Observed Extreme Air–Sea Heat Flux Variations during Three Tropical Cyclones in the Tropical Southeastern Indian Ocean

Xiangzhou Song, Chunlin Ning, Yongliang Duan, Huiwu Wang, Chao Li, Yang Yang, Jianjun Liu, and Weidong Yu

Abstract: Six-month buoy-based heat flux observations from the poorly sampled tropical southeastern Indian Ocean are examined to document the extremes during three tropical cyclones (TCs) from December 2018 to May 2019. The most striking feature at the mooring site (16.9°S, 115.2°E) during the TCs is the extensive suppression of the daytime net surface flux (Qnet) together with a concomitant reduction of 470 W m⁻², a peak decrease at approximately noon of 695 W m⁻² and an extreme drop during TC Riley of 800 W m⁻². The mean surface cooling in the daytime is primarily contributed by the 370 W m⁻² decrease in shortwave radiation associated with increased cloudiness. The air–sea turbulent heat flux increases by approximately 151 W m⁻² in response to the enhanced wind speed under near-neutral boundary conditions. The daily mean rainfall-induced cooling is 8 W m⁻², with a maximum magnitude of 90 W m⁻². The mean values, seasonal variation, and synoptic variability of the characteristic heat fluxes are used to assess the new reanalysis data from ERA5 and MERRA2 and the analyzed OAFlux. The overall performance of the high-frequency net heat flux estimates at the synoptic scale is satisfactory, but the four flux components exhibit different quality levels. A serious error is that ERA5 and MERRA2 poorly represent TCs, and they show significant daily mean Qnet biases with opposite directions, −59 W m⁻² (largely due to the overestimated latent heat with a bias of −76 W m⁻²) and 50 W m⁻² (largely due to the overestimated shortwave radiation with a bias of 41 W m⁻²), respectively.

KEYWORDS: Indian Ocean; Madden-Julian oscillation; Tropical cyclones; Air-sea interaction; Heat budgets/fluxes; Buoy observations

1. Introduction

The warmest sea surface temperature (SST) in the Indo-Pacific warm pool drives global atmospheric circulation through deep convection (Matsumo 1966; Bjerknes 1969; Gill 1980; Graham and Barnett 1987). Understanding how the air–sea heat fluxes vary is important for the study of tropical air–sea interactions. The net air–sea heat flux (Qnet), that is, the sum of solar shortwave radiation (SW), longwave radiation (LW), latent heat (LH), and sensible heat (SH), plays a central role in determining the seasonal evolution of SST (Moisan and Niler 1998; Yu et al. 2006). In tropical regions, LH and SW are normally the two dominant terms in Qnet (Nigam and Chao 1996; Carton and Zhou 1997; Foltz and McPhaden 2005; Foltz et al. 2010).

Tropical cyclones (TCs) are a significant weather-scale phenomenon in the tropics and are among the most powerful events in the atmospheric circulation system. TCs cause significant sea surface cooling (Leipper 1967; Withke and Johnson 1976; McPhaden et al. 2009a), which produces a feedback mechanism that moderates their intensity (Schade and Emanuel 1999; D’Asaro et al. 2007; Lin et al. 2009). A case study by Lin et al. (2009) indicates that TC Nargis in 2008 rapidly intensified from a weak category 1 storm to an intense category 4 storm within a day in the Bay of Bengal in response to a preexisting warm ocean anomaly. Thus, accurate knowledge of the air–sea exchanges of heat, moisture and momentum associated with TCs is of major relevance for predicting TC tracks and intensity model physics (Emanuel 1995; Lin et al. 2009). However, understandings of the variations in Qnet and the contributions of the heat flux components during TC passage remain poor due to the limited availability of full-flux observations.

The Madden–Julian oscillation (MJO) is the primary source of intraseasonal (30–90 days) variability in the tropics (Madden and Julian 1971, 1972; Hendon and Salby 1994; Zhang 2005; DeMott et al. 2015). It originates in the equatorial Indian Ocean, propagates eastward at a typical speed of 5 m s⁻¹ and weakens in the equatorial Pacific (Wheeler and Hendon 2004; Zhang 2005). Previous studies indicate that strong surface-level westerly wind bursts (WWBs) and heavy precipitation during active MJO phases lead to reduced insolation and enhanced evaporative LH (Krishnamurti et al. 1988; Zhang and McPhaden 1995; Jones and Weare 1996; Cronin and McPhaden 1997; Benedict and Randall 2007; Chi et al. 2014; DeMott et al. 2015). The transition from dry to rainy conditions is accompanied by increased cloudiness (Myers and Waliser 2003; Riley et al. 2009).
et al. 2011; Rowe and Houze 2015) and decreased SW (Weller and Anderson 1996; Webster et al. 1996).

TCs are widely shown to be modulated by the MJO. The genesis of TCs is generally enhanced during active MJO phases, which are associated with enhanced convective activity (Hall et al. 2001; Bessafi and Wheeler 2006; Chand and Walsh 2010). During active MJO phases, convectively coupled equatorial waves (equatorial Rossby waves, Kevin waves, and Rossby–gravity waves) play dynamic roles in influencing TC activities in terms of low-level relative vorticity, upper-level divergence, and vertical wind shear (Bessafi and Wheeler 2006). The air–sea heat flux changes during TCs contribute to the maintenance and propagation of deep MJO convection (Emanuel 1987; Neelin et al. 1987; Riley Dellaripa and Maloney 2015). Thus, the feedback associated with SW and turbulent heat fluxes between the ocean and the atmosphere provides an important energy source for driving MJO activities (Sobel et al. 2008). However, how the heat flux components of SW, LW, LH, SH, and Qnet quantitatively change during TCs under active MJO phases is poorly understood.

Despite recent projects such as the Dynamics of the MJO (DYNAMO; Yoneyama et al. 2013) and the Research Moored Array for African–Asian–Australian Monsoon Analysis and Prediction (RAMA; McPhaden et al. 2009b), the tropical southeastern Indian Ocean and the Maritime Continent region, where the MJO-related SST variability (Fig. 1) and outgoing longwave radiation (OLR) variations are most significant (Sobel et al. 2008; Viallard et al. 2013), remain undersampled, particularly in terms of in situ observations of air–sea heat fluxes. Using newly deployed buoy observations with full-flux parameters in the eastern Indian Ocean warm pool off the northwestern coast of Australia (Feng et al. 2020), the present study aims to obtain a quantitative understanding of the rarely captured extreme air–sea heat fluxes associated with three TCs in order to further understand the oceanic responses and feedbacks associated with cyclone and MJO development (Emanuel 1987; Neelin et al. 1987; Riley Dellaripa and Maloney 2015).

However, estimates of the air–sea heat fluxes suffer from various uncertainties, which arise from both the observational air–sea variables and empirical parameters (Weare 1989; Gleckler and Weare 1997; Grist and Josey 2003; Brunke et al. 2011; Yu et al. 2013; Yu 2019). Heat flux errors can be identified through direct intercomparisons between point-to-point observations and flux products (Gleckler and Weare 1997; Brunke et al. 2011; Song 2020). For example, reanalysis and analyzed heat flux products have been compared with in situ measurements in the Atlantic Ocean (Josey 2001; Renfrew et al. 2002; Sun et al. 2003; Yu et al. 2004). Using buoy observations in the subduction region of the North Atlantic, Josey (2001) assessed the heat fluxes of the first general atmospheric reanalysis and found that the reanalysis biases are primarily due to a combination of underestimated SW and overestimated LH. In addition to there being uncertainties in heat flux products, significant discrepancies in LH have been found in CMIP5 models (e.g., Zhang et al. 2018). Based on the oceanic heat budget balances in a specified control region, such as an enclosed temperature volume of the western Pacific warm pool (Song and Yu 2013) or a semienclosed basin of the Mediterranean Sea (Song and Yu 2017), the Qnet climatology can be inversely estimated and used to assess the heat flux climatologies in different heat flux models and products. These studies provide physical constraints on understanding heat flux biases in current models. However, studies of extreme heat fluxes...
Table 1. Information on the observed variables and associated instruments on the buoy. Negative values of the observational heights in the last column indicate the CTD sensors beneath the sea surface, while the positive numbers represent the heights of meteorological sensors above the surface.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Manufacturer and sensor</th>
<th>Resolution</th>
<th>Sampling configuration</th>
<th>Accuracy</th>
<th>Nominal depth or height (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sea surface temperature</td>
<td>SeaBird, IM-37 CT</td>
<td>0.001°C</td>
<td>30 min</td>
<td>0.003°C</td>
<td>−1</td>
</tr>
<tr>
<td>Sea surface salinity</td>
<td></td>
<td>0.0001 S m⁻¹</td>
<td>30 min</td>
<td>0.02 psu</td>
<td>−1</td>
</tr>
<tr>
<td>Surface air temperature</td>
<td>Vaisala, HMP155</td>
<td>0.003°C</td>
<td>10 min</td>
<td>0.2°C</td>
<td>+3</td>
</tr>
<tr>
<td>Relative humidity</td>
<td></td>
<td>0.003% RH</td>
<td>10 min</td>
<td>1% RH</td>
<td>+3</td>
</tr>
<tr>
<td>Sea level pressure</td>
<td>Vaisala, PTB110</td>
<td>0.02 hPa</td>
<td>10 min</td>
<td>0.08%</td>
<td>+3</td>
</tr>
<tr>
<td>Wind speed</td>
<td>Gill Inst. Ltd., WindSonic</td>
<td>0.003 m s⁻¹</td>
<td>10 min</td>
<td>2%</td>
<td>+4</td>
</tr>
<tr>
<td>Wind direction</td>
<td></td>
<td>0.02°</td>
<td>10 min</td>
<td>3°</td>
<td>+4</td>
</tr>
<tr>
<td>SW</td>
<td>Eppley Laboratory, SPP</td>
<td>0.03 W m⁻²</td>
<td>10 min</td>
<td>1%</td>
<td>+3.5</td>
</tr>
<tr>
<td>LW</td>
<td>Eppley Laboratory, PIR</td>
<td>0.08 W m⁻²</td>
<td>10 min</td>
<td>1%</td>
<td>+3.5</td>
</tr>
<tr>
<td>Precipitation</td>
<td>R. M. Young, 50203</td>
<td>0.02 mm h⁻¹</td>
<td>10 min</td>
<td>6 mm h⁻¹</td>
<td>+3.5</td>
</tr>
</tbody>
</table>

fluxes at the weather scale during TCs can provide another important benchmark for assessments of heat flux products and models, especially in the southeastern Indian Ocean.

This paper has two purposes. The first is to thoroughly document the observed extreme air–sea heat flux variations in the tropical southeastern Indian Ocean. The second is to assess two newly released atmospheric reanalysis datasets and one analyzed air–sea flux product against the observations. The remainder of this paper is organized as follows: section 2 introduces the data and the air–sea flux algorithms; the buoy-based heat flux variations associated with TCs during active MJO phases are analyzed in section 3; in section 4, the heat flux comparisons among the buoy observations, reanalysis datasets and analyzed air–sea flux product are shown; finally, a summary and discussion are given in section 5.

2. Data and method

a. Data description

1) BUOY OBSERVATIONS

The Bailong Buoy (Freitag et al. 2016) was deployed in the tropical southeastern Indian Ocean at 16.9°S, 115.2°E (Fig. 1 and Feng et al. 2020) in November 2018 through a collaboration between the First Institute of Oceanography (FIO), Ministry of Natural Resources (MNR), China, and CSIRO, Australia. The local water depth at the buoy location is 1734 m, and the buoy is situated approximately 470 km off the northwestern coast of Australia. This flux buoy station is recommended by the CLIVAR Global Ocean Observing System (GOOS) GOOS decadal review on Indian Ocean Observing System (IndOOS), and its scientific merit is highlighted in Beal et al. (2020). The present pilot study is inspiring discussion of possible long-term deployment through international cooperation under the coordination of IndOOS Resource Forum, particularly among Australia, China, the United States, and other countries. Unfortunately, the anchor chain of the buoy broke under severe weather conditions in May 2020. The future plan is being discussed among the potential international partners under the coordination of IndOOS Resource Forum.

This buoy provided high-resolution observations of full-set air–sea flux variables (Table 1), including surface wind speed (WS), wind direction (WD), surface air temperature (SAT), relative humidity (RH), sea level pressure (SLP), and SST. The meteorological sensors are mounted on the top of the buoy with specific nominal heights (see Table 1 and Fig. 2). The SST and sea surface salinity (SSS) are observed with a conductivity and temperature (CT) recorder (SeaBird, IM-37 CT) at the depth of 1 m. The wind sensor was mounted approximately 1 m above the other instruments, thereby avoiding the effects of the other objects on the accuracy of wind speed and direction observations. To avoid the effects of other instruments on the radiation observations, the radiometers and other sensors were placed on opposite sides of the buoy. Nevertheless, there remain inevitable uncertainties in the radiation measurements due to the objects mounted above the radiometers (Fig. 2). In this paper, these uncertainties are assumed to be unsubstantial; however, in future studies and observations, attention should be devoted to improving the mounting scheme.

The temporal resolution of the recorded atmospheric parameters is 10 min, whereas the temporal resolution of SST is 30 min. It should be noted that the 10-min recorded meteorological values and 30-min recorded SST are averages of higher-frequency samples. For meteorological variables, the values are sampled at a high frequency of 2 Hz during the starting 2 min. The 240 samples in 2 min are averaged to obtain the 10-min mean value, which is recorded. However, SST is sampled every 10 min. The 30-min recorded SST and SSS are obtained by averaging the three values collected over a half-hour period. In this paper, the quality of the high-resolution raw data is controlled by three basic steps: first, the values of air–sea variables beyond the reasonable range [e.g., 0–40 m s⁻¹ for wind speed and 0–40°C for temperature, similar to the procedure in Wang et al. (2017)] are removed; second, the missing values are linearly interpolated based on the nearest samples; third, the hourly variables are then obtained by averaging the high-resolution measurements. Fortunately, no physical variable is out of the physically reasonable range, and there is only one missing sample over the whole observational period; a replacement value was obtained by the second step.
Six months of data (December 2018–May 2019) are used here to investigate the extreme variations in air–sea heat fluxes at the weather scale associated with TCs in the southeastern Indian Ocean. The precipitation data from the Tropical Rainfall Measuring Mission (TRMM) were incorporated to help identify cyclonic activities. The TC activities and trajectories in the six-month observation dataset are incorporated based on the International Best Track Archive for Climate Stewardship (IBTrACS; Knapp et al. 2010) from the National Oceanic and Atmospheric Administration (NOAA).

2) OBJECTIVELY ANALYZED AIR–SEA FLUX PROJECT

The Objectively Analyzed Air–Sea Flux (OAFlux) project for the global oceans (Yu and Weller 2007) provides a multi-decadal analysis of air–sea heat and momentum fluxes for research on the global energy budget and climate change. Synthesized observations/estimates from various sources, including satellite observations and reanalysis, were used to reduce the errors, and the best estimate was produced with a minimum error variance. The Coupled Ocean–Atmospheric Response Experiment (COARE) bulk flux algorithm, version 3.0 (Fairall et al. 1996, 2003), is used. OAFlux outputs daily LH and SH and the associated parameters of SST, SAT, specific humidity, and WS and has been widely regarded as an important benchmark for air–sea flux intercomparisons and the global surface heat budget (e.g., Cronin et al. 2019; Yu 2019).

3) NEWLY RELEASED ATMOSPHERIC REANALYSIS DATASETS

Two reanalysis datasets are used here. One is the Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA2; Gelaro et al. 2017), which provides data from 1980 to the present. This new global reanalysis replaces and extends the original MERRA dataset (Rienecker et al. 2011). The other is the European Centre for Medium-Range Weather Forecasts (ECMWF) Reanalysis version 5 (ERA5), which provides hourly estimates of a large number of atmospheric, land and oceanic variables (Hersbach et al. 2019). ERA5 replaces ERA-Interim (Dee et al. 2011). Both ERA5 and MERRA2 have combined vast historical observations and state-of-the-art models to produce a global reanalysis dataset using advanced data assimilation systems. Different from the use of the COARE algorithm in OAFlux, the turbulent parameterization of the vertical fluxes of LH, SH, and momentum must be stable, give smooth results and interact harmoniously with other modulations of the operational modeling system (Louis et al. 1981). Thus, ERA5 and MERRA2 employed the Louis scheme (Louis 1979, and subsequent studies since the 1970s) in the ECMWF system and the Goddard Earth Observing System-5 (GEOS-5), respectively. However, MERRA2 has replaced the Louis scheme with the parameterization of Helfand and Schubert (1995), which modifies the turbulent exchange coefficients under different boundary conditions (Molod et al. 2015).

OAFlux and MERRA2 incorporate the high-resolution daily 1/4° optimum interpolation SST (OISST) products from NOAA (Reynolds et al. 2007) as the boundary conditions, whereas ERA5 assimilates the Met Office Operational SST and Sea Ice Analysis (OSTIA) product. OAFlux produces daily variables without diurnal variations, whereas MERRA2 and ERA5 output the diurnal (hourly) heat fluxes with modeled atmospheric variables but a constant daily SST. Thus, there is no diurnal cycle in OAFlux. Diurnal variations in heat fluxes in MERRA2 and ERA5 are mainly determined by atmospheric dynamics instead of ocean processes. The spatial resolutions of OAFlux, MERRA2, and ERA5 are 1° × 1°, 0.625° (longitude) × 0.5° (latitude), and 0.25° × 0.25°.
respectively. In this paper, the grids in these three products that are nearest to the buoy location are used for comparison rather than the four-grid mean results. As this study focuses on the effects of extreme weather systems (TCs) on heat flux variations, the four-grid mean results were not considered because they may cover up extreme values. The buoy observations have not yet been assimilated in the reanalysis, and thus, they can be regarded as an independent benchmark for the reanalysis comparisons. In this paper, the hourly heat fluxes based on buoy observations are directly compared with the reanalysis; however, the daily mean results are constructed from hourly data with statistical errors to provide a basis for the comparison with OAFlux in section 4. The location of the buoy and grids of both OAFlux and the reanalysis are shown in Fig. 1.

b. Method of air–sea flux calculation

$Q_{\text{net}}$ consists of four components as follows:

$$Q_{\text{net}} = Q_{\text{SW}} + Q_{\text{LW}} + Q_{\text{LH}} + Q_{\text{SH}},$$  

(1)

where $Q_{\text{SW}}$ represents the net downward SW, $Q_{\text{LW}}$ represents the net LW, $Q_{\text{LH}}$ represents the LH, and $Q_{\text{SH}}$ represents the SH. Positive values denote that the ocean is receiving heat.

Constant surface albedo ($\alpha = 0.06$) and nonconstant surface albedo from MERRA2 were used to estimate the net SW ($Q_{\text{SW}} \downarrow$) for comparison:

$$Q_{\text{SW}} \downarrow = Q_{\text{SW}} \downarrow (1 - \alpha),$$  

(2)

where $Q_{\text{SW}} \downarrow$ denotes the incoming solar radiation directly measured from the buoy. Although the surface albedo $\alpha$ in MERRA2 is not constant but changes with the surface state, the mean value is 0.06, which is the same as the fixed surface albedo in ERA5 and the conventional choice for buoy observations (e.g., Feng et al. 2020). The nonconstant surface albedo in MERRA2 is slightly lower from 0200 to 0800 UTC when the downward SW is stronger with warmer SST associated with diurnal variations and is slightly higher when the SW is weaker (Figs. 3a–c). The six-month mean downward SW in MERRA2 is slightly lower from 0200 to 0800 UTC when the downward SW is stronger with warmer SST associated with diurnal variations and is slightly higher when the SW is weaker (Figs. 3a–c). The six-month mean downward SW in MERRA2 is slightly lower from 0200 to 0800 UTC when the downward SW is stronger with warmer SST associated with diurnal variations and is slightly higher when the SW is weaker (Figs. 3a–c). The six-month mean downward SW in MERRA2 is slightly lower from 0200 to 0800 UTC when the downward SW is stronger with warmer SST associated with diurnal variations and is slightly higher when the SW is weaker (Figs. 3a–c).
found for buoy observations, MERRA2 and ERA5. However, a
six-month mean difference in SW of only 1 W m$^{-2}$ is obtained by
using the MERRA2 scheme for all products, as the sea surface is
less reflective with higher downward SW.

Following Dickey et al. (1994), the net LW ($Q_{\text{LW}} \uparrow \downarrow$) is
calculated from the upward LW ($Q_{\text{LW}} \uparrow$) associated with the
surface thermal state and the observed downward LW ($Q_{\text{LW}} \downarrow$) as

$$ Q_{\text{LW}} \uparrow \downarrow = Q_{\text{LW}} \downarrow - Q_{\text{LW}} \uparrow, $$

where $Q_{\text{LW}} \downarrow$ was observed from the buoy. The upwelling LW
leaving the sea surface includes two parts [Eq. (4)]: one part is
the emissions associated with SST, and the other part is the
reflection of downwelling LW:

$$ Q_{\text{LW}} \uparrow = e_s \sigma (SST)^4 + (1 - e_s) Q_{\text{LW}} \downarrow, $$

where $\sigma = 5.7 \times 10^{-8}$ is the Stefan–Boltzmann constant and $e_s = 0.985$ is the surface emissivity, which is dependent on
temperature and wavelength. In this paper, the surface net LW
is defined as the difference between the downwelling longwave
radiation (DLW) associated with atmospheric absorption, emission and scattering within the entire atmospheric column
and the upwelling longwave radiation (ULW) emitted and reflected by the ocean surface [Eq. (4)].

Following the Monin–Obukhov similarity theory (Monin
and Obukhov 1954), turbulent heat fluxes, namely, LH and
SH, are conventionally estimated by bulk formulas (Liu et al.
1998). Radiative fluxes include SW and LW, both of which are
closely associated with the cloud fraction in the climate sys-
tem, clouds absorb and reflect solar radiation (Cess et al. 1995;
Ramanathan et al. 1995; Pilewskie and Valero 1995) and
radiate LW back into space (Chen et al. 2000), which reduces the
amount of energy penetrating to the surface. The water vapor in
clouds effectively absorbs and emits (reemits) LW as a
blackbody (e.g., Kiehl and Trenberth 1997). Based on Eq. (4),
the extreme heat fluxes induced by the TCs in the reanalysis can be
compared with the buoy-observed fluxes. The dates of the three
TCs passing the buoy site in 2019 are 26 January, 2 March, and 8 April. The calm weather conditions were
constructed by averaging the states two days before and after
the TC passages [23 and 29 January for TC Riley (TC 1), 20 and
25 March for TC Veronica (TC 2), and 4 and 11 April for TC
Wallace (TC 3)] to analyze the extreme air–sea heat flux changes associated with these TCs.

### 3. Diagnosis of cyclone-induced extreme air–sea heat fluxes

#### a. Identification of a cyclone

TCs and the MJO are characterized by heavy precipitation
and low SLP. In 2019, the buoy observations reveal three heavy
precipitation events on 25–27 January, 21–22 March, and
8 April (Fig. 4). The SLP was also observed to be less than
1005 hPa. The precipitation intensity exceeded 50 mm h$^{-1}$ on
26 January 2019. Although TRMM, ERA5, and MERRA2 all
captured this extreme precipitation, their values are signifi-
cantly lower, especially in the two reanalysis datasets. These
differences may be partly attributed to the different resolutions
between the observations and the satellite/reanalysis products.
However, discussion of this difference is beyond the scope of
this paper.

The passage of a TC is identified by the combination of precipitation and SLP. Three TCs occurred between
December 2018 and May 2019 off the northwestern coast of
Australia: TC Riley (21 January–2 February 2019), severe TC
Veronica (18–28 March 2019), and TC Wallace (3–10 April
2019). The trajectories of these TCs based on IBTrACS have
been included in Fig. 1. Figure 5 shows the daily mean mini-
imum central SLP and trajectories based on the reanalysis data
(MERRA2 and ERA5) and IBTrACS during TC Riley
(26 January 2019), severe TC Veronica (22 March 2019), and
TC Wallace (8 April 2019). Significant low pressure centers can
be found in association with cyclones. The trajectories associ-
ated with the minimum central SLP in MERRA2 and ERA5
show good agreement with the IBTrACS data. Therefore, the
extreme heat fluxes induced by the TCs in the reanalysis can be
compared with the buoy-observed fluxes. The dates of the three
TCs passing the buoy site in 2019 are 26 January, 2 March, and 8 April. The calm weather conditions were
constructed by averaging the states two days before and after
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25 March for TC Veronica (TC 2), and 4 and 11 April for TC
Wallace (TC 3)] to analyze the extreme air–sea heat flux
changes associated with these TCs.

#### b. Radiative fluxes

Radiative fluxes include SW and LW, both of which are

In addition to the above four heat flux components at the
air–sea interface, the rainfall-induced SH flux (RF) during the
TC passage is estimated based on the following algorithm
(Gosnell et al. 1995; Fairall et al. 1996):

$$ Q_{\text{RF}} = \rho L_e c_{eL} |\vec{U}_z| (\Delta q), $$

$$ Q_{\text{SH}} = \rho c_p c_{sh} |\vec{U}_z| (\Delta T), $$

where $\rho$ is the density of air, $L_e$ is the LH of evaporation, $c_p$ is
the specific heat capacity of air, and $U_z$ is the wind vector at a
height $z$. The turbulent exchange coefficients for the LH and
SH are denoted by $c_{eL}$ and $c_{sh}$, respectively. The terms $\Delta q$ and $\Delta T$ represent the sea–air humidity and temperature difference,
respectively. In this paper, the COARE bulk flux algorithm
version 3.0 (Fairall et al. 1996, 2003; Edson et al. 2013) was used
as in the OAFlux project (Yu and Weller 2007).

In the three TC passages [23 and 29 January for TC Riley
(TC 1), 20 and 25 March for TC Veronica (TC 2), and 4 and 11 April for TC
Wallace (TC 3)] to analyze the extreme air–sea heat flux changes associated with these TCs.

![Image](https://example.com/figure7.png)
resolution of SLP, maximum WS, hourly precipitation and 10-min downward SW during the three TC passages. The observed minimum SLP and maximum WS (shaded in Fig. 7) can further confirm the passage of TCs at the buoy location. Significant rainfall is found by buoy observations, TRMM and reanalysis, and the differences among them are within the range of acceptable uncertainties. Further individual TC cases associated with SW and LW are given in Figs. 7 and 8, respectively, to illustrate the diversity. These figures reveal apparent case-to-case differences in the diurnal cycle of SW and LW variations between TC conditions and calm weather conditions. TC Riley demonstrates the most striking contrast between TC conditions and calm weather conditions, with a peak SW (LW intensity) drop of 916 (75) W m\(^{-2}\). Severe TC Veronica follows but with weaker changes, with a peak SW (LW intensity) drop of 777 (48) W m\(^{-2}\). The behavior of TC Wallace looks different. Its SW drop was much weaker than those in TCs Riley and Veronica, while the LW intensity decrease was stronger, with a value of approximately 64 W m\(^{-2}\) from 1400 to 1600 UTC 8 April 2019. The mean heat fluxes between each of the three TCs and calm weather during the day and night are summarized in Table 2. The statistics of the values are calculated separately for daytime (0600–2000 UTC) and nighttime (2000–0600 UTC) to distinguish any clear diurnal differences. The daytime mean reduction (relative to the calm days) of SW averaged over the three TCs is 370 W m\(^{-2}\). No significant reduction is observed at night. The mean increases in LW in the daytime and nighttime are 47 and 39 W m\(^{-2}\), respectively, resulting in amplification of the LW diurnal cycle. The maximum reduction (increase) in the magnitude of SW (LW) is approximately 916 (90) W m\(^{-2}\).

The above results indicate that SW plays a central role in TC-induced surface cooling during the passage of a TC. Why does the TC-affected net SW show significant case-to-case differences at the time of the TC passage? The net SW during the passage of TC Riley is greatly suppressed, with a maximum value of 916 W m\(^{-2}\), whereas that on 8 April 2019 during TC Wallace is only slightly reduced. Cloud variations may account for these differences. Although an increased cloud fraction has been identified during TC passage (e.g., Houze 2014), low cloud fraction conditions can be found in the eye/eyewall region with low-topped stratus and/or stratocumulus clouds, which might help the downward penetration of SW to the sea surface. However, the eyes of the three TCs did not pass over
the buoy based on the SLP patterns and trajectories in Fig. 5. From the observed heavy rainfall (Fig. 7) and the distance between the minimum SLP grid and the buoy site (greater than 50 km, Fig. 5), the buoy is inferred to have been located in the rainbands beyond the eye/eyewall regions. The strength of the SW suppression during the three TC passages should depend on the timing of the TC passage by the buoy. TC Riley and TC Veronica passed the buoy during the daytime when the SW was strong, and thus, the net SW arriving at the sea surface could have been suppressed by the increased cloudiness. When TC Wallace passed by the buoy during the night on 8 April, the SW in the daytime was not suppressed significantly. However, the SW on 9 April was reduced by a mean magnitude of 500 W m$^{-2}$ as a result of the extended effects of TC Wallace.

c. Turbulent heat fluxes

The atmospheric boundary conditions (Fig. 6) are examined before the air–sea turbulent heat fluxes are addressed. The stability is estimated by the Monin–Obukhov stability parameter ($z/L$), where $z$ is the height of the turbulent transfer coefficient, and $L$ is the Obukhov length scale estimated based on the COARE 3.0 algorithm (see Song 2020). Unstable, near-neutral, and stable conditions are determined by $z/L < -0.4$, $-0.4 < z/L < 0.1$, and $z/L > 0.1$, respectively. The stability during the observational period was dominated by near-neutral and unstable conditions, although a few stable conditions were observed when the SAT was occasionally higher than the SST. Under near-neutral and unstable boundary conditions, turbulent heat fluxes transferred from the ocean to the atmosphere (negative magnitude, Figs. 8 and 9). During the three TCs, near-neutral boundary conditions were dominant.

During the TC period, the LH magnitude significantly increased (Table 2) as the diurnal cycle intensified (Fig. 9). The mean daytime (nighttime) increase in the LH magnitude averaged over the three TCs is 118 (118) W m$^{-2}$, as shown in Table 2, which is largely explained by the WS increase of 8 (10) m s$^{-1}$. The RH increases during the passage of a TC. Although the air–sea humidity difference in terms of the near-surface specific humidity, which is dependent on SST and the specific humidity, does not show significant enhancement (Figs. 9d–f), the increased WS during the TC passages can account for the enhanced LH under the near-neutral boundary conditions.

Even though the SH is low and usually plays a minor role in the surface net heat flux, the three TCs studied here clearly show an increase in SH (Fig. 8), with a mean value of 27 (39) W m$^{-2}$ in the daytime (nighttime) (Table 2). Such intensified SH is mainly due to the increased WS and enlarged air–sea temperature difference. Both SAT and SST decrease during the TC passages, and SAT decreases more than SST (Figs. 8a–c). The average increase in LH (118 W m$^{-2}$) in the daytime is equivalent to that (118 W m$^{-2}$) in the nighttime, but the actual value varies greatly among TC events (see the following analysis). SH increases by 27 W m$^{-2}$ in the daytime and by an equivalent magnitude of 39 W m$^{-2}$ in the nighttime; these values are also dependent on the different TC features.

The hourly LH (SH) anomaly during the TCs, with wind anomaly and air–sea humidity (temperature) differences under near-neutral boundary conditions, is shown in Fig. 10. It is evident that the LH and SH increase with increases in the wind anomaly, which is consistent with previous studies (Krishnamurti et al. 1988; Weller and Anderson 1996; Hendon and Glick 1997; Inness and Slingo 2003; Sobel et al. 2008; Araligidad and Maloney 2008; Chi et al. 2014). When the atmospheric boundary is under near-neutral conditions, the wind effect associated with the Reynolds stress exerts more control over LH and SH than the thermal effect. With the wind anomaly changing from 3 to 15 m s$^{-1}$, the LH and SH increase by 270 and 75 W m$^{-2}$, respectively. The $\Delta Q$ and $\Delta T$ anomalies do not show significant increases with increases in the wind anomaly or LH + SH. The scatterplots (Fig. 10) indicate that $\Delta Q$ and $\Delta T$ are sometimes weakened during the TC passages; however, enhanced LH and SH are also found as a result of a stronger wind anomaly. Therefore, in light of the mean Reynolds stress, any changes in the wind anomaly will help overcome the stability related to buoyancy, which determines the variations in LH and SH.
In addition to the conductive air–sea SH, the RF based on Eq. (7) is related to precipitation and the air–sea temperature difference. Previous studies have indicated that intense rainfall in the tropical ocean can produce surface cooling with a maximum magnitude of approximately 200 W m$^{-2}$ (e.g., Gosnell et al. 1995; Cronin and McPhaden 1997). However, on average, the RF is less than 5 W m$^{-2}$ in the tropical oceans. In this study, with the heavy precipitation during TCs, the RF may be an

FIG. 6. As in Fig. 4, but for time series of (a) Qnet (W m$^{-2}$), (b) SW (W m$^{-2}$), (c) LW (W m$^{-2}$), (d) LH (W m$^{-2}$), (e) SH (W m$^{-2}$), (f) boundary layer stability ($z/L$), (g) SST ($^\circ$C), (h) SAT ($^\circ$C), (i) RH (%), and (j) WS (m s$^{-1}$). The calm weather conditions and TC statuses are indicated by the blue and red bars in (a) and (f), respectively. The stable atmospheric stability parameter ($z/L > 0.1$) is marked by black circles in (f), with near-neutral conditions ($-0.4 < z/L < 0.1$) in magenta and unstable conditions ($z/L < -0.4$) in green.

In addition to the conductive air–sea SH, the RF based on Eq. (7) is related to precipitation and the air–sea temperature difference. Previous studies have indicated that intense rainfall in the tropical ocean can produce surface cooling with a maximum magnitude of approximately 200 W m$^{-2}$ (e.g., Gosnell et al. 1995; Cronin and McPhaden 1997). However, on average, the RF is less than 5 W m$^{-2}$ in the tropical oceans. In this study, with the heavy precipitation during TCs, the RF may be an

FIG. 7. (left) The observations of (a) SLP (hPa), (d) maximum wind speed (m s$^{-1}$), (g) precipitation (mm h$^{-1}$), and (j) downward SW during TC Riley (TC 1). Note that the SLP, maximum WS, and downward SW are based on the 10-min observations, while the observed precipitation data are averaged hourly for a comparison with different products of ERA5 (magenta), MERRA2 (orange), and TRMM (green). The situations during (center) TC Veronica (TC 2) and (right) TC Wallace (TC 3). The blue shaded patches indicate the calm weather before and after the TC passages, while the dark yellow patches indicate the TC passing by the buoy location, with significantly minimum SLP, enhanced WS, and precipitation and suppressed downward SW. The calm weather conditions are obtained by averaging the hourly observations of variables on 23 and 29 Jan for TC Riley (TC 1), 20 and 25 Mar for TC Veronica (TC 2), and 4 and 11 Apr for TC Wallace (TC 3), and the active TC state is calculated on 26 Jan, 22 Mar, and 8 Apr 2019 for TCs 1, 2, and 3, respectively.
important driver of sea surface cooling. Using the high-resolution full-flux buoy observations, the RF was estimated based on the Gosnell et al. (1995) algorithm to identify the extreme RF values associated with the passage of the TCs. Figure 11 shows the RF during the three TC passages and the corresponding ratios between the RF and the conductive SH. The maximum cooling can be found in TC Riley, with an approximate value of 90 W m$^{-2}$, which is twice the conductive SH flux associated with the air–sea temperature difference. The maximum RF during TC Wallace is 73 W m$^{-2}$, which is greater than the SH, with a maximum ratio of approximately 1.5. However, the RF during TC Veronica is low due to the low rainfall rate (Figs. 4 and 7). On average, the daily mean RF during the three TCs is 28 W m$^{-2}$ (Table 2), which indicates that the RF is an essential component of surface cooling and the heat budget balance.

d. Net surface heat flux

Previous studies have focused on the diurnal variations in precipitation and cloud populations (e.g., Yang and Slingo 2001; Sakaeda et al. 2017), while the diurnal variations in heat fluxes and the extreme changes during TCs have been little explored due to the lack of data. The composite heat flux diurnal cycles during TCs and calm weather conditions are shown in Fig. 12 to emphasize the common features. SW exhibits the most striking changes during the TC period. Its diurnal cycle was dramatically suppressed during the daytime, and the peak value averaged over the noon period (1200–1400 LT) decreased from a calm weather value of 870 W m$^{-2}$ to a TC value of 75 W m$^{-2}$ during TC Riley and TC Veronica and 672 W m$^{-2}$ during TC Wallace. The significant case-to-case differences in SW suppression during the TC passages (Table 2) depend on the diurnal cycles of SW associated with the time of TC passage. LW also exhibits significant changes during the TC period. The magnitude of net LW leaving from the sea surface during the TC period significantly decreases, and its diurnal cycle is also suppressed with a more profound reduction in daytime. The LW peak value averaged over the early afternoon (1400–1600 LT) decreases (warming effect) by 53 W m$^{-2}$ from the calm weather value of $-81$ W m$^{-2}$ to the TC value of $-28$ W m$^{-2}$.

### Table 2. Summary of heat flux differences (W m$^{-2}$) between calm weather conditions and TCs during active MJO phases.

<table>
<thead>
<tr>
<th>Variables</th>
<th>TC Riley</th>
<th>TC Veronica</th>
<th>TC Wallace</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Sun</td>
<td>Dark</td>
<td>Sun</td>
<td>Dark</td>
</tr>
<tr>
<td>$\Delta Q_{\text{net}}$</td>
<td>$-565$</td>
<td>$-62$</td>
<td>$-542$</td>
<td>$-112$</td>
</tr>
<tr>
<td>$\Delta S W$</td>
<td>$-484$</td>
<td>$1$</td>
<td>$-404$</td>
<td>$2$</td>
</tr>
<tr>
<td>$\Delta L W$</td>
<td>$54$</td>
<td>$38$</td>
<td>$37$</td>
<td>$25$</td>
</tr>
<tr>
<td>$\Delta R F$</td>
<td>$-2$</td>
<td>$-25$</td>
<td>$-3$</td>
<td>$-1$</td>
</tr>
</tbody>
</table>
LH increases by 150 W m$^{-2}$ after sunrise until midnight. However, the SH increases from the early morning to midnight. Both LH and SH increase with increases in the WS during the TC passages, as shown in Fig. 10. The RF, which is related to the rainfall rate, also shows significant diurnal variations and case-to-case differences during the three TCs, which depend on the time of the heavy rain (Fig. 11 and Table 2). The sum of SW, LW, LH, SH, and RF leads to unexpectedly compressed net surface heat flux during TCs, with a mean Qnet reduction of 470 (131) W m$^{-2}$ during the day (night) (Table 2). The most significant mean decrease occurs at approximately noon (1200–1400 LT), with Qnet dropping 695 W m$^{-2}$

FIG. 9. As in Figs. 7 and 8, but for (a)–(c) RH (%), (d)–(f) specific air humidity at a height of 3.5 m (blue) and near-surface specific humidity at saturation (red) (g kg$^{-1}$) in terms of SST with a reduction of 2% associated with salinity effect, and (g)–(i) LH (W m$^{-2}$).

FIG. 10. The variations in LH and SH differences between calm weather conditions and active TCs with the observed differences in air–sea gradients $\Delta Q$ and $\Delta T$ (y axis; g kg$^{-1}$ and $^\circ$C, respectively) and WS (x axis; m s$^{-1}$) under near-neutral ($-0.4 < z/L < 0.1$) atmospheric boundary layer conditions. Colors show the magnitudes of LH and SH anomalies. The calm weather conditions during TCs in this paper are estimated as the average of the hourly air–sea variables one day before and after the calm states shown by the blue shading in Figs. 7–9.
In individual cases, Qnet decreases as much as approximately 800 W m\(^{-2}\) during TC Riley. The Qnet experiences significant diurnal variations: in the daytime, the surface cooling is dominated by the reduction in SW, while in the nighttime, the surface cooling is mainly determined by the increased LH, followed by SH. The reduced LW contributes to surface warming, and the magnitude is counterbalanced by the SW, LH, and SH.

In turn, the SST cools (Figs. 6 and 8) in response to such Qnet changes following the mixed layer heat budget diagnosis (Price et al. 1986). The SST, the mixed layer (ML) temperature, and the upper-layer heat budget, namely, the daily mean SST tendency and surface heat forcing, can be estimated by a simple dynamic relation as in Cronin and McPhaden (1997):

\[ \frac{dT_S}{dt} = \frac{Q_{NET}}{c_{pw} \rho_w h_{ML}} + \text{residual}, \]  

where \( T_S \) represents the SST and \( t \) is the time interval of one day in term A. The SST tendency is obtained by the SST difference between 2300 and 0000 UTC on a calendar day; \( c_{pw} \) is the heat capacity of the seawater, \( \rho_w \) is the seawater density, and a constant ML depth \( h_{ML} = 15 \) m is used for the estimation of term B. The daily mean Qnet is constructed from hourly buoy observations, which can drive the SST changes in a day together with the residual term. The residual term includes the ocean dynamics associated with advection (geostrophic and ageostrophic currents), entrainment and mixing processes. Figure 13 indicates that the relationship between the daily SST tendency (red) and surface Qnet forcing (blue) shows good consistency during the entire observational period from December 2018 to May 2019. The most significant dynamic structure is the surface cooling and Qnet reduction during the passage of the three TCs, similar to previous findings (Leipper 1967; Withee and Johnson 1976; McPhaden et al. 2009a).

In this paper, it is found that SW suppression (\(-184\) W m\(^{-2}\)) in response to enhanced cloudiness and increased LH (\(-118\) W m\(^{-2}\)) associated with increases in the wind anomaly under near-neutral boundary conditions dominate the Qnet reduction (\(-301\) W m\(^{-2}\)) during the passage of TCs, which determines the SST cooling tendency. These cooling processes, in particular the enthalpy fluxes, are important components of feedback that influences the TC intensity, as shown in previous studies (Schade and Emanuel 1999; D’Asaro et al. 2007; Lin et al. 2009). However, an in-depth upper-layer heat budget analysis based on the mooring conductivity–temperature and current observations will be performed following the surface heat flux analysis in future works.

### 4. Assessment of OAFlux and reanalysis data against buoy observations

Errors in heat fluxes are obstacles to our accurate understanding of upper ocean dynamics (Song and Yu 2013, 2017; Yu et al. 2013; Yu 2019). A point-to-point comparison between the observations and analyzed products is a powerful tool to identify the potential heat flux errors in OAFlux/reanalysis (Gleckler and Weare 1997; Brunke et al. 2011). In this section, the comparison among the heat fluxes from the buoy observations and OAFlux, MERRA2 and ERA5 datasets is presented, including the six-month mean state (section 4a), extreme flux changes in association with TCs (section 4b), and seasonal (section 4c) and synoptic-scale variations (section 4d).

#### a. Heat flux comparisons of six-month mean result

Figure 14 shows the six-month mean and heat flux differences in Qnet, SW, LW, LH, and SH between the TC passages and calm weather based on buoy observations, ERAS, MERRA2, and OAFlux. As no radiation flux was obtained in OAFlux, only the LH and SH were compared among products.
and $-7 \, \text{W m}^{-2}$ for buoy estimates, $62 (264, -68, -127, \text{and } -6) \, \text{W m}^{-2}$ for ERA5, and $46 (254, -70, -133, \text{and } -6) \, \text{W m}^{-2}$ for MERRA2. The LH and SH from OAFlux are $-114$ and $-3 \, \text{W m}^{-2}$, respectively (Table 3). ERA5 with higher SW, LH, and LW values produces slightly lower Qnet values than the buoy observations. The mean LH in MERRA2 is highest, which leads to a Qnet value that is $23 \, \text{W m}^{-2}$ lower than that of the buoy observations. The turbulent heat fluxes of LH and SH in OAFlux are close to those of the buoy observations, which is possibly due to the use of the same algorithm of COARE 3.0.

b. Comparison of the extreme flux changes associated with TCs

How well the reanalysis data and OAFlux capture the extreme flux variations during TCs is of great scientific and application value. Here, the buoy-based extreme flux measurements during TCs in the poorly sampled tropical southeastern Indian

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**Fig. 12.** The ensemble mean diurnal variations in (a) Qnet, (b) SW, (c) LW, (d) LH, and (e) SH during calm weather conditions (red) and TCs (blue/green) as defined in Figs. 7–9. The UTC time was translated into local time for plotting convenience of SW. (f)–(j) The heat flux differences (black) between calm weather (red) and TCs (blue). Note that the mean Qnet and SW in (a) and (b) during active TC 1 and 2 are indicated by blue (differences shown by solid in the right panels), while those during active TC 3 are marked by green (differences shown by dashed lines in the right panels, accordingly).
Ocean provide valuable validation information for the ERA5 and MERRA2 reanalysis datasets and analyzed flux products. Their surface net heat flux and four flux components are assessed against those of buoy observations in Fig. 14b, where the chosen criteria are the daily mean difference during the calm-to-TC transition in the above analyses. The observed daily mean difference is calculated from the composite diurnal cycle from Figs. 12a–e, and those from the reanalysis and OAFlux follow the same method. It is surprising to note the large discrepancies in the reanalysis datasets (Table 4). ERA5 suffers from a significant negative bias \((-59 \text{ W m}^{-2}\)) in Qnet, which is largely due to the overestimated LH (with bias \(-76 \text{ W m}^{-2}\)). In contrast, MERRA2 suffers from a significant positive bias (50 W m\(^{-2}\)) in Qnet, which is largely due to the overestimated SW (with bias 41 W m\(^{-2}\)). A weaker LW is found in ERA5 (with bias \(-14 \text{ W m}^{-2}\)) and MERRA2 (with bias \(-15 \text{ W m}^{-2}\)) than in the buoy observations. The SH anomaly during TC passages from MERRA2 is equivalent to that of the buoy observations, whereas that from ERA5 is 7 W m\(^{2}\) lower than the buoy observations. Although OAFlux yields high-quality data, as evidenced by the general consistency of six-month mean LH with the buoy observations (in contrast to the significant bias in ERA5 and MERRA2 LH), the LH in OAFlux during TCs is 34 W m\(^{-2}\) higher than that of the buoy observations. However, the SH in OAFlux during TCs is 9 W m\(^{2}\) lower than that of the observations. The apparently poor ability of OAFlux, ERA5, and MERRA2 to resolve TC-associated extreme heat flux changes is discouraging. It is therefore necessary to further check the overall performance of the reanalysis and flux product data on the synoptic scale to identify the major error sources and help improve the flux accuracy.

c. Comparison of monthly mean heat fluxes

Figure 15 shows the monthly mean results of Qnet, SW, LW, LH, and SH from December 2018 to May 2019. In general, the Qnet at the observation site switches from positive (austral summer) to negative (austral winter), with seasonal decreases in SW and increases in LW and LH. Compared to the buoy
observations, ERA5 yields better Qnet results than MERRA2. The root-mean-square (RMS) error (summarized in Table 4) of the Qnet differences between buoy observations and ERA5 is 12 W m\(^{-2}\) in magnitude, which is much smaller than that between the buoy measurements and MERRA2 (27 W m\(^{-2}\)). The same RMS error for SW (17 W m\(^{-2}\)) can be found between the observations and two reanalyses. ERA5 produces a six-month higher SW (14 W m\(^{-2}\)) than the buoy measurements, while MERRA2 shows a four-month higher SW and two-month lower SW than the buoy results. Over six months, both ERA5 and MERRA2 indicate higher estimates of LW and LH (cooling effect) with respect to the observations. The LW and LH in ERA5 are slightly closer to those of the observational results, with lower RMSs than those of MERRA2. The LH in OAFlux shows a four-month (December–March) lower estimate and a two-month higher estimate than that of the observations. Positive and negative differences offset each other and cause a better six-month mean value in OAFlux than in ERA5 and MERRA2 (Fig. 14a) compared to the observations. OAFlux estimates a smaller SH magnitude than the reanalysis and observations, resulting from the weak air–sea temperature difference (see the following analysis). The RMS error for SH between the buoy observations and OAFlux is 4 W m\(^{-2}\), whereas that between the buoy observations and the reanalysis is only 1 W m\(^{-2}\).

d. Comparisons of daily mean variations in heat fluxes and associated air–sea variables

The daily mean Qnet and its four components SW, LW, LH, and SH from ERA5, MERRA2 and OAFlux are compared against those of the buoy observations, as shown by time series of heat flux differences (Fig. 16) and in Taylor diagrams (Fig. 17). From December 2018 to May 2019, significantly larger heat flux discrepancies can be found during TC passages (Fig. 16) than during calm weather. In addition, the heat flux discrepancies show case-to-case discrepancies during the three TCs. During TC Riley, surface cooling (negative Qnet) in ERA5 and MERRA2 is higher than that in the buoy observations, which is induced by excessive LW + LH + SH in ERA5 and MERRA2, although the SW in ERA5 and MERRA2 is slightly higher than that of the buoy results at a magnitude of approximately 100 W m\(^{-2}\). However, significantly higher Qnet in ERA5 and MERRA2 can be found during TC Veronica, which is determined by the higher SW and lower LW + LH in the reanalysis products than in the buoy observations. Compared to MERRA2, ERA5 shows a better time series of Qnet relative to the daily mean buoy observations (Fig. 16a), primarily due to a better estimate of SW (Fig. 16b). The LW in ERA5 and MERRA2 show larger discrepancies from late February to early March in 2019. The LH and SH differences between products (ERA5, MERRA2, and OAFlux) and buoy observations are strongly affected by synoptic-scale processes.

Even though ERA5 and MERRA2 cannot capture the extreme Qnet changes during TCs, as concluded in section 4b and shown in Figs. 14b and 16a, it is encouraging to see that they both statistically reproduce the high-frequency net heat flux variability well (Fig. 17a), with similar standard deviation (STD) magnitudes, high correlations and small RMS errors compared with the buoy measurements. It is important to note that the good agreement of the Qnet estimates does not necessarily guarantee that each of its four components is accurate, as is clearly reflected in the diverse distribution in the Taylor diagrams (Figs. 17b–e). During the entire observational period, the correlation coefficient between Qnet in ERA5 (MERRA2) and buoy observations is 0.95 (0.88). The Qnet value of ERA5 is relatively closer to that of the buoy result than that of MERRA2 due to a better simulation of SW (Fig. 17b). The qualities of LW, LH, and SH in ERA5 and MERRA2 compared to buoy observations are nearly equivalent in the Taylor diagrams and have acceptable uncertainties. However, larger uncertainties indicate the poor ability of the reanalysis products to produce reasonable extreme air–sea heat fluxes during TC passages (Fig. 16). Such stratification of the quality level among the four flux components is consistent with the quality level of Qnet in Figs. 16a and 17a, considering the dominant role of SW and LH in determining Qnet.

The above comparison reveals a less satisfactory quality level of the turbulent flux LH and SH, and this requires further

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**Table 3. Summary of six-month (December 2018–May 2019) mean heat flux components (columns 2–6 with overbars; W m\(^{-2}\)) based on various datasets and heat flux differences (columns 7–11 with primes) between calm weather conditions and TCs, as shown in Fig. 14.**

<table>
<thead>
<tr>
<th></th>
<th>Qnet (W m(^{-2}))</th>
<th>SW (W m(^{-2}))</th>
<th>LW (W m(^{-2}))</th>
<th>LH (W m(^{-2}))</th>
<th>SH (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>MERRA2</td>
<td>46</td>
<td>254</td>
<td>–70</td>
<td>–133</td>
<td>–6</td>
</tr>
</tbody>
</table>

**Table 4. The RMS errors of the differences between the monthly mean measured heat fluxes (W m\(^{-2}\)) and those in the products (ERA5, MERRA2, and OAFlux).**

<table>
<thead>
<tr>
<th></th>
<th>Qnet (W m(^{-2}))</th>
<th>SW (W m(^{-2}))</th>
<th>LW (W m(^{-2}))</th>
<th>LH (W m(^{-2}))</th>
<th>SH (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>RMS (buoy, ERA5)</td>
<td>12</td>
<td>17</td>
<td>12</td>
<td>14</td>
<td>1</td>
</tr>
<tr>
<td>RMS (buoy, MERRA2)</td>
<td>27</td>
<td>17</td>
<td>14</td>
<td>19</td>
<td>1</td>
</tr>
<tr>
<td>RMS (buoy, OAFlux)</td>
<td>—</td>
<td>—</td>
<td>—</td>
<td>15</td>
<td>4</td>
</tr>
</tbody>
</table>
diagnosis to understand the potential error sources, especially during TC passages. Following Eqs. (5) and (6), the relevant daily mean variables SST, SAT, $\Delta T$, specific humidity of air (QA), $\Delta Q$, and WS from reanalysis and OAFlux are compared with the buoy observations (Fig. 18). Reanalysis products show clear positive bias in $\Delta T$ (Fig. 18c) and $\Delta Q$ (Fig. 18e) against buoy observations. The positive bias in air–sea temperature difference is primarily induced by a colder atmosphere (lower SAT) in the reanalysis products. Similar to $\Delta T$, the higher estimates of $\Delta Q$ in the reanalysis products are attributed to a drier atmosphere (lower QA), which produces a higher LH in ERA5 and MERRA2, as shown in Fig. 14a. Their WS at the buoy observational heights of 4 m in terms of the Smith (1988) algorithm is less biased but shows unreasonable overestimates during the three TCs (Fig. 18f), which is consistent with the overestimated values of the TC-associated LH and SH changes in MERRA2 and ERA5.

Day-to-day variations in variables from OAFlux are directly compared with buoy observations since no hourly data are available. OAFlux exhibits an upward trend in $\Delta Q$ (LH, Fig. 16) than the observations but a higher estimate of $\Delta Q$ (LH) in April and May. A warmer atmosphere and higher SAT (Fig. 18b) are found in OAFlux than in the observations, which accounts for the lower bias of the $\Delta T$ error (Fig. 18c). Compared with the buoy observations, OAFlux features a higher QA from December 2018 to February 2019 but a lower QA from March to May 2019. This explains the upward trend (Fig. 18e) in the $\Delta Q$ error. The WS of OAFlux at a translated height of 4 m also shows good agreement with that of the observations, with a small mean difference of 0.2 m s$^{-1}$ and an STD of 0.7 m s$^{-1}$. However, the daily mean bias in air–sea variables shown in Fig. 18 is strongly affected by synoptic-scale processes, for example, TC passages.

The above time series comparison is further translated into the statistical Taylor diagram (Fig. 19) to obtain a clearer picture. Once again, the wind quality is fairly good for the reanalysis products and OAFlux (Fig. 19f). The daily mean SST
is well reproduced in the reanalysis products and OAFlux, with high correlation coefficients over 0.9. The SST in MERRA2 and ERA5 shows a better quality than that of OAFlux compared to the buoy observations. This might be due to the relatively low resolution (1°) in the current OAFlux project. A high-resolution version (0.25°) is expected to better incorporate the high-resolution OISST. Similar results can also be seen for SAT (Fig. 18b) and QA (Fig. 18d) between the products and observations. However, the $\Delta Q$ and $\Delta T$ are not well reproduced in all three datasets, resulting in less significant correlations and relatively large RMS errors against the buoy observations and representing considerable room for further improvement. The discrepancy identified here also calls for the enhancement of in situ measurements of SAT and RH in the regional/global ocean observing systems, considering the fact that present remote sensing technology still does not resolve the retrieval of surface atmospheric variables well.

5. Summary and discussion

This study addresses the requirements for better understanding the complex air–sea interactions in the data-sparse tropical southeastern Indian Ocean, which hosts a rich variety of important oceanic, atmospheric and climate processes, such as Indonesian Throughflow (ITF; Gorden 2005; Yuan et al. 2018), TC, MJO, Indian Ocean dipole (IOD; Saji et al. 1999), and Ningaloo Niño (Feng et al. 2013). Coupled models suffer from poor performance in this wider Maritime Continent region, and the errors could load their impacts onto remote regions through teleconnection processes; this is a driving motivation of the recent Year of Maritime Continent project (YMC Science Plan available at http://www.jamstec.go.jp/ymc/docs/YMC_SciencePlan_v2.pdf). The present study provides the first 6-month (December 2018–May 2019) time series data of in situ heat flux measurements in this critical ocean, meeting the above goals by depicting the extreme air–sea heat flux during TCs, validating reanalysis data, and analyzed flux products.

New data shed light on the magnitude of extreme heat flux during TCs, which has rarely been measured. For three cyclones during the observation period, namely, TC Riley, severe TC Veronica, and TC Wallace, the composite (averaged over three TCs) diurnal cycle of the surface net heat flux ($Q_{net}$) and its four components (SW, LW, LH, and SH) are constructed during the TC passage period and compared with that during the calm weather before the TC. The most striking feature is the dramatic suppression (mainly during the daytime) in the composite diurnal cycle of $Q_{net}$ during TC passage, with a mean daytime (nighttime) reduction of 470 (131) W m$^{-2}$, a maximum reduction of 695 W m$^{-2}$ around noon (1200–1400 LT), and an extreme decrease observed during TC Riley of 800 W m$^{-2}$. The extensive mean $Q_{net}$ decreases during TCs are attributed to the dominant contribution from the decreased SW (mean daytime reduced magnitude of 370 W m$^{-2}$), followed by the significant contribution from the increased LH (daytime and nighttime mean increased magnitude of 118 and 118 W m$^{-2}$, respectively), followed by a comparable contribution from the decreased LW (daytime and nighttime mean decreased magnitude of 118 and 118 W m$^{-2}$, respectively), and finally supplemented by the minor SH enhancement of 27 and 39 W m$^{-2}$ in daytime and nighttime, respectively. The mean
RF during TC passages is found to be 2 and 15 W m\(^{-2}\) in the daytime and nighttime, respectively, and depends on the magnitude of the precipitation.

The six-month time series heat flux data provide a valuable chance to make a detailed assessment of reanalysis datasets, such as ERA5 and MERRA2, and analyzed flux products, such as OAFlux, based on long-term means, extreme values, seasonal cycles, and synoptic variability. The main conclusions of the reanalysis data assessment can be summarized as follows:

1) For the six-month mean, ERA5 exhibits better consistency with the observations than does MERRA2. OAFlux also shows good consistency with observations in terms of LH and SH. However, the resolution (1\(^{\circ}\)) is lower.

2) During TCs, neither ERA5 nor MERRA2 capture the extreme heat flux changes. In addition, OAFlux cannot effectively capture the extreme changes of LH and SH as a result of the biases in air–sea variables, although the same algorithm COARE 3.0 is used for the calculation of the buoy observations. ERA5 suffers from a significant negative bias (−59 W m\(^{-2}\)) in Qnet, largely due to the overestimated LH (with bias −76 W m\(^{-2}\)) associated with a drier atmospheric environment. In contrast, MERRA2 suffers from a significant positive bias (50 W m\(^{-2}\)) in Qnet, largely due to the overestimated SW (with bias 41 W m\(^{-2}\)), which may be induced by a heavier cloudiness in MERRA2.

3) On the synoptic scale, both ERA5 and MERRA2 can reproduce the daily mean Qnet variability. Among its four flux terms, SW in ERA5 exhibits good quality, but the LW terms in the reanalysis are of poor quality.

4) Errors in the turbulent flux terms are found to be mainly due to poor performance in \(\Delta Q\) and \(\Delta T\), while the WS quality is reasonable.

It should be noted that the buoy observations of air–sea variables and estimated heat fluxes are not free of uncertainty.
In addition, being affected by the instrumental accuracies (Table 1), the uncertainty of the heat fluxes may be impacted by the buoy status on the sea, the choice of algorithms employed for estimating the heat fluxes and the averaging methods for the high-resolution variable samples. For example, the choice of surface albedo value can significantly affect the eventual net SW at the surface (e.g., Fig. 3). At present, how much downward SW is biased due to the pitch-and-roll movements remains unknown and needs to be explored and verified in future studies. The net LW is not only affected by the direct accuracy of the downward SW but also strongly dependent on the choices of LW parameterizations regarding SST, SAT, and so on (Fung et al. 1984). In this study, the observed air–sea turbulent heat fluxes are empirically estimated based on COARE 3.0, which is different from the Louis scheme in the reanalysis system. The choice of algorithms for estimating turbulent heat fluxes accounts for the major heat flux discrepancies among different heat flux products (Yu 2019). Thus, quality control is undoubtedly a priority for observations, which are an important test bed for the validation of reanalysis and other products.

In summary, the present study demonstrates the critical value of in situ heat flux observations over the wider Maritime Continent region. The six-month time series data reveal extreme heat flux changes during TCs, which have rarely been measured, convey complex synoptic variability, and partly
capture the seasonal cycle. All these characteristics are valuable in assessing the quality of the reanalysis datasets, such as ERA5 and MERRA2 adopted here, and the objectively analyzed flux products, such as OAFlux. The most serious problem identified here is the inability of ERA5 and MERRA2 to properly resolve the extreme heat flux changes during TCs. The less satisfactory quality in the turbulent heat flux in ERA5, MERRA2, and OAFlux needs further attention and more SST and RH measurements from regional/global ocean observing systems.

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