipot et al. 2010).

The eastern south Indian Ocean (SIO) has a rich mesoscale eddy field with a clear seasonal and annual cycle (Feng et al. 2005, 2007; Jia et al. 2011). The circulation of the eastern SIO consists of eastward flowing geostrophic currents in the upper 200–300 m, identified as the South Indian Countercurrent (SICC) between 20° and 30°S (Menezes et al. 2014) and an underlying westward current (Fig. 1; Palastanga et al. 2007; Siedler et al. 2006; Schott et al. 2009; Menezes et al. 2013, 2014); and poleward flowing Leeuwin Current (LC) which is accompanied by a northward flowing undercurrent, Leeuwin Undercurrent (LUC). The rich mesoscale field of the SIO is fed by the instabilities of both the SICC and LC systems (Feng et al. 2005, 2007; Zhang et al. 2020). In addition, the annual and semiannual Rossby waves emanating from the eastern boundary (Morrow and Birol 1998) also contribute to the mesoscale eddy field in the eastern SIO. The mesoscale eddy field offshore of the LC is unusually strong relative to other eastern boundary regions and contributes to large ocean to atmosphere fluxes of heat (Domingues et al. 2006).

Global studies show that weak to moderate wind work is done on mixed layer near-inertial motions in the eastern SIO (Alford 2003). The higher inertial horizontal kinetic energy during austral winter (Chaigneau et al. 2008) is consistent with the offshore migration of high eddy kinetic energy and long-lived anticyclonic eddies from the LC (Feng et al. 2007; Fang and Morrow 2003). Partly because of the paucity of observations, the NIW field in the eastern SIO is not well known. Moreover, it is even less understood how much of the near-inertial energy flux is locally dissipated through diapycnal mixing. Regional ocean model studies in the SIO suggest that the strength of the large-scale circulation is sensitive to the magnitude of vertical mixing (Furue et al. 2013; Benthuysen et al. 2014). Thus, more observations of near-inertial energy flux are needed to improve the representation of NIWs in regional and global ocean models. Due to the ubiquity of the mesoscale features in this region, the interaction between eddies and NIWs could be an important source of turbulent mixing in the ocean interior.

Data collected from an array of Electromagnetic Autonomous Profiling Explorer (EM-APEX) (Sanford et al. 2011) floats allow us to examine the near-inertial, mesoscale, and large-scale processes in the eastern SIO. The EM-APEX in this experiment collected on average eight profiles of velocity, temperature, and salinity each inertial period (=27 h) down to 1200-m depth between 100° and 110°E, 23°–32°S. Thus, they..
offer a unique opportunity to examine the internal wave field using both a time series analysis approach as used for mooring data (e.g., Alford et al. 2012) and drifter data (Chaigneau et al. 2008), and also through examination of the vertical propagation of internal waves that is possible with high vertical resolution data (Polzin 2008; Meyer et al. 2016; Waterman et al. 2021).

This paper investigates the characteristics of NIWs and their interactions with the mesoscale field in the eastern SIO. In addition, we attempt to estimate the fraction of near-inertial energy flux that propagates into the thermocline from the sea surface. We also look at the horizontal propagation and final dissipation of the waves by considering diapycnal mixing estimates from a finescale parameterization published in a companion paper (Cyriac et al. 2021).

The paper is organized as follows. In section 2, the characteristics of EM-APEX floats and other auxiliary data are described. The data analysis techniques used to identify the NIWs and a brief description of the regional environment are discussed in section 3. Section 4 describes the characteristics of the observed NIW packets whereas section 5 analyses the interaction of NIW with mesoscale eddies. Section 6 discusses these results and provides a conclusion of the study.

2. Data and regional environment

a. EM-APEX floats

1) DEPLOYMENT

The primary data used for this study (temperature, salinity, pressure, and horizontal velocity) are recorded by five EM-APEX floats deployed in the eastern SIO between 25° and 32°S along 105°E, in July 2013 (Fig. 1). The deployment was part of Voyage SS2013 V04 of the Australian Marine National Facility RV Southern Surveyor. The selected latitude band covers the strongest and deepest part of the eastward flows, the southern branch of the SICC (Menezes et al. 2014). Eddy kinetic energy is also high in this band surrounding 25°S (Jia et al. 2011).

The floats were deployed immediately following a full-depth CTD/lowered ADCP cast, as the ship moved away from the station. The five floats were spaced at intervals of 1.5° latitude along 105°E. In Fig. 1, the rich mesoscale environment is revealed in the float tracks, overlaid on a long-term surface geostrophic velocity field that highlights the background eastward flows. The shipboard CTD data were used as high-quality reference profiles to calibrate the EM-APEX CTD and velocity in the first few profiles. Further information about the voyage and the data processing report for the shipboard hydrographic data is available at https://www.cmar.csiro.au/data/trawler.

(i) EM-APEX background

The EM-APEX float is a combination of absolute velocity profiler (Sanford et al. 1978) and Argo profiler (Roemmich et al. 2004). In addition to the standard Argo components for measuring temperature, salinity and pressure, the EM-APEX includes a compass, accelerometer, five electrodes and external fins to rotate the float as it moves vertically. The electrodes measure the electrical potential difference across the float that is generated by the movement of seawater through
Earth’s magnetic field. The theory of motional induction then allows the depth varying ocean velocity relative to a depth-independent reference velocity to be calculated (Sanford et al. 1978). The reference velocity is found by estimating the displacement caused by the measured relative velocities along a path from the surface to the bottom of a down profile and back to the surface. The difference between this integrated displacement and that measured from GPS surface positions is then attributed to the reference velocity (see Phillips and Bindoff 2014).

The float tracks are analyzed using the SLA maps to identify float profiles associated with mesoscale eddies. SLA is interpolated to the position and time of each float profile to construct along-trajectory records of SLA. Anticyclonic (cyclonic) eddies are associated with positive (negative) SLA.

Daily sea level anomaly (SLA) and absolute surface geostrophic velocities are obtained from Copernicus Marine Environment Monitoring Service (CMEMS) on a 0.25° × 0.25° spatial grid. The daily SLA data are computed with respect to a 20-yr mean reference period (1993–2012). The data are obtained from https://cds.climate.copernicus.eu. The float tracks are analyzed using the SLA maps to identify float profiles associated with mesoscale eddies. SLA is interpolated to the position and time of each float profile to construct along-trajectory records of SLA. Anticyclonic (cyclonic) eddies are associated with positive (negative) SLA.

Hourly wind stress data are obtained from the National Centers for Environmental Prediction (NCEP) Climate Forecast System version 2. It is a fully coupled model in which the interaction between atmosphere, ocean, land, and sea ice is incorporated. The data have a spatial resolution of 0.5° × 0.5° (Kanamitsu et al. 2002).

Tides contribute to mixing in the ocean especially in regions of significant bathymetric features (Egbert and Ray 2000).
The barotropic tidal currents are extracted from the predic-
tions of TOPEX/Poseidon 7.2 (TPXO7.2) barotropic data-
assimilated global tidal model that uses the Laplace tidal
equations and along track altimeter data from TOPEX/Posei-
don satellites to estimate the depth averaged barotropic cur-
tections of TOPEX/Poseidon 7.2 (TPXO7.2) barotropic data-
expression, as described in the model results of Rennie et al.
(2007). EM-6663 tracked northward and eastward, orbiting
the western side of CE1, the southern edge of two anticy-
clonic eddies to the north (AE1 and AE2), and along the
western boundary of another cyclonic eddy (CE3) (Fig. 4b).
EM-6664 was deployed in a region of weak positive SLA with
anticyclonic vorticity. It moved westward, northward, and
then eastward on the southern side of a large anticyclonic
eddy (AE3) before becoming caught up in CE2 (Figs. 3
and 4). The excursions of the two shallow floats (down to
300 m), EM-6217 and EM-6218, is shown in Fig. S1 in the
online supplemental material.

MLD varies from 30- to 180-m depth (magenta line, Fig. 4b) and has no apparent dependence on the position of
the float relative to eddies. The maximum current speeds
experienced in cyclonic and anticyclonic eddies are up to
0.8 m s\(^{-1}\). Anticyclonic eddies have surface intensified cur-
rents, whereas cyclonic eddies have subsurface velocity maxi-
ma around 200-400 m. Strong vertical shear is observed in
the upper 400 m and below 26.9 kg m\(^{-3}\) isopycnal. The floats
also provide a finer picture of the different watermasses in
this region. From the float data, we can identify the warm,
salty subtropical underwater (STUW), the layer of potential
vorticity minimum, the Subantarctic Mode Water (SAMW,
26.7-26.9 kg m\(^{-3}\), and the layer of salinity minimum, the
Antarctic Intermediate Water (AAIW, 27.1-27.3 kg m\(^{-3}\)).

A more detailed description of the regional characteristics
observed along the float tracks and shipboard stations of
SS2013 V04 is given in Cyriac et al. (2021).

3. Methods

a. Near-inertial signal

Velocity rotary spectra can reveal the characteristics of vari-
ability at different time scales and decomposes the velocity
vector into counterclockwise (CCW) and clockwise (CW)
rotating circular components (Leaman and Sanford 1975; Eli-
pot and Lumpkin 2008). We treated the velocity data from
each float as a time series and examined the rotary spectrum
of velocity on depth surfaces. The data were divided into
30-day moving windows with 30% overlap and averaged to
create a mean rotary spectrum for each float. In the resulting
velocity rotary spectra, there is a prominent peak at the local
inertial frequency in all the floats (Fig. 5). The floats covered
a wide range of inertial frequencies (Table 1) since they
moved in a 7° latitude window. Additional peaks at diurnal

### Table 1. Deployment information and the inertial band covered by each float based on the latitudinal extent of their tracks.

<table>
<thead>
<tr>
<th>Float</th>
<th>6664</th>
<th>6663</th>
<th>6218</th>
<th>6662</th>
<th>6217</th>
</tr>
</thead>
<tbody>
<tr>
<td>Longitude range</td>
<td>104.2°–107.5°E</td>
<td>103.5°–106.6°E</td>
<td>103.7°–105.2°E</td>
<td>104.3°–107.6°E</td>
<td>104.9°–108.5°E</td>
</tr>
<tr>
<td>Depth (dbar)</td>
<td>1200</td>
<td>1200</td>
<td>300</td>
<td>300</td>
<td>300</td>
</tr>
<tr>
<td>No. of profiles</td>
<td>520</td>
<td>518</td>
<td>1058</td>
<td>528</td>
<td>528</td>
</tr>
<tr>
<td>f range (cpd)</td>
<td>0.95–1.0</td>
<td>0.80–0.97</td>
<td>0.94–1.0</td>
<td>0.90–0.99</td>
<td>0.8–0.95</td>
</tr>
</tbody>
</table>
(O₁, K₁), semidiurnal (M₂), and higher harmonics (f + M₂, M₂ + S₂) are also observed.

Since the diurnal tidal frequencies are very close to the local inertial frequency in this region, it can be difficult to separate the wind-generated near-inertial oscillations from the barotropic and baroclinic diurnal tides. We thus first removed the tidal contribution by subtracting the depth averaged barotropic tidal velocity (TPXO7.2 model) from the float-measured horizontal velocity corresponding to each location and time of the float, following Elipot et al. (2010). The tidal velocity arising from the eight primary tidal constituents estimated along the float track using the TPXO 7.2 tidal inversion model is quite small (0.01–0.02 m s⁻¹) compared to the amplitude of the float-measured velocities (0.33–1.4 m s⁻¹). The spectrum of the detided velocity is indistinguishable from the spectrum of total velocity (not shown), indicating that the peak at diurnal frequency is primarily due to near-inertial motions rather than diurnal tides. The baroclinic tides arising from the diurnal and semidiurnal tides could survive the detiding process. However, the eastern south Indian Ocean is a region of weak energy flux into internal tides and weak barotropic to baroclinic tidal energy conversion (Melet et al. 2013; Ansong et al. 2017). Furthermore, the amplitude of the semidiurnal tides in the tidal analysis in this region are significantly larger than the amplitudes of the diurnal tides and the distinct semidiurnal tides are themselves much weaker than the peak at the inertial frequency (Fig. 5). These two lines of evidence suggest the energy near diurnal frequency is most likely to be primarily due to inertial oscillations and is the main assumption underlying the complex demodulation approach described in the next section.

While the TPXO model is widely used to predict barotropic tidal velocities (e.g., Elipot et al. 2010; Alford 2010), tidal models can have imperfect amplitude and phase of the tidal...
constituents. We thus compared the barotropic tidal velocities from the TPXO7.2 tidal model with a harmonic analysis of the EM-APEX velocities using the MATLAB tidal analysis software UTide (Codiga 2011). For the UTide analysis, we treat the EM-APEX data as a time series at the mean latitude of the float track. There is no perceptible difference between the spectrum of velocity detided using the TPXO tidal velocities and that detided using the tidal velocities reconstructed from harmonic analysis of the EM-APEX velocities (not shown). This is the case despite the latitude variations experienced by the EM-APEX floats. We are thus confident that the TPXO7.2 tidal model is efficient in predicting the tidal velocities in the study region.

b. Complex demodulation

Complex demodulation is the process of extracting a particular frequency component from a velocity or scalar time series to determine the temporal variation at that frequency. A general approach to complex demodulation is to use the least squares method to fit the desired parameters (e.g., velocity) to sequential segments of the time series data. An advantage of this approach is that the data need not be regular in time and the inertial frequency can be explicitly targeted. The method separates the velocity components for a particular frequency into counterclockwise (CCW) and clockwise (CW) rotating circular components (Gonella 1972; Leaman and Sanford 1975). Depending on the hemisphere, only one of the above rotary components (CW component in the Northern Hemisphere and CCW component in the Southern Hemisphere) dominates the inertial currents (Elliot and Lumpkin 2008; Leaman and Sanford 1975; Martini et al. 2014; Thomson and Emery 2014). Thus, we can then analyze a single rotary component rather than two scalar components.

Here we estimate the amplitudes and phases of the internal waves at near-inertial frequency using complex demodulation of the observed velocity components. The velocity profiles were Wentzel-Kramers-Brillouin (WKB) scaled, following Leaman and Sanford (1975) as \( u_{wkb} = u/\sqrt{N(z)/N_0} \). A WKB stretched depth is also calculated as \( z_{wkb} = \int_0^z N(z)/N_0 \, dz \) to account for the wavenumber changes due to stratification. Here, \( N_0 \) is a reference buoyancy frequency of 3 cph. We then applied the complex demodulation to the velocity time series in each depth level at the center time \( t_c \) in a window of \( \pm 1 \) inertial periods (\( \pm 48 \) h, 16 profiles). The amplitude and phase of the inertial component in the velocity data are extracted using back rotation of the velocity vector with respect to a reference time \( t_0 \) (1 July 2013) before the floats were deployed (D’Asaro et al. 1995). Back rotation reduces the space–time aliasing which may occur due to the rotation of the near-inertial currents during long periods of sampling. This method removes the temporal phase propagation that distorts the near-inertial currents (Martínez-Marrero et al. 2019). Here, the phase of the inertial currents at \( t \) correspond to the direction of inertial currents at \( t_0 \). The measured velocity component can be expressed as

\[
U_c = A_c^+ \exp\left\{ i \left[ f_c (t_1 - t_0) + \epsilon_c^+ \right] \right\} \\
+ A_c^- \exp\left\{ -i \left[ f_c (t_1 - t_0) + \epsilon_c^- \right] \right\} \\
+ \nabla_c^+ \cdot \nabla_c^- + a_{uc} (t_1 - t_c) + i a_{uc} (t_1 - t_c),
\]

where \( U \) is the complex velocity \( u + iv \) and \( u \) is the eastward (northward) velocity component at each depth level \( c \). Parameters \( (A^+, A^-) \) are the amplitudes and \( (\epsilon^+, \epsilon^-) \) are the phases of the CCW and CW rotating inertial components, respectively, at each depth. The time index in each window, \( j \), varies from 1 to \( n \), where \( n \) is the number of data points in the chosen window. The demodulation method is based on the assumption of a “quasi-steady” state so that the wave properties do not change much over the period of the window (Federik and Allen 1996). The floats are moving in space, so we take \( f_c \) to be the average of local inertial frequencies in each window. The terms \( \nabla_c^+ \cdot \nabla_c^- \) are the mean eastward and northward velocities in each window. The last two terms allow for a linear low-frequency trend in the background current in each window, and \( a_{uc} \) and \( a_{uc} \) are the slopes of the trend in the eastward and northward direction. The percentage of the variance captured by the model can be written as

\[
R^2 = 1 - \left( \frac{\sigma_{res}^2}{\sigma_{tot}^2} \right),
\]

where \( \sigma_{res}^2 \) is the variance of the model residual and \( \sigma_{tot}^2 \) is the variance of the total velocity in each moving window. The average \( R^2 \) value for the data from all the floats is 0.57 (0.72 for float EM-6662), implying that 57% (72%) of the variance in the observed velocities can be explained by slowly varying near-inertial waves.

The wave frequency at the location of each profile using this method can be obtained as (Federik and Allen 1996)

\[
\omega_k = f_c - \frac{\partial \epsilon_c^+}{\partial t},
\]

and the vertical wavenumber as

\[
m_c = \frac{\partial \epsilon_c^+}{\partial z}.
\]

The dispersion relation for the internal wave can be written as

\[
k_h^2/m_c^2 = \frac{f_c^2 - \omega_0^2}{\omega_0^2 - N^2},
\]

where \( \omega_0 \) is the intrinsic frequency and \( k_h \) is the horizontal wavenumber \( (k_h = \sqrt{k_l^2 + F}) \).

From the dispersion relation, \( k_h \) can be obtained as

\[
k_h = \pm m_c \frac{f_c^2 - \omega_0^2}{\omega_0^2 - N^2}.
\]

The group velocity \( C_g \) is given by the gradient of the frequency with respect to the wavenumber. By implementing the hydrostatic approximation, assuming \( \omega_0^2 \ll N^2 \), the magnitude of the horizontal group velocity can be obtained as
FIG. 4. Along-stream evolution of (a) sea level anomaly (blue line), relative vorticity (green line) and DHT (pink), (b) Conservative Temperature, (c) Absolute Salinity, (d) speed, (e) buoyancy frequency, and (f) vertical shear of horizontal velocity. The light gray lines are isopycnals with an interval of 0.7 kg m\(^{-3}\). The heavy gray contours in all panels show the density range of AAIW (27.1–27.3 kg m\(^{-3}\)), and heavy black lines show SAMW (26.7–26.9 kg m\(^{-3}\)). The STUW is the high salinity near-surface water.

EM-6662 profiled two cold core eddies (CE1 and CE2). EM-6663 profiled two warm core eddies (AE1 and AE2). EM-6664 profiled another warm core (AE3) and the same cold-core eddy in EM-6662 (CE2). The evolution of the mixed layer along the float tracks are marked over temperature and salinity (magenta line). The horizontal extent of each cyclonic (anticyclonic) eddy is marked with cyan (red) horizontal lines at the top of (a) and is projected through all panels.
\[ C_{gh} = \sqrt{\left( \frac{kN^2}{\alpha_0 m_c^2} \right)^2 + \left( \frac{\hat{N}N^2}{\alpha_0 m_c^2} \right)^2}, \]  

(6)

where \( k = k_h \sin \epsilon \) and \( l = -k_h \cos \epsilon \) are the horizontal wave vector components.

The vertical group velocity can be written as

\[ C_{gz} = \frac{\alpha_0^2 - f^2}{\alpha_0 m_c}. \]  

(7)

c. Bandpass filtering

As a check on the effectiveness of the complex demodulation in resolving near-inertial frequency variations, we bandpass filtered the detided velocity from the float data in the time domain. First, we interpolated the velocity data onto 3-h intervals at all depths. We then applied a minimum order digital bandpass filter on each time series at each depth. The pass band frequency limit is guided by the mean Coriolis frequency of each float track (Table 1). The pass band for float EM-6662 corresponds to the frequency range of 0.96\( f \)–1.05\( f \), for EM-6663 corresponds to the frequency range of 0.88\( f \)–1.08\( f \) and for EM-6664, the pass band frequency ranges from 0.96\( f \) to 1.01\( f \), where \( f \) is the mean Coriolis frequency of each float track.

Figure 6 shows the separation of the detided velocity into bandpass filtered amplitudes in the near-inertial band for each float along with the complex demodulated amplitudes. The bandpass filtered near-inertial amplitudes show strong wave motion in regions where we observe strong NIW amplitudes from the complex demodulation (Fig. 6, middle and bottom panels). There are some differences between the two as well. For instance, the complex demodulated signal is larger in amplitude and smoother than the bandpass filtered data by a factor of 2. Part of this difference is because the complex demodulated estimates of the rotary spectra target the near-inertial frequency over a 48-h period and also take into account the trends in the mean velocity field. The bandpassed estimates do not take the trends and are potentially noisier estimates compared with the complex demodulated estimates. Nevertheless, the complex demodulation is capturing the near-inertial variations at the right frequencies most of the time and likely to be a more reliable estimate than the raw bandpassed data.

d. Coherent features

Internal waves can also be identified from coherent features in vertical profiles of horizontal velocity (Polzin 2008; Meyer et al. 2016; Waterman et al. 2021). Following Meyer et al. (2016), we identified coherent features in velocity profiles at depths corresponding to the packets of high near-inertial amplitude by plotting anomalies of \( u \) and \( v \) relative to the vertically smoothed profile. The smoothed profiles were created using a vertical moving average window of 500 m. Figure 7 shows an example of such a packet from EM-6662 between 650- and 1100-m depth. Elevated NIW amplitudes from the demodulation are found between profiles 440 to 520. Figure 7 zooms in on profiles 472–480 to provide a clearer view. The upward phase propagation implies a downward energy propagation suggesting that the packet was likely generated at the sea surface. This packet is discussed more in section 4c. We observe similar coherent features in the velocity profiles corresponding to other regions of high NIW amplitude identified through the complex demodulation.

e. Relative vorticity and mixed layer depth

The vertical component of the relative vorticity (\( \zeta \)) can be written as

\[ \zeta = \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y}. \]  

(8)
where $u$ ($v$) is the daily surface geostrophic eastward (northward) velocity component obtained from satellite altimetry. The $\zeta$ values are then subsampled at the time and location of each float profile to construct the along-trajectory variations. Relative vorticity helps us to identify the mesoscale eddies in the float tracks where Southern Hemisphere anticyclonic (cyclonic) core eddies have positive (negative) $\zeta$ values within their perimeter.

The mixed layer depth (MLD) is defined as the depth at which the potential density changes by 0.03 kg m$^{-3}$ relative to that at 15-m depth (de Boyer Montegut et al. 2004) for all floats except EM-6217. Since the salinity data from EM-6217 were erroneous (Fig. S1), we estimated the MLD using a temperature criterion of 0.2$^\circ$C threshold (de Boyer Montegut et al. 2004).

f. Wind energy flux and near-inertial wave energy

The amount of energy transferred from wind to the near-inertial motions is estimated as

$$\Pi = \tau U_{in},$$

where $\Pi$ is the power per unit area or the energy flux, $\tau$ is the wind stress, and $U_{in}$ is the mean near-inertial velocity with components $(u_{in}, v_{in})$ in the mixed layer (Silverthorne and Toole 2009) estimated from the complex demodulation.

When the wind stress and the inertial currents are in the same direction (positive power), the energy will transfer from the atmosphere into the ocean. The energy flux is negative when the wind stress is in the opposite direction to the surface current, which will increase the near-surface shear in both atmospheric and oceanic boundary layers resulting in turbulent dissipation (D’Asaro 1985). Further, the wind work input into near-inertial motions can be estimated as a time integral of the energy flux, $\int \Pi dt$.

The near-inertial kinetic energy (KE$_{in}$) along the float tracks is calculated as

$$KE_{in} = \frac{1}{2} \rho_0 (u_{in}^2 + v_{in}^2),$$

where $\rho_0$ is the density (Silverthorne and Toole 2009).

4. Observations of near-inertial internal waves

The isopycnal heights displayed in Fig. 4 clearly show high-frequency fluctuations with depth ranges of tens of meters. The current speed in Fig. 4d also shows alternating strong and weak currents every few profiles. In the following we isolate these near-inertial oscillations from the lower-frequency variability using complex demodulation.

Using complex demodulation, we obtained the CCW and CW components of the NIWs along the float tracks. Most of the strong packets are observed near the surface (Fig. 8).
suggestive of wind-forced near-inertial motions. The CCW component of the amplitudes are dominant compared to the CW component as is expected for the Southern Hemisphere. After the passage of a storm, wind-generated NIWs are known to propagate both horizontally and vertically from the mixed layer (e.g., Zervakis and Levine 1995). We observe strong packets that span a range of isopycnals near the surface and also at depth (e.g., EM-6662, 800 m depth). Comparatively weak packets propagate along the isopycnals (e.g., EM-6663). Here we consider packets with amplitudes larger than 0.1 m s$^{-1}$ in the CCW component, the global average of NIW amplitude (Chaigneau et al. 2008), to be strong packets.

**a. Wave properties**

Several downward propagating packets can be identified from the NIW velocities, reconstructed from the amplitude and phase from complex demodulation of the deep floats (Fig. 9). The group velocities of the packets are similar to those found in the North Pacific (Alford et al. 2012). The packets were identified by visual inspection in close-up plots of amplitude and corresponding coherent features in velocity profiles (Fig. 9). We identified a total of 15 clear NIW packets with a mean amplitude of 10 ± 3 cm s$^{-1}$ from all the floats and document their properties in Table 2. The properties of the packets from EM-6217 were not estimated since the float’s salinity measurements were erroneous. We first estimate the wave frequency and vertical wavenumber for each profile using Eqs. (2) and (3). We consider this frequency as the intrinsic wave frequency, $\omega_0$, since the float follows the mean flow. This is a major difference between drifting and moored observations where one has to infer $\omega_0$. Then we use $\omega$ and $m$ in Eqs. (5) and (6) to estimate the horizontal wavenumber and group velocity. Finally, we average the profiles that span each packet to get the wave properties.

The properties are highly sensitive to the $\omega$ and $m$ estimates. The mean and standard deviation of the properties are shown in Table 2. The low standard deviation shows that all of the waves have similar frequency and vertical wavenumber. Most of the packets are observed within the upper 500 m and very few packets are observed below the thermocline. Packet 10 has the largest vertical wavelength of 180.6 m whereas packet 13 has the smallest vertical wavelength (17.2 m). Similarly, packet 15 has the largest horizontal wavelength (346.9 km) whereas packet 14 in the thermocline has the smallest wavelength of 9.4 km. The packets have a mean frequency of $(0.98 \pm 0.09)$/ with a vertical wavelength of $89 \pm 63$ m. The mean horizontal wavelength of the packets is $69 \pm 85$ km and horizontal group velocity of $3 \pm 2$ cm s$^{-1}$. The packets have a mean vertical group velocity of $9 \pm 7$ m day$^{-1}$.

**b. Wind work and kinetic energy distribution**

The wind energy flux into near-inertial motions, $\Pi$, is quite intermittent along the float tracks with large energy inputs during strong wind events (Figs. 10a–c). The mean value of $\Pi$ is
0.22 mW m\(^{-2}\) ranging between a maximum value of 78.4 mW m\(^{-2}\) and a minimum value of 80.6 mW m\(^{-2}\) over the duration of four months. Even though it is small, the mean positive value of \(P\) suggests that there is a net energy transfer from wind to inertial motions in the mixed layer. Higher wind energy input often generates strong near-inertial currents in the mixed layer which then excite NIWs that propagate into the ocean interior. Below the mixed layer, strong NIW amplitudes are often seen.

**Fig. 8.** Near-inertial amplitudes of (a) CCW and (b) CW components along the float tracks. The isopycnals and the watermasses are the same as in Fig. 4. The dashed box in (a) highlights a strong packet of NIW amplitude. The red dashed lines inside the black box represents profiles that are examined in Fig. 7. Note that the x axis in each column represents the time period unique to each float between 10 Jul 2013 and 25 Oct 2013.

**Fig. 9.** The meridional component of the near-inertial currents (shading) from complex demodulation. The vertical velocity of each event estimated using linear wave theory is identified with a black line oriented along their direction of propagation. Only packets with amplitudes larger than 0.1 m s\(^{-1}\) are highlighted. The mixed layer depth is contoured in blue. Note that the x axis in each column represents the time period unique to each float between 10 Jul 2013 and 25 Oct 2013.
below cyclonic (e.g., EM-6662, profile numbers from 400 to 450) and within anticyclonic (e.g., EM-6663, profile numbers from 900 to 1000) eddies. The cumulative energy input or the wind work has a step like structure which increases only during strong wind events (Fig. 10d).

Strong wind events, driving elevated wind work, are associated with high KEin near the surface. High KEin (∼30 J m⁻³) is associated with stronger inertial currents and energy transfer as expected (Fig. 11). These high values are of similar magnitude with those observed in mooring data in the North Pacific (Alford et al. 2012; Plueddemann and Farrar 2006). Below the mixed layer, energy is weaker (<10 J m⁻³) except within strong packets. The depth-integrated KEin varies from 0.2 to 3.4 J m⁻³ with a long-term mean of 1.1 J m⁻³ (Fig. 11a). The time mean KEin from all the profiles from all the deep floats decreases in energy from 3.6 to 0.4 J m⁻³ at 1200 m (not shown). There are few subsurface increments in the time mean energy at 140 m (2.3 J m⁻³) and between 650 and 1000 m (0.9 J m⁻³) due to the presence of strong packets at depth.

c. Case studies

1) UPPER OCEAN

Impulsive wind stress imparts energy into the ocean surface mixed layer generating near-inertial oscillations. The convergences and divergences of these oscillations perturb the base of the mixed layer and transfer energy into downward propagating NIWs (D’Asaro 1985; Alford et al. 2016). Strong wind events often generate large inertial velocities in the mixed layer (Fig. 10). Moderate wind events also result in comparatively large velocities. One such packet was observed by float EM-6663 by end of September (Fig. 8) immediately below the mixed layer. Even though the wind stress is moderate (∼0.1 N m⁻²) in the region at the time of these profiles, this packet has a strong amplitude of 0.13 m s⁻¹ and higher inertial energy (>15 J m⁻³, Fig. 11). It has a vertical wavelength of 51 m and a horizontal wavelength of 120 km (Table 2, packet 7). This packet is also visible from the velocity profiles as a coherent feature similar to that in Fig. 7. The nonassociation of strength between wind and inertial amplitudes could be because the wind components and the inertial currents are out of phase and their lateral structures do not match (Alford et al. 2012).

2) DEEP OCEAN (500–1200 m)

We observe high KEin below 700 m associated with a downward propagating NIW (Fig. 7). This is the only deep reaching packet with energy >15 J m⁻³ which is similar in magnitude to the packets near the surface. It has a vertical wavelength of 191.4 m, horizontal wavelength of about 34.54 km and vertical group velocity of 8.23 m day⁻¹ (Table 2, packet 11). Since the packet does not seem to propagate from the surface in the float track (Fig. 8), it must have been generated at another place at some early time and propagated downward and equatorward (Zervakis and Levine 1995; Alford and Gregg 2001) to its observed position and depth. With a vertical group velocity of 8.2 m day⁻¹, this packet would take about 85 days to reach 700 m. By allowing 2–5 days for internal wave downward propagation to begin, we looked at the surface maps of wind speed around that time (Fig. 12). From the NCEP reanalysis winds, we identified a cyclonic wind system that evolved over several days (13–15 July 2013) with southerly wind speeds larger than 10 m s⁻¹, and then faded away. The packet observed by EM-6662 could have been generated

Table 2. The NIWs and their properties derived from the float data. The float number, amplitude, frequency $\omega$, $\omega/\omega_f$, vertical wavenumber $m$, vertical wavelength $\lambda$, horizontal wavenumber $K_h$, horizontal group velocity $C_{gh}$, vertical group velocity $C_{gg}$, depth, and profile numbers in which the waves are observed. The mean and standard deviation of the properties of the packets are at the bottom of the table.

<table>
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<th>No.</th>
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<th>Depth (m)</th>
<th>Amplitude (cm s⁻¹)</th>
<th>$\omega \times 10^{-5}$ (rad s⁻¹)</th>
<th>$\omega/\omega_f$</th>
<th>$m$ (cm)</th>
<th>$K_h \times 10^{-5}$ (rad m⁻¹)</th>
<th>$\lambda$ (m)</th>
<th>$C_{gh}$ (cm s⁻¹)</th>
<th>$C_{gg}$ (m day⁻¹)</th>
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<td>0.03</td>
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<td>—</td>
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from this passing wind system and propagated toward lower latitudes where the local inertial frequency is lower than that at the wave’s origin site (Alford 2003; Simmons and Alford 2012).

If we further assume that the vertical flux of the deepest packet transiting 700 m, $3.2 \times 10^{-4}$ W m$^{-2}$ is dissipated in the immediate 100-m depth below (AAIW layer, Fig. 4), it will result in a dissipation rate of $\epsilon = F_m/\rho \times 100 = 3.15 \times 10^{-9}$ m$^{-2}$ s$^{-3}$, where $\rho = 1027.13$ kg m$^{-3}$, with an associated diffusivity of $K_v = \Gamma\epsilon/N^2 = 4.2 \times 10^{-5}$ m$^2$ s$^{-1}$. This is of the same order of magnitudes for dissipation rate and diffusivity, respectively, as that estimated through finescale parameterization at this depth range for the corresponding profiles (Cyriac et al. 2021). Further, following Alford et al. (2012), if we assume that the vertical flux of this packet transiting 700 m is evenly distributed up to 1200 m of the water column, it will result in a very moderate dissipation rate of $2.63 \times 10^{-10}$ m$^2$ s$^{-3}$ and a diffusivity of $5.6 \times 10^{-6}$ m$^2$ s$^{-1}$. This suggests that NIWs play an important role in providing energy to the deeper levels in regions where the surface currents are not very strong.

d. Meridional propagation

The inertial waves observed at a given location can be a combination of locally generated (local wave field) waves in a response to local wind forcing and remotely generated waves that have propagated to that location from their generation site (global wave field) (Fu 1981; Alford 2003). We thus estimated the blue shift along the float trajectories and with depth. We calculated the velocity rotary spectra over a moving window of width 30 days along the float trajectory and then compared the peak inertial frequency with the mean local inertial frequency of the window, following Simmons and Alford (2012), at different depth levels. The blue shift is evident from the float EM-6663 which covered the most latitudes (Fig. 13a). The ratio of $f_{\text{peak}}/f_{\text{local}}$ increases toward the equator and reaches up to 10%–15% of the local inertial frequency. There is a tendency of the blue shift to increase with depth but not very strongly. We also see a small tendency for weak red shift especially in the upper layers (up to 1000 m). This is in agreement with the mooring observations of blue

FIG. 10. (a) Wind stress components from NCEP reanalysis subsampled along the float tracks, (b) near-inertial currents averaged over the mixed layer, (c) wind energy flux into near-inertial motions ($\Pi$), and (d) time integral of the wind energy flux along the float tracks. The pink dashed lines in (c) are one standard deviation. The location of anticyclonic (AE) and cyclonic (CE) mesoscale eddies along the float track are marked. Note that the x axis in each column represents the time period unique to each float between 10 Jul 2013 and 25 Oct 2013.
shift in Simmons and Alford (2012). The horizontal velocity spectra at different depths show a decrease in power with increase in depth as expected (Fig. 13b). The peak around $f$ is broad most likely due to the presence of a strong mesoscale background field and latitudinal variations along the float track. The peak varies at different depths, but we do not have enough depth ranges to clearly show the equatorward shift in $f$ peak with depth.

We further looked at the implied latitude of origin for the wind system which generated the deep NIW packet in Fig. 8 [section 4c(2)]. At the latitude of the NIW packet (~29°S), the ratio $f_{\text{peak}}/f_{\text{local}}$ is less than 1, suggesting a red shift. With a projected value of 0.9 for $f_{\text{peak}}/f_{\text{local}}$, this would lead to a latitude of origin of ~26°S, which matches with the wind system that we tracked (Fig. 12).

5. Interaction of NIWs and mesoscale eddies

Anticyclonic eddies can reduce the propagating frequency of the NIWs to an effective frequency, $f_{\text{eff}} = f + \zeta_z/2$, where $\zeta_z$ is the vertical component of relative vorticity. In the Southern Hemisphere, the cyclonic vorticity regions have $(f_{\text{eff}}/f) > 1$ whereas the anticyclonic vorticity regions have $(f_{\text{eff}}/f) < 1$ (Figs. 14a,c). The rich mesoscale field of the eastern SIO has strongly influenced the movement of the floats (Fig. 3). The floats profiled during July–October, during which the LC anticyclonic eddies have higher amplitudes, eddy kinetic energy and relative vorticity (Zhang et al. 2020). Even though none of the floats profiled across an eddy since they follow streamlines, we have two floats profiled along the periphery of anticyclonic eddies which is evident from the deepening of the isopycnals and strong relative vorticity (Figs. 4a,b).

In Figs. 14b and 14d, we plot the near-inertial shear variance, $S_{\text{in}}^2 = \left(\frac{\partial u_{\text{in}}}{\partial z}\right)^2 + \left(\frac{\partial v_{\text{in}}}{\partial z}\right)^2$, where $u_{\text{in}}$ and $v_{\text{in}}$ are the zonal and meridional components of NIW velocities, respectively, and $dz$ is taken as 10 m, for floats EM-6663 and EM-6664 and overlay the locations of cyclonic (CE) and anticyclonic (AE) mesoscale eddies. The panels above (Figs. 14a,c) show the corresponding $(f_{\text{eff}}/f)$. The near-inertial shear variance is higher when there is an anticyclonic eddy present (Fig. 14). To illustrate the characteristics of near-inertial shear variance in mesoscale eddies, we calculate the mean vertical profile of near-inertial shear variance for all profiles within AE3 and CE2 (Fig. 15). Near-inertial shear is clearly elevated in the upper 400 m within the anticyclonic eddy. This is in agreement with Elipot et al. (2010) where they found that the near-inertial variance is higher in anticyclonic vorticity regions compared to cyclonic vorticity regions, using global drifter and altimetry data. Lee and Niiler (1998) also found high inertial energy shear in the core of an anticyclonic eddy in their primitive equation model, suggesting that the NIWs were trapped inside the anticyclonic eddy (e.g., Kunze 1985; Lee and Niiler 1998; Jaimes and Shay 2010). Here we examine two floats (EM-6663 and EM-6664) that encountered anticyclonic eddies in their tracks.

a. EM-6663

The anticyclonic eddy (AE1) encountered on the track of EM-6663 has a mean relative vorticity of $1.3 \times 10^{-5}$ s$^{-1}$. The eddy vorticity reduces the effective frequency to about 90% of the local inertial frequency (Fig. 14a). This suggests that the anticyclonic eddy could trap the wind generated NIWs by reducing their frequency. This is also evident from the increased near-inertial shear variance in the region of anticyclonic eddies AE1 and AE2 (Fig. 14b), although there is no apparent reduction in effective frequency for AE2. The presence of NIWs can also be seen by simply inspecting the

![Fig. 11](image-url). (a) Depth integrated near-inertial KE along the float tracks. (b) The variation of KE$_{\text{in}}$ with depth along each float track. The locations of anticyclonic (AE) and cyclonic (CE) mesoscale eddies along the float track are marked.
near-inertial velocity profiles associated with the anticyclonic eddies. There is also a clear amplification of $u_n$ and $v_n$ in the upper 400 m with a wave-like pattern as in Fig. 7 (not shown).

b. EM-6664

The part of the anticyclonic eddy (AE3) sampled by the float is comparatively weak with a mean relative vorticity of $3.1 \times 10^{-6}$ s$^{-1}$. Since the float does not have any profiles in the eddy core, we cannot examine the theory of inertial chimney (Lee and Niiler 1998) where the strong relative vorticity core of the eddy acts as a critical layer where the near-inertial energy reaches its maximum. However, the background vorticity has influenced the inertial frequency (Figs. 14c,d). The ratio of $f_{\text{crit}}$ to $f$ is less than 1 for part of the anticyclonic eddy region of the float track where the isopycnals are deepest suggesting that the vorticity of the eddy reduces the frequency of NIWs resulting in trapping them and leading to the elevated NIW shear variance within AE3 (Fig. 14d). Note that the ratio of $f_{\text{crit}}/f < 1$ is not valid everywhere in AE3. We suspect this discrepancy is due to the low resolution of the relative vorticity estimates from altimetry data compared to the high-resolution data from the floats. However, we can see amplified wave-like patterns in the velocity profiles suggesting the presence of NIWs (Cyriac et al. 2021, their Fig. 11). Furthermore, there is a strong wind energy input ($\sim 50$ mW m$^{-2}$) into near-inertial motions in region of strong anticyclonic vorticity of AE3 (Fig. 10c). This is in agreement with the near-inertial wave trapping scenario of Kunze (1985) where the waves are required to originate inside the regions of anticyclonic vorticity.

Moreover, the near-inertial shear variance is distinctively higher within the AAIW layer than within the SAMW layer (Figs. 14c,d). This is in agreement with Cyriac et al. (2021), where they noted that the diapycnal diffusivity estimates were

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**Fig. 12.** Daily maps of surface wind speed (color shading) with the arrows showing wind direction during the period 11–18 Jul 2013 when the deep reaching NIW packet observed by float EM-6662 was likely generated. The track of the float EM-6662 is also marked (black line). The red dots are the float locations at each day, and the pink triangle is where the packet was observed.
lower in the SAMW layer of anticyclonic eddies than at other depths along the float tracks. This suggests the possible role of NIWs in watermass transformations of AAIW layers in this region.

6. Discussion and conclusions

The observations of NIWs in the eastern SIO and their interactions with LC eddies are presented for the first time.
down into the ocean interior. We observed stronger near-inertial shear variance in anticyclonic eddies near the surface compared to cyclonic eddies suggesting that NIWs are trapped inside anticyclonic eddies. This suggests that the presence of anticyclonic eddies increases the likelihood that wind energy will penetrate into the interior of the ocean (Fig. 11b).

The NIW characteristics obtained in this study are in good agreement with previous studies in different regions (Table 3). Compared to the Northern Hemisphere, NIW characteristics are less studied in the Southern Hemisphere. Using global drifter data from 1999 to 2006, Chaigneau et al. (2008) estimated a mean inertial current amplitude of 10 cm s$^{-1}$ in every ocean basin. They found a seasonal cycle in the inertial amplitude with higher amplitudes during July–September in the southern Indian Ocean, during which time the floats were deployed.

From our study in the eastern SIO, we observe near-inertial current amplitudes larger than 20 cm s$^{-1}$, which is stronger than those observed in the North Pacific subtropical front (Kunze and Sanford 1984; Alford et al. 2013). D’Asaro et al. (1995) reported amplitudes and energy 2–3 times stronger in the North Pacific where the NIWs were generated in the wake of a strong, isolated storm. They observed the inertial energy spreading downward from the mixed layer, and reaching as deep as 1000 m. This is consistent with our study where we observed a NIW packet propagating below 700 m and that of Alford et al. (2012) in the North Pacific where they found that the inertial energy propagated to depths of 800 m. They also found that the KE$_{in}$ is higher in winter relative to summer. Chaigneau et al. (2008) also reports larger inertial amplitudes and mixed layer energy during winter for the northern Pacific.

Although the southern Indian Ocean does not have a strong seasonal cycle in wind strength and direction, it has stronger near-inertial amplitudes and energy than other Southern Hemisphere basins (Chaigneau et al. 2008). This is associated with the large energy transfer from wind to near-inertial motions during austral autumn and winter (Alford 2003). We found that strong energy transfer is often associated with strong KE$_{in}$ in the mixed layer. Anticyclonic eddies were observed to influence the vertical propagation of near-inertial energy into the ocean interior in agreement with eddy-rich channel model results (Zhai et al. 2005). Considering the richness of eddies in this region, it is important to include the effect of mesoscale eddies in mixed layer models.

Most of the turbulent mixing in the ocean interior is due to internal wave breaking. NIWs are one of the most energetic types of internal waves and mostly contribute to turbulent mixing in the upper ocean (e.g., Alford and Gregg 2001; Waterman et al. 2013; Whalen et al. 2018) although they have the potential to reach the deep ocean (Silverthorne and Toole 2009; Alford et al. 2012). By comparing the location of the NIWs identified in this study with the distribution of diapycnal turbulent mixing from the same float observations using a shear-strain finescale parameterization (Cyriac et al. 2021), we observe that most of the elevated NIW amplitudes are

<table>
<thead>
<tr>
<th>Shear variance ($10^{-2}$ s$^{-1}$)</th>
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<tbody>
<tr>
<td>Pressure (dbar)</td>
</tr>
<tr>
<td>0</td>
</tr>
<tr>
<td>0.5</td>
</tr>
<tr>
<td>1</td>
</tr>
<tr>
<td>1.5</td>
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<td>2</td>
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<td>2.5</td>
</tr>
<tr>
<td>3</td>
</tr>
<tr>
<td>3.5</td>
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</table>

**Fig. 15.** Mean shear variance from float EM-6664 in anticyclonic eddy AE3 (red) and cyclonic eddy CE2 (blue).
associated with enhanced turbulent mixing along the float tracks (Fig. 16). Turbulent mixing is further elevated in anti-cyclonic eddies where NIWs are present. Thus, the interaction of eddies and NIWs could be a key factor in the mixing budget of this region, which is important for the unique circulation in the eastern south Indian Ocean.

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![Fig. 16](image_url)  
**Fig. 16.** The turbulent diffusivity along the float tracks from Cyriac et al. (2021) (color shading). The vertically propagating NIWs identified from complex demodulation below 250 m are also marked (inverted triangles). The gray contours show where the near-inertial amplitudes exceed 0.1 m s\(^{-1}\). Note that the x axis in each column represents the time period unique to each float between 10 Jul 2013 and 25 Oct 2013.


