

On the Diurnal Cycle in UTLS Mesoscale Potential Vorticity Variability Using 9-km-Resolution ECMWF Operational Forecast Analyses: A Climatology of South Pacific Tropical Cyclones during 2016–24

SHELLIE M. ROWE^a AND MATTHEW H. HITCHMAN^a

^a *Department of Atmospheric and Oceanic Sciences, University of Wisconsin–Madison, Madison, Wisconsin*

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ABSTRACT: This study investigates mesoscale potential vorticity (PV) variability in the upper troposphere and lower stratosphere (UTLS) in tropical cyclones (TCs) using 9-km-resolution, 6-hourly ECMWF operational forecast analyses, to characterize its behavior during TC life cycles. TCs exhibit a diurnal cycle in deep convection with a maximum overnight. Since mesoscale variability of UTLS PV is caused by deep convection, it should also exhibit a diurnal cycle. Spatial standard deviations σ_{PV} and means are calculated in a $6^\circ \times 6^\circ$ domain following a storm track. The distinctive maximum in σ_{PV} at the tropopause is caused by the upward increase in mean static stability. A key time was chosen to be the first 1800 LT after reaching category 1 (C1). All storms reached category 2 (C2) the night after the key time. We required at least a 6-day track and compared 3 days before and after reaching C2. This method was first applied to TC Yasa. We then analyzed a total of 23 TCs which met these selection criteria in the South Pacific during 2016–24 [13 C1–C2 and 10 category 3 (C3)–category 5 (C5)]. A strong diurnal cycle in 100-hPa σ_{PV} was found, with a development phase (days 1–3) and a mature phase (days 4–6). From days 1–3 to days 4–6, there is a reduction in mean σ_{PV} by $\sim 25\%$ and in diurnal range by $\sim 45\%$ for C1–C2 and $\sim 60\%$ for C3–C5 and elimination of the preference for a 0600 LT peak. These results are consistent with cloud-radiative forcing theories of TCs.

SIGNIFICANCE STATEMENT: We investigated variability of mesoscale potential vorticity (PV) in tropical cyclones (TCs) in high-resolution global forecast model output to characterize its behavior and contribute toward understanding TC life cycles. We analyzed 23 TCs in the South Pacific during 2016–24. The spatial standard deviation of PV (the product of relative vorticity and static stability) is a sensitive measure of updrafts which reach the tropopause. It exhibits a diurnal cycle at the tropopause, with a 1800 LT minimum. Prior to reaching strength C2 (development phase), peak amplitude tends to occur at 0600 LT. After reaching C2 (mature phase), this preference disappears, and its range decreases by half. These results are consistent with cloud-radiative forcing theories of the diurnal cycle in TCs.

KEYWORDS: Tropical cyclones; Potential vorticity; Convective adjustment; Diurnal effects; South Pacific Ocean; Statistics

1. Introduction

a. UTLS mesoscale PV and its diurnal cycle in TCs

We have been investigating mesoscale horizontal potential vorticity (PV) dipoles in the upper troposphere and lower stratosphere (UTLS) in tropical cyclones (TCs). Hitchman and Rowe (2017) studied UTLS PV dipoles in TC Talas and found that the inertially unstable member tends to lie radially outward and that outflow PV filament pairs can merge with the subtropical jet. Hitchman and Rowe (2019) compared UTLS PV dipoles and filament pairs in TCs Talas, Edouard, and Ita and found amplitudes exceeding ± 10 PVU ($1 \text{ PVU} = 1 \times 10^{-6} \text{ K kg}^{-1} \text{ m}^2 \text{ s}^{-1}$). They showed that UTLS mesoscale PV structures occur only when updrafts reach the tropopause. In the equation for relative vorticity ζ , dipole production is caused by an updraft in the tilting/twisting term (Davies-Jones 1984; Chagnon and Gray 2009; Conzemius and Montgomery

2009; Davis and Galarneau 2009). As an updraft arrives at the tropopause, flow will accelerate locally due to vertical advection, with a resulting vorticity dipole.

Mesoscale PV structures can enhance the longevity of convective phenomena (Weijenborg et al. 2017). UTLS PV dipoles are associated with dynamical adjustment toward a balanced state and radiation of gravity waves (Snyder et al. 2007; Wang et al. 2009; Hernandez-Duenas et al. 2013; Plougonven and Zhang 2014). Cohen et al. (2017) found that nonbalance in the UTLS is related to TC intensification. Here, we use a statistical approach to investigate UTLS mesoscale PV variability, its diurnal cycle, and how it evolves during a TC life cycle.

To analyze the aggregate behavior of mesoscale PV features, we calculated the spatial mean ($\overline{PV}$) and standard deviation (σ_{PV}) following the track of a TC and applied it to 6-hourly, 9-km resolution ECMWF operational forecast analyses, available since 2016. The number of PV anomalies and their intensity both contribute toward larger values of σ_{PV} . We first studied TC Yasa, which exhibited a distinct diurnal cycle in 100-hPa σ_{PV} , with minimum amplitude at 1800 LT. After reaching strength C2, there was a reduction in mean amplitude and range of the diurnal cycle and elimination of a preference for a 0600 LT maximum.

The first goal of this paper is to analyze the evolution of the diurnal cycle of mesoscale variability in TC Yasa in detail.

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Corresponding author: Shellie Rowe, rowe1@wisc.edu

The second goal is to analyze the diurnal cycle in 100-hPa σ_{PV} during the life cycle of South Pacific TCs with statistical significance tests for observed changes in amplitude and phase. We studied a total of 23 TCs in the South Pacific during 2016–24 which passed our selection criteria. The third goal is to compare the evolution of the diurnal cycle in 100-hPa σ_{PV} with expectations from cloud-radiative forcing theories.

A fourth goal is to show that the strong local maximum in σ_{PV} at the tropopause is a result of the product of the profile of σ_z and mean static stability, which abruptly increases at the tropopause. Variations in the magnitude of 100-hPa σ_{PV} are related to the number and intensity of updrafts which reach the tropopause. In this work, we investigate the evolution of 100-hPa σ_{PV} in light of cloud-radiative forcing theories of the diurnal cycle in TCs.

b. Diurnal cycle in TCs and cloud-radiative forcing theories

Browner et al. (1977) found an early morning precipitation maximum in eight Atlantic TCs. Bowman and Fowler (2015) studied ~1500 TCs and confirmed that peak rainfall occurs in the early morning. A diurnal cycle in gravity wave activity is also observed above TCs in numerical simulations and satellite observations, which strengthens during intensification (Kim et al. 2014; Hoffmann et al. 2018; Wright 2019). The diurnal cycle in gravity wave activity is caused by a diurnal cycle in deep convection, as part of the adjustment process (Rossby 1938; Gill 1982; Raymond 1992; Lin 2007; Lane and Zhang 2011).

Cloud-top radiative cooling at night promotes deep convection by reducing static stability (Melhauser and Zhang 2014; Navarro and Hakim 2016; Trabing et al. 2019). Organized convection also traps upwelling infrared, increasing the core temperature in a “cloud greenhouse effect” (Ruppert et al. 2020). These two effects increase the lapse rate in the upper troposphere within clouds, here collectively called the “lapse rate mechanism.” The contrast between warm cores in clusters of deep convection and cooling in patches of clear air can also enhance convection, here called the “dynamic cloudy-clear mechanism.” From idealized numerical simulations of radiative feedbacks in TCs, Ruppert and Hohenegger (2018) and Ruppert and O’Neill (2019) concluded that the lapse rate mechanism peaks at midnight and is the primary governor of the diurnal cycle amplitude, while the dynamic cloudy-clear mechanism amplifies the diurnal range and delays the peak by up to ~5 h.

Downs and Kieu (2020) showed that the intensity of observed TCs in the northwest Pacific and North Atlantic is particularly responsive to variations in upper-tropospheric static stability. Using hyperspectral infrared soundings, Duran et al. (2021) found a destabilization of the mid- to upper troposphere near the inner core at night. It will be shown below that $\overline{PV}$ at 150 at 200 hPa exhibits a diurnal cycle consistent with reduced static stability overnight.

Muller and Roms (2018) found that infrared emission from clear air can facilitate amplification. Wing (2022) showed that development occurs sooner when cloud-radiative feedbacks are

included. Chen et al. (2023) found that diurnal variation provides a favorable environment for rapid intensification (RI). Kaplan and DeMaria (2003) define RI to be an increase in maximum wind speed of at least 30 kt (1 kt $\approx 0.51 \text{ m s}^{-1}$) in 24 h. They found that ~30% of all TCs and 100% of category 4 (C4) and category 5 (C5) TCs in the North Atlantic undergo RI.

In the genesis phase of a TC, organized convection tends to occur in mesoscale convective cloud clusters, with regions of clear air between. This would promote the dynamic cloudy-clear mechanism, which amplifies the diurnal range and shifts the peak toward sunrise. This contrast at the cloud-cluster scale disappears when a symmetric vortex forms in the core region during intensification, with a uniform cloudy disk [cf. Figs. 4 and 5 in Houze (2010)]. Miyamoto and Takemi (2015) showed that RI in simulated TCs “commenced following the formation of a circular cloud, which occurred about 12 h after the TC became essentially axisymmetric.” A reduction in area with clear/cloudy contrast at the cloud-cluster scale upon achieving axisymmetry might imply a reduction in amplitude and a reduced delay in peak timing. Dominance of the lapse rate mechanism during the mature phase may imply a tendency to peak closer to midnight.

In this work, we document changes in the diurnal cycle in σ_{PV} before and after reaching category 2 (C2) and compare results with expectations from cloud-radiative forcing theories. To test the statistical significance of results, we pose the following hypotheses:

- H1: There is a diurnal cycle in mesoscale PV variability in the UTLS above TCs, with a minimum near sunset and a maximum at 0600 LT.
- H2: The mean and range of the diurnal cycle in UTLS mesoscale PV variability decrease after reaching C2.
- H3: The preference for a sunrise amplitude maximum in UTLS mesoscale PV variability disappears after reaching C2.

c. Plan of paper

A statistical approach is introduced to study mesoscale dynamical variability following the motion of TCs in 9-km ECMWF forecast analyses. The spatial standard deviation and mean are calculated for each altitude for specified domains moving with the storm track: $\sigma_{PV}(z, t)$ and $\overline{PV}(z, t)$. These are scalar quantities, which are invariant to coordinate rotation. This method was first used to study the evolution and diurnal cycle of TC Yasa, which occurred near Fiji during 12–20 December 2020.

We then sought statistical confirmation by studying other TCs in the South Pacific basin. A diurnal cycle in TCs is found in all basins around the world (Bowman and Fowler 2015). However, each basin has distinctive characteristics, including the distribution of sea surface temperature (SST), land, proximity to midlatitude troughs, precursor waves, wind shear, and airborne dust. The South Pacific basin (~10°–30°S, 140°E–150°W) coincides with the South Pacific convergence zone and the west Pacific warm pool (Chand 2020). It is distinguished by having a large region of uniformly high SSTs, relatively little land to alter the diurnal cycle, and relatively little influence by midlatitude troughs. Precursors in the South Pacific include equatorial

Rossby waves, mixed Rossby–gravity waves, Kelvin waves, and tropical depressions (Schreck et al. 2012). For this study of the evolution of the diurnal cycle in σ_{PV} , we analyzed a total of 23 TCs which occurred in the South Pacific basin during 2016–24.

Analysis methods are described in section 2, including datasets, case selection criteria, compositing techniques, and significance testing. Results for TC Yasa are described in section 3. A comparison of features in ECMWF 100-hPa PV and *Himawari-8* satellite cloud-top temperature illustrates the close relationship between regions of enhanced mesoscale variations and updrafts that penetrate the tropopause. The diurnal cycle in 100-hPa σ_{PV} is shown to be in phase with the diurnal cycle in vertical velocity variance, small mean PV in the upper troposphere, and cold cloud-top area. Section 4 presents time series of 100-hPa σ_{PV} and statistical analysis of the diurnal cycle for the 23 TCs with a 6-day record, comparing days 1–3 (development) and days 4–6 (mature). Section 5 discusses the life cycle of the diurnal cycle in 100-hPa σ_{PV} and compares results with cloud-radiative forcing theories. Conclusions are given in section 6.

2. Analysis methods and datasets

a. Statistical definitions

The mean value of PV over a square domain (I, J) at a given level is given by

$$\overline{PV} = \frac{1}{I \times J} \sum_{i,j} PV(i, j).$$

Mesoscale variability of PV is characterized by the spatial standard deviation,

$$\sigma_{PV} = \sqrt{\overline{PV^2}} = \sqrt{\frac{1}{I \times J} \sum_{i,j} [PV(i, j) - \overline{PV}]^2}.$$

RMS and mean values of PV are expressed in PVU. Results are shown for a $6^\circ \times 6^\circ$ “full domain” following the moving storm track of the storm center. Selected results for Yasa are also shown for a $2^\circ \times 2^\circ$ “inner domain.” The square domains are compatible with the gridded model and satellite data and provide a reference frame for comparing features in ECMWF 100-hPa PV and satellite cloud-top temperature. This method is applied at each level and time, with time–altitude sections showing the “full storm evolution” of a TC life cycle.

We chose 100 hPa as a level of particular interest, since σ_{PV} exhibits a distinctive maximum at 100 hPa in all South Pacific TCs, except for two of the strongest storms, which exhibited large values of σ_{PV} at 70 hPa. We also calculated the standard deviations of ECMWF temperature σ_T , pressure velocity σ_ω , relative vorticity σ_ζ , and divergence $\sigma_{\nabla \cdot \mathbf{v}}$.

b. Datasets

This study uses 6-hourly 0.08° (~ 9 km) resolution ECMWF operational forecast analyses, available beginning 1 January 2016. At the equator, the spacing between gridded values is 7.8 km, with the number of grid points in longitude decreasing

with latitude to maintain similar spacing. The vertical resolution in the UTLS of this pressure-coordinate model was upgraded from ~ 400 to ~ 300 m (Polichtchouk et al. 2022). The “ $6^\circ \times 6^\circ$ domain” refers to a square grid of 86×86 points, centered at the location of the storm track, which is 6° in latitude and longitude at the equator. The range of horizontal wavelengths for mesoscale variability which are resolved is ~ 35 – 600 km for the $6^\circ \times 6^\circ$ domain. Bechtold et al. (2013) showed that a new ECMWF convective scheme based on boundary layer CAPE markedly improved the timing and amplitude of the diurnal cycle. The maximum in deep convection in the South Pacific in ECMWF forecasts is in the early morning, in agreement with TRMM observations.

We have also investigated 100-hPa absolute vorticity in 28-km-resolution NOAA GFS data and 100-hPa PV in 27-km-resolution ERA5 data for TC Yasa. UTLS mesoscale curvilinear structures in absolute vorticity and PV are coarser and much smaller in amplitude in the 27- and 28-km-resolution data compared to the 9-km ECMWF data. Mesoscale PV features in 15-km-resolution ECMWF analyses for TCs Talas, Edouard, and Ita are also poorly represented relative to 9-km-resolution ECMWF analyses. The degraded horizontal resolution in 28-km GFS, 15-km ECMWF, and 27-km ERA5 data leads to irregular, small values of σ_{PV} relative to 9-km ECMWF data.

Since the 27-km ERA5 data have 1-h time resolution, we tried calculating the diurnal cycle in 100-hPa σ_{PV} for TC Yasa. The resulting time series is compared with that for 6-hourly 9-km ECMWF data in Fig. S1 in the online supplemental material. The coarser 27-km resolution markedly degrades the amplitude of curvilinear mesoscale PV structures. The degraded spatial resolution renders the temporal signal weak and variable relative to 9-km ECMWF, making ERA5 unsuitable for evaluating the timing of the overnight peak.

ECMWF forecasts are regularly used by WMO warning centers and commercial enterprises, with TC track forecasts achieving a high degree of accuracy (Magnusson et al. 2019). This dataset provides the opportunity to study mesoscale PV at 9-km resolution for any TC in the world since 2016. However, heating rates and static stability factors are not available, the 6-h archival interval does not allow for calculating tendencies for mesoscale processes in conservation equations, and the vertical resolution of archived data makes it difficult to calculate vertical derivatives near the tropopause. Nevertheless, insight can be gained by investigating the statistical properties of the diurnal cycle in mesoscale PV features. This statistical approach is consistent with the nature of convection-dominated flows in TCs.

Gridded fields of cloud-top temperature and cloud-top pressure from the 2-km-resolution *Himawari-8* satellite are used to identify locations of deep convection and to compare with fields of ECMWF 100-hPa PV for TC Yasa. The location of the 202-K contour is used to indicate the presence of deep convection. The diurnal cycle in fractional area with cloud-top temperature < 202 K will be compared with the diurnal cycle in 100-hPa σ_{PV} .

To help characterize the growth potential for TCs, SST maps were obtained from the NOAA climate data archive. This 0.25° -resolution dataset was created at NOAA by modeling diurnal variability in conjunction with Advanced Very High Resolution Radiometer (AVHRR) observations.

For TC Yasa, four different storm track estimates were used to evaluate sensitivity of results to storm location. We created an “ECMWF storm track” by inspection of 100-hPa PV charts, where the curvilinear aspect of 100-hPa PV features was used to subjectively identify the center of the radius of curvature every 6 h. Storm tracks were also obtained from the Cooperative Institute for Meteorological Satellite Studies (CIMSS), the Joint Typhoon Warning Center (JTWC), and the International Best Track Archive for Climate Stewardship (IBTrACS), which are based on the Advanced Dvorak Technique (Olander and Velden 2009). The JTWC and IBTrACS estimates are available from NOAA (Schreck et al. 2014). Since IBTrACS has emerged as an international standard, IBTrACS is used as the standard for this study.

Estimates of storm intensity using the Saffir–Simpson scale [tropical depression (TD), tropical storm (TS), category 1 (C1)–category 5 (C5)], minimum mean sea level pressure (MSLP), and maximum wind speed ($V_{\max}$) are provided by CIMSS, JTWC, and IBTrACS. Threshold wind speeds for intensity categories are TS (34 kt), C1 (64 kt), C2 (83 kt), C3 (96 kt), C4 (113 kt), and C5 (137 kt). The IBTrACS users guide suggests an uncertainty in wind speed of ± 10 kt and in position of 10–15 km for $V_{\max}$ exceeding 100 kt and 30–40 km for $V_{\max}$ less than 60 kt. Storm tracking begins when circulation around a center can be identified—a TD. After reaching 34 kt, it is designated a TS, with values of MSLP and $V_{\max}$ then recorded in IBTrACS. Strength and duration values were cross-checked with the NOAA Historical Hurricane Track database.

Fiji standard time (FJT) is 12 h ahead of UTC. Time references will be given primarily in FJT or local time. Longitude (hence local time) can vary along a TC storm track. In creating local time composites for the four ECMWF data times 0000 UTC (“T1”), 0600 UTC (“T2”), 1200 UTC (“T3”), and 1800 UTC (“T4”), we used local time bins 1000–1300 FJT (T1, noon), 1600–1900 FJT (T2, sunset), 2200–0100 FJT (T3, midnight), and 0400–0700 FJT (T4, sunrise), respectively. These local time windows correspond to the longitude range $\sim 140^\circ\text{E}$ – 160°W .

c. ECMWF PV

Ertel’s PV is defined to be the product of static stability and absolute vorticity (Hoskins et al. 1985):

$$P = \frac{1}{\rho_o} \nabla \theta \cdot (f + \zeta) = \frac{1}{\rho_o} \frac{\partial \theta}{\partial x} (\zeta_1) + \frac{1}{\rho_o} \frac{\partial \theta}{\partial y} (\zeta_2) + \frac{1}{\rho_o} \frac{\partial \theta}{\partial z} (f + \zeta_3), \quad (1)$$

where the components of relative vorticity are $\zeta_1 = (\partial v / \partial z) - (\partial w / \partial y)$, $\zeta_2 = (\partial u / \partial z) - (\partial w / \partial x)$, and $\zeta_3 = (\partial v / \partial x) - (\partial u / \partial y)$; $\theta = T(1000/p)^{R/c_p}$ is the potential temperature; $f = 2\Omega \sin \phi$ is the Coriolis parameter; ρ_o is the basic-state density; and wind components are u , v , and w [Holton and Hakim 2013; Eq. (3.1.4) in Andrews et al. 1987]. At ECMWF, PV is calculated from the forecast values of u , v , ω , and T on pressure surfaces at each time, rather than by integrating a PV conservation equation (ECMWF 2023). In pressure coordinates, Ertel’s PV is evaluated from the distributions of u , v , ω , and T using

$$P = \frac{1}{\rho_o} \frac{\partial \theta}{\partial z} \left(f + \frac{\partial v}{\partial x} - \frac{\partial u}{\partial y} \right) - \frac{1}{\rho_o} \frac{\partial \theta}{\partial x} \left(\frac{\partial v}{\partial z} - \frac{\partial w}{\partial y} \right) + \frac{1}{\rho_o} \frac{\partial \theta}{\partial y} \left(\frac{\partial u}{\partial z} - \frac{\partial w}{\partial x} \right), \quad (2a)$$

where $\partial z = -dp/(\rho_o g)$ and $w = \omega/(\rho_o g)$. When isentropic surfaces are oriented primarily horizontally, scale analysis shows that the other two terms are much smaller than the first, so that

$$P = \frac{1}{\rho_o} \frac{\partial \theta}{\partial z} (f + \zeta_3) = -g \frac{\partial \theta}{\partial p} (f + \zeta_3) = S(f + \zeta_3), \quad (2b)$$

where $S = -g(\partial \theta / \partial p)$ is the static stability parameter. Millán et al. (2021) used (2b) to evaluate how consistently modern global reanalysis datasets represent PV.

The coarseness of archived vertical resolution makes it difficult to calculate static stability accurately from temperature. However, since ECMWF provides P and ζ_3 calculated from the high-vertical resolution model grid, one may estimate S from

$$S = P/(f + \zeta_3) \quad (3)$$

(see also Dunkerton and Delisi 1986). An anomalous value of PV (sign opposite to f in a given hemisphere) can indicate either inertial instability or static instability. One may use $f(f + \zeta_3) < 0$ to indicate inertial instability (anomalous absolute vorticity) and $S < 0$ to indicate convective instability.

The relative contributions of perturbations in relative vorticity and temperature to perturbations in PV may be estimated by linearizing (2b) and subtracting $\overline{PV}$:

$$P' \approx \frac{1}{\rho_o} \frac{\partial \bar{\theta}}{\partial z} \zeta' + \frac{1}{\rho_o} \frac{\partial \theta'}{\partial z} (f + \bar{\zeta}) = P'_\zeta + P'_T. \quad (4)$$

A scale analysis of the ratio of these two terms shows that $P'_T/P'_\zeta \sim [(f + \bar{\zeta})/\zeta'](\theta'/\bar{\theta})$, so that the two contributions to σ_{PV} have the ratio $[(f + \bar{\zeta})/\sigma_\zeta](\sigma_\theta/\bar{\theta}) = [(f + \bar{\zeta})/\sigma_\zeta](\sigma_T/\bar{T})$. It will be shown that at 100 hPa, $\sigma_T \sim 1$ K and $\sigma_\zeta \sim 10^{-4} \text{ s}^{-1}$. Assuming that near the tropopause $\bar{T} \sim 200$ K and that $\bar{\zeta} \sim 1.5 \times 10^{-5} \text{ s}^{-1}$ in the anticyclonic outflow, with $f \sim -5 \times 10^{-5} \text{ s}^{-1}$, then $|(f + \bar{\zeta})/\sigma_\zeta| \lesssim 1$ and $|\sigma_T/\bar{T}| \sim [1/200]$. It is therefore expected that the first term in (4) would dominate. Assuming that $|P'_\zeta| \gg |P'_T|$, then from (3) and (4), $\sigma_{PV} \approx \bar{S} \sigma_\zeta$.

d. Storm case selection

Tropical storms were selected using IBTrACS for the period 2016–24. Our four selection criteria were as follows: 1) The storm reached at least C1 during its lifetime. 2) It occurred in the South Pacific in the longitude range 140°E – 160°W . 3) Maximum speed values are available each day. 4) The storm center can be tracked for six continuous days or more. Applying criteria 1–3, we found the following number of TCs for each year: 2016 (4), 2017 (3), 2018 (4), 2019 (4), 2020 (4), 2021 (1), 2022 (1), 2023 (5), and 2024 (1), for a total

TABLE 1. South Pacific tropical storms during 2016–24 which satisfied our selection criteria, including name, maximum strength, maximum wind speed (kt), and dates used. Storms which reached C1–C2 are listed chronologically in the upper group, while C3–C5 are listed in bold in the lower group. The four TCs which exceeded 125 kt are shown in bold italics.

Name	Strength	Max speed	Dates	Year
Victor	C1	80	14–21 Jan	2016
Amos	C2	80	18–24 Apr	2016
Debbie	C2	95	24–31 Mar	2017
Cook	C2	85	7–14 Apr	2017
Hola	C2	90	5–12 Mar	2018
Keni	C2	85	6–12 Apr	2018
Owen	C2	80	10–17 Dec	2018
Oma	C2	70	16–23 Feb	2019
Trevor	C2	75	20–27 Mar	2019
Uesi	C2	70	8–15 Feb	2020
Dovi	C2	90	7–13 Feb	2022
Gabrielle	C2	80	7–14 Feb	2023
Megan	C2	70	14–21 Mar	2024
Winston	C5	150	14–21 Feb	2016
Donna	C4	110	3–10 May	2017
Gita	C4	110	8–15 Feb	2018
Harold	C4	125	2–9 Apr	2020
Yasa	C4	125	12–19 Dec	2020
Niran	C4	110	28 Feb–7 Mar	2021
Kevin	C4	125	28 Feb–7 Mar	2023
Judy	C3	100	26 Feb–5 Mar	2023
Lola	C4	115	19–26 Oct	2023
Jasper	C4	115	4–11 Dec	2023

of 27 TCs. Criterion 4 allows us to quantify differences in the diurnal cycle during development (days 1–3) and mature (days 4–6) phases. Criterion 4 eliminated four TCs which lasted 5 days or less. This procedure yielded 23 storms during 2016–24 which satisfied our criteria, including 13 C1–C2 and 10 C3–C5 (Table 1). Three of these storms lasted only 6 days (Amos, Dovi, and Keni), while the other 20 lasted 7 days or more.

The four TCs which exceeded 125 kt showed significant variability in 70-hPa σ_{PV} , with Winston and Kevin exhibiting a maximum at 70 hPa instead of 100 hPa. We therefore include 70-hPa σ_{PV} for Winston and Kevin in the statistical analysis along with σ_{PV} at 100 hPa for the other 21 TCs.

A key time was chosen for each storm based on observed intensification to C1 in IBTrACS, which is used to superimpose time series and to form time-of-day composites for assessing H2 and H3. The key time for superposition is chosen to be 0600 UTC (1800 FJT) since a minimum in the diurnal cycle in σ_{PV} reliably occurs at 1800 FJT in each TC. The key date is defined as the date on which 1800 FJT first occurs after reaching C1. All TCs reached C2 during the night after the key time. Seven days of data were obtained for 20 TCs. Each seven daytime series is shown to illustrate the evolution of the diurnal cycle from the development to mature to decay phase of a TC. These values are then binned by time of day for all 7 days to evaluate H1. For evaluating H2 and H3, we used the

first 6 days of these 20 records, plus the 6 days available for Amos, Dovi, and Keni, for a total of 23 TCs.

Composites were formed for each of the four times of day: 1000–1300 LT (“noon”), 1600–1900 LT (“sunset”), 2200–0100 LT (“midnight”), and 0400–0700 LT (“sunrise”). We created separate time-of-day composites for the 13 storms that reached only C2 and for the 10 storms that reached C3–C5 and tested hypotheses H2 and H3 for these two subgroups and for all TCs. To test H1, we composited 20 storms for each time of day for all 7 days, which includes 560 values of σ_{PV} . To test H2 and H3, we composited all 23 storms for each time of day separately for the 2 days prior to and including the key date (days 1–3) and for the 3 days after the key date (days 4–6), which includes 552 values of σ_{PV} .

We used a one-sided Student’s t test to assess whether two means, $[X]_1$ and $[X]_2$, are significantly different from each other, given standard deviations σ_1 and σ_2 . The t statistic is given by

$$t = \frac{[X]_1 - [X]_2}{\sqrt{\sigma_1^2 + \sigma_2^2}} \sqrt{N}, \quad (5)$$

where N is the number of independent samples and $\nu = 2N - 2$ is the number of degrees of freedom. Values of t and ν generate a p value, where a p value of 0.01 indicates that the means are significantly different from each other at the 99% level. Statistical significance increases in proportion to the square root of the sample size.

3. Evolution of diurnal cycle in UTLS mesoscale variability for TC Yasa

a. Overview of TC Yasa life cycle

TC Yasa caused significant damage on Vanua Levu and Viti Levu, affecting ~100 000 people (International Federation of Red Cross and Red Crescent Societies 2022). SSTs were ~28°–29°C during Yasa’s intensification and period of maximum intensity approaching Vanua Levu (Fig. 1a). Yasa became a TS (34–63 kt) on 13 December FJT (Fig. 1b) near 15°S, 173°E, about 500 km west and 250 km north of Viti Levu (Fig. 1a). During 13–15 December, Yasa strengthened from TS to C1 (64–82 kt) (Figs. 1a,b), while exhibiting a strong diurnal cycle in 100-hPa σ_{PV} (Fig. 2). Yasa reached C2 (82–95 kt) at 1200 FJT 15 December (Fig. 1b) and began to propagate rapidly toward Vanua Levu (Fig. 1a). A satellite IR movie of Yasa’s evolution is shown in Fig. S2.

Yasa developed from C1 to C4 in ~18 h (Fig. 1b). From 1200 FJT 15 December to 1200 FJT 16 December, $V_{\max}$ increased 45 kt in 24 h, satisfying the criterion for RI. The greatest intensity was reached at 0600 FJT 17 December, with $V_{\max} \sim 125$ kt and MSLP ~ 918 hPa (Fig. 1b). CIMSS and JTWC tracks estimated that Yasa reached C5, with $V_{\max} \sim 140$ kt and MSLP ~ 914 hPa. On the afternoon of 17 December, Yasa passed over Vanua Levu and began to decay rapidly, diminishing to C1 by 0600 FJT 18 December (Fig. 1b). From 0600 FJT 17 December to 0600 FJT 18 December, $V_{\max}$ declined from 125 to 75 kt (–50 kt in 24 h), a rapid weakening event. Yasa

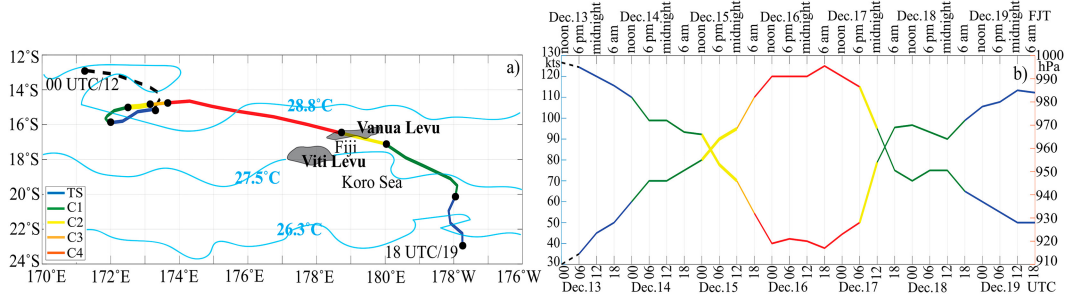


FIG. 1. (a) AVHRR SSTs (°C) at 1330 FJT 16 Dec 2020 and the storm track for Yasa (using IBTrACS) colored by the Saffir–Simpson intensity scale (color legend) and (b) evolution of MSLP (hPa; right scale) and $V_{\max}$ (kt; left scale) every 6 h, colored by storm intensity, during 1200 FJT 13 Dec–1800 FJT 20 Dec 2020. The dashed portion in (a) shows the track during 1200 FJT 12 Dec–1200 FJT 13 Dec, when Yasa was still a TD and MSLP and $V_{\max}$ were not defined.

strengthened ~ 10 kt overnight on 18 December and then diminished as it propagated southward over cooler SSTs (Fig. 1a), becoming a TS on 19 December (Fig. 1b).

b. Diurnal cycle of 100-hPa σ_{PV} in TC Yasa

A notable diurnal cycle in 100-hPa σ_{PV} is evident, with overnight maxima and a minimum at 1800 FJT, which persists for 7 days (Fig. 2a). The key time for Yasa is 1800 FJT 15 December. The maxima on days 1–3 (13–15 December) are symmetric about 0600 FJT, while the maxima on days 4–6 (16–18 December) are asymmetric about 0000 FJT, with a rapid rise and slower decay (Fig. 2a). On 19 December, a minimum in σ_{PV} occurred at 1200 FJT and a maximum at 1800 FJT (Fig. 2a), after the regions of cold cloud top and 100-hPa PV variability had become weak and disorganized.

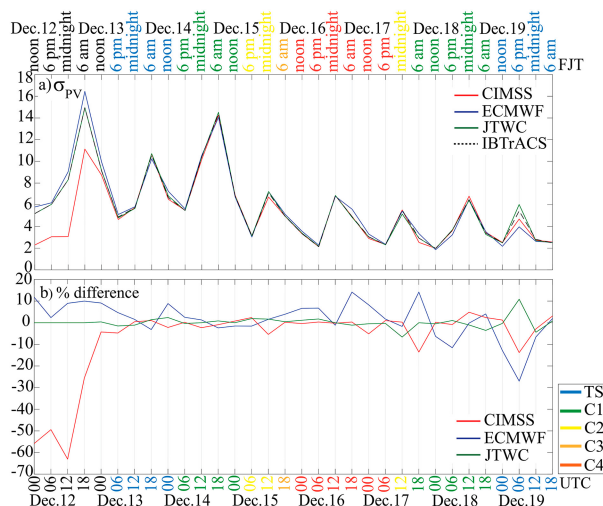


FIG. 2. (a) Time series of 100-hPa σ_{PV} for the $6^\circ \times 6^\circ$ domain from 1200 FJT 12 Dec to 1200 FJT 19 Dec 2020 for TC Yasa using storm tracks from CIMSS (red line), “ECMWF” (blue line), JTWC (green line), and IBTrACS (dashed gray line). (b) Time series of the percent difference between CIMSS and IBTrACS (red line), ECMWF and IBTrACS (blue line), and JTWC and IBTrACS (green line). IBTrACS is used for this study.

The diurnal cycle in 100-hPa σ_{PV} is nearly identical for each of the storm tracks (Fig. 2a). During the first three cycles, maxima of ~ 11 – 16 PVU occurred at 0600 FJT, with minima of ~ 6 PVU at 1800 FJT. During the next four cycles, maxima of ~ 5 – 7 PVU occurred at 0000 FJT, with minima of ~ 2 PVU at 1800 FJT. The range of ~ 5 – 16 PVU in the diurnal cycle for days 1–3 was reduced to ~ 2 – 6 PVU for days 4–6.

Percent differences from using different tracks, $[(\sigma - \sigma_{\text{IBTrACS}})/\sigma_{\text{IBTrACS}}] \times 100$, with IBTrACS as standard, are shown in Fig. 2b. JTWC and IBTrACS agree very well, with differences less than $\sim 5\%$ during amplification and $\sim 10\%$ during weakening (Fig. 2b). This suggests that there is less than 10% uncertainty in estimating 100-hPa σ_{PV} due to storm track location.

To test H1, whether the diurnal cycle in 100-hPa σ_{PV} is statistically significant, we asked whether the mean at 0600 FJT (T4) is higher than the mean at 1800 FJT (T2) and analyzed days 1–3 and days 4–6 separately. The ratio of the maximum to minimum values of σ_{PV} in the diurnal cycle for TC Yasa is 2.3 for days 1–3 and 2.8 for days 4–6 (Fig. 2a, Table 2), an average of 2.5. For days 1–3, the p value is 0.0025, which is significant at the 99.75% level. For days 4–6, the p value is 0.0004. Therefore, H1 is supported for TC Yasa.

To test H2, we asked whether the mean value of 100-hPa σ_{PV} for days 1–3 is significantly higher than the mean for days 4–6. The mean value is 7.7 PVU for days 1–3 and 4.0 PVU for days 4–6 (Fig. 2a, Table 2), with p value 0.0004, supporting H2 for Yasa. This decrease in amplitude of 50% after reaching C2 is highly statistically significant. Similarly, the diurnal range in 100-hPa σ_{PV} was ~ 8 PVU for days 1–3 and ~ 4 PVU for days 4–6. This decrease in the range of the diurnal cycle in 100-hPa σ_{PV} by 50% is also highly statistically significant.

To test H3, we asked whether the mean value of 100-hPa σ_{PV} at 0600 FJT (T4) is significantly higher than at 0000 FJT (T3). For days 1–3, the mean at 0600 FJT (13.3 PVU) is significantly higher than the mean at 0000 FJT (8.2 PVU), with a t score of 2.7 and p score of 0.06, consistent with a preference for a maximum at 0600 FJT, with 95% significance (Fig. 2a, Table 2). For days 4–6, the mean at 0600 FJT (4.2 PVU) is significantly smaller than the mean at 0000 FJT (6.5 PVU), with

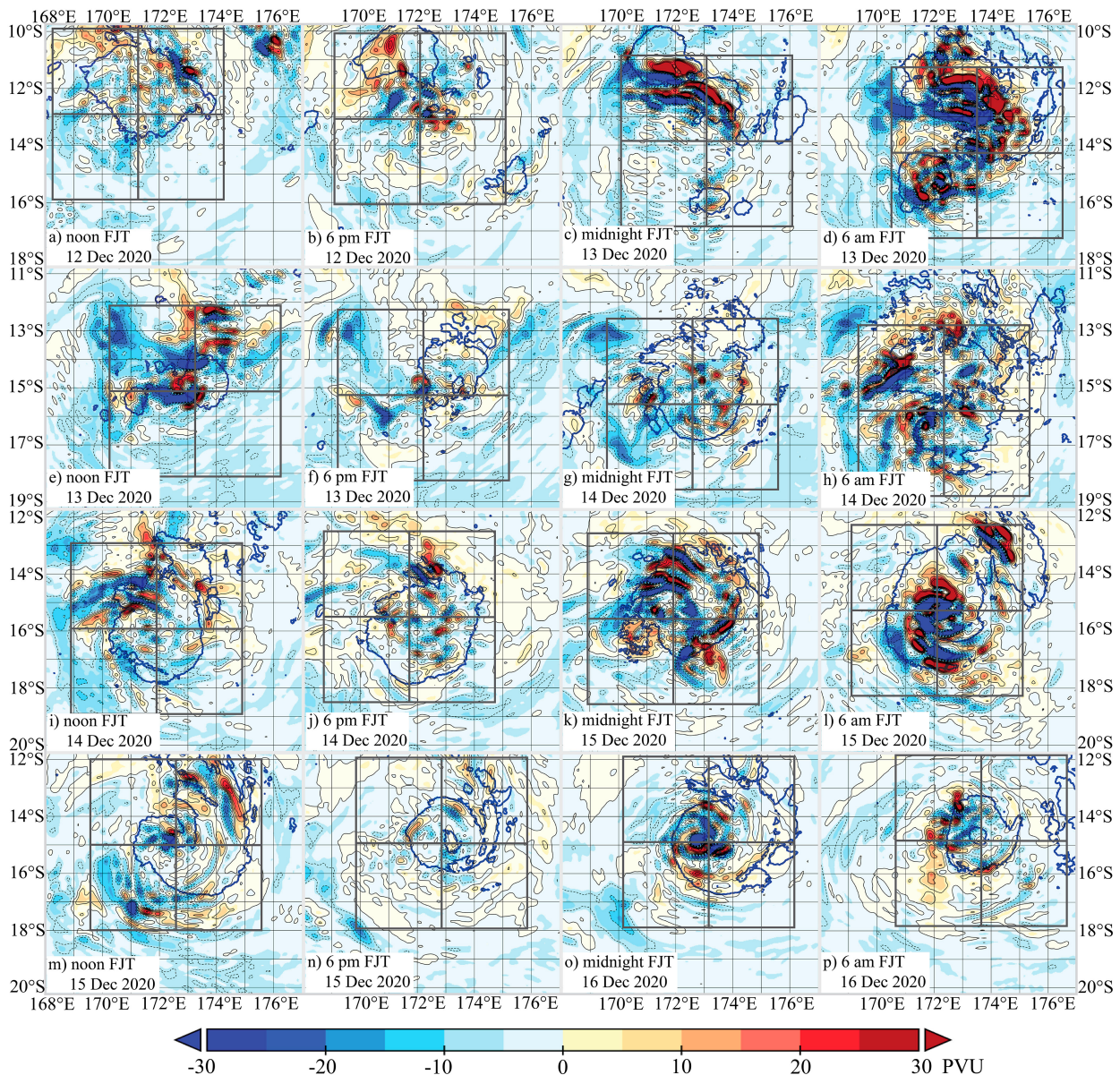


FIG. 3. Six-hourly sequence of 9-km ECMWF 100-hPa PV charts (contours every 5 PVU; negative dashed) and region of cold *Himawari-8* cloud-top temperature (heavy blue 202-K contour) from 1200 FJT 12 Dec to 0600 FJT 16 Dec 2020. (first column) Local noon, (second column) sunset, (third column) midnight, and (fourth column) sunrise. The location of the $6^{\circ} \times 6^{\circ}$ domain used for calculations is shown at each time, with center estimated by IBTrACS. The longitude range for each panel is 168° – 177° E. The latitude range is (a)–(d) 10° – 18° S, (e)–(h) 11° – 19° S, (i)–(l) 12° – 20° S, and (m)–(p) 12° – 20° S.

a t score of -2.7 and p score of 0.05 . The negative t score implies that the peak at 0000 FJT is higher than the peak at 0600 FJT. These results support H3 for TC Yasa, indicating that the preference for a 0600 FJT maximum disappears since it exhibits a sunrise maximum during days 1–3 and a 0000 FJT maximum during days 4–6 (Fig. 2).

Statistically significant results for Yasa include a reduction in the mean amplitude and diurnal cycle range of σ_{PV} by 50% and a shift in the timing of the maximum from 0600 FJT to 0000 FJT after reaching C2.

c. Comparison of ECMWF 100-hPa PV and *Himawari-8* cloud-top temperature charts

A 6-hourly sequence of ECMWF 100-hPa PV charts for the 6-day period 1200 FJT 12 December–0600 FJT 18 December is shown in Figs. 3 and 4. During days 5–6, Yasa's rapid propagation required a shifting field of view (Fig. 4). Mesoscale PV variability is dominated by radial scales of ~ 60 km, with azimuthal scales of ~ 100 – 200 km. Six-hourly charts of ECMWF 100-hPa PV are shown side by side with *Himawari-8* cloud-top

TABLE 2. Tests for TC Yasa of hypotheses H1 (0600 LT minus 1800 LT, for D1–3 and D4–6), H2 (mean of D1–3 minus mean of D4–6), and H3 (0600 LT minus 0000 LT, for D1–3 and D4–6). Columns show the hypothesis, times and days, the two mean values $[X]_1$ and $[X]_2$ of 100-hPa σ_{PV} (PVU), and number of degrees of freedom ν used in the test, with the resulting t score and p value. Time 1 (T1) is noon, T2 is sunset, T3 is midnight, and T4 is sunrise. The key time is 1800 LT day 3 (1800 FJT 15 Dec), with D1–3 and D4–6 indicating the 3 days before and after the key date.

Hypothesis	Times and days	$[X]_1$	$[X]_2$	ν	t score	p value
H1	T4–T2, D1–3	13.0	5.7	4	5.6	0.0025
H1	T4–T2, D4–6	5.4	1.9	4	9.1	0.0004
H2	D1–3 – D4–6	7.7	4.0	20	4.2	0.0004
H3	T4–T3, D1–3	13.3	8.2	4	2.7	0.06
H3	T4–T3, D4–6	4.2	6.5	4	–2.7	0.05

temperature in Figs. S3–S5. Regions of deep convection are enclosed by the 202-K contour in Figs. 3 and 4. The 202-K contour corresponds closely to the 136-hPa cloud-top pressure contour. Within the 202-K contour, updrafts are likely to reach the 100-hPa level.

Regions of enhanced ECMWF 100-hPa PV variability, characterized by numerous high-amplitude PV filaments, occur preferentially in regions where cloud tops are less than 202 K (Figs. 3 and 4). The general spatial agreement between regions of cold *Himawari-8* cloud-top temperature and high 100-hPa mesoscale PV variability suggests that operational ECMWF forecasts provide useful information regarding significant convective features in TCs.

Consistent with the diurnal cycle in 100-hPa σ_{PV} shown in Fig. 2, on each day there is a minimum in PV variability near sunset (second column), with a maximum overnight (third and fourth columns). During the first 3 days, the maxima in PV variability occurred at sunrise (Figs. 3d,h,i). A similar diurnal cycle is seen in the area enclosed by the 202-K contour, which is smallest at sunset (Figs. 3b,f,j,n) and largest at sunrise (Figs. 3d,h,i) during the first 3 days.

This behavior changed on day 4 (fourth row, Fig. 3). A period of RI from C1 to C4 began at 0600 FJT 15 December (Fig. 3l). After reaching a peak in convection near sunrise on 15 December (Fig. 3l), Yasa reached C2 and formed an eye by noon (Fig. 3m), with the area of deep convection contracting and becoming more axisymmetric by sunrise on 16 December (Fig. 3p). In contrast to the previous 3 days, on day 4, PV variability reached a maximum at midnight (Fig. 3o) instead of at sunrise (Fig. 3p). In association with the formation of a compact inner core, areas of outlying convection diminished. By sunrise on 16 December (Fig. 3p), PV variability was reduced relative to the three previous sunrises (Figs. 3d,h,i), both in the surrounding area and in the inner core.

After 0600 FJT 16 December, a diurnal cycle is still evident, with a peak near midnight, but amplitudes are noticeably reduced (cf. Figs. 3 and 4), consistent with H2. This reduction in 100-hPa PV variability coincided with the onset of greater axisymmetry in cloud-top temperature and rapid growth from C2 to C4 (Fig. 2b). Days 4–6 are characterized by a peak in mesoscale PV variability and area of cold cloud tops which is slightly larger near 0000 FJT than at 0600 FJT (Figs. 3o,p and 4c,d,g,h), in contrast with days 1–3, which show a larger peak at 0600 FJT than at 0000 FJT (Figs. 3c,d,i,h,k,l), consistent with H3. By

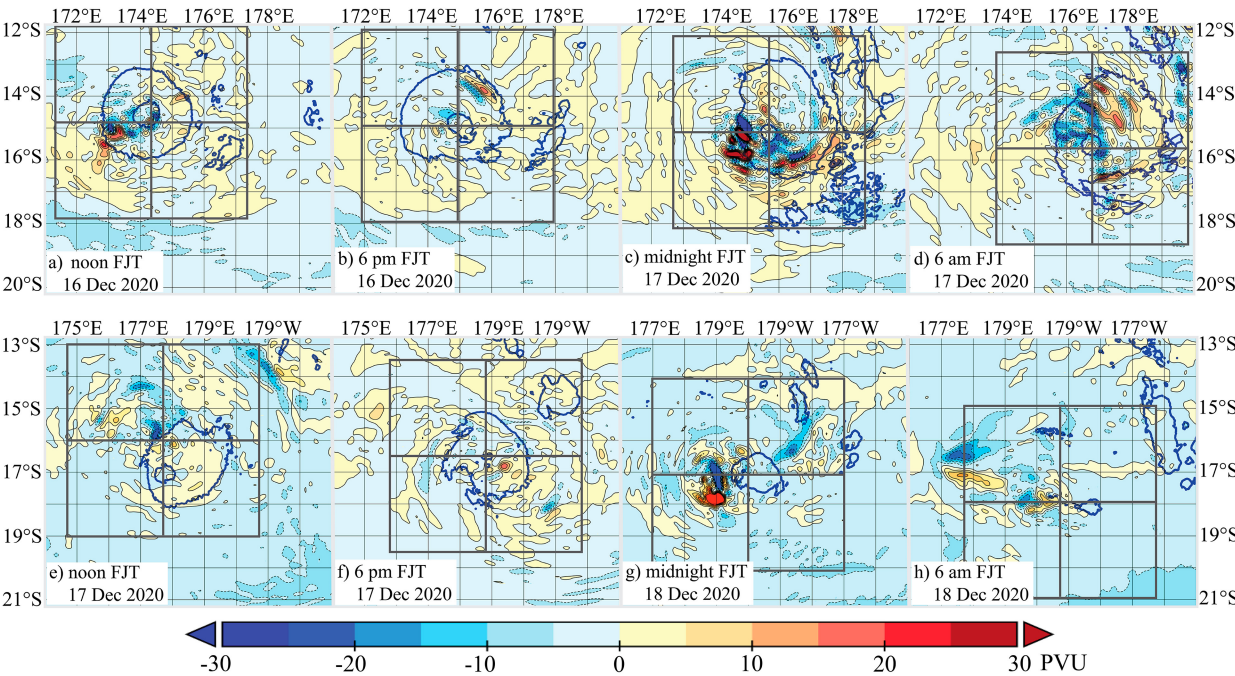


FIG. 4. As in Fig. 3, but for the period 1200 FJT 15 Dec–0600 FJT 18 Dec 2020. (a)–(d) The latitude range is 12°–20°S and the longitude range is 171°E–180°. The latitude range is 13°–21°S and the longitude range is (e),(f) 174°E–179°W and (g),(h) 176°E–175°W.

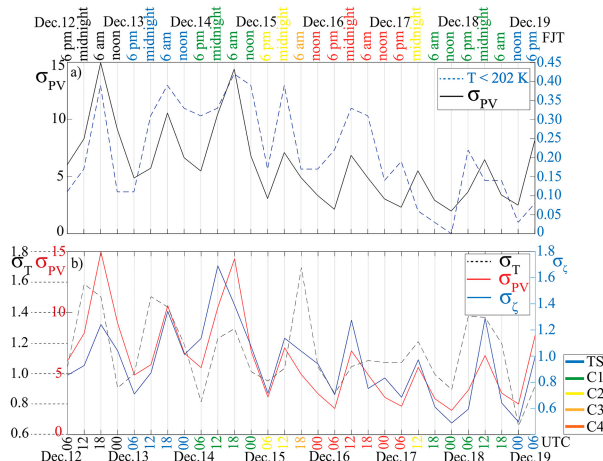


FIG. 5. Comparison of time series of (a) ECMWF 100-hPa σ_{PV} (PVU; solid black) and fractional area in the $6^\circ \times 6^\circ$ domain with *Himawari-8* cloud-top temperature < 202 K (dashed blue) and of (b) ECMWF σ_{PV} (PVU; red), σ_{ζ} ($\times 10^{-4} \text{ s}^{-1}$; blue), and σ_T (K; dashed) at 100 hPa for TC Yasa during 1800 FJT 12 Dec–1800 FJT 19 Dec 2020. The strength category for each time is indicated (color bar legend), with times in black indicating a TD.

00000 FJT 18 December (Fig. 4g), weak PV variability coincided with a reduced area of coldest clouds, and by 0600 FJT 18 December, it was hard to locate the center of the TC (Fig. 4h).

Convective features prior to reaching C2 were loosely organized, while a more symmetric and compact convection pattern appeared afterward (Figs. 3 and 4, and Fig. S2). The reduction in 100-hPa PV variability after reaching C2 suggests a change in dynamical adjustment.

d. Diurnal cycle in 100-hPa σ_{PV} and cold cloud-top area

Time series of ECMWF 100-hPa σ_{PV} and *Himawari-8* cold cloud-top area, calculated as the fractional area in the $6^\circ \times 6^\circ$ domain having brightness temperature $T < 202$ K, are compared in Fig. 5a. These independent estimates both exhibit peaks at 0600 FJT 13 December–0600 FJT 15 December and at 0000 FJT 16 and 17 December. Their relationship is less

coherent during the late decay phase on 18 and 19 December. The correlation coefficient for these two time series is 0.6 for the 7 days 12–19 December. The phase agreement in Fig. 5a provides significant support for interpreting the diurnal cycle in ECMWF σ_{PV} at 100 hPa as a result of model physics producing a diurnal cycle in updrafts which reach the tropopause. It will be shown below that σ_w exhibits a diurnal cycle with peaks which are in phase with those of σ_{PV} and are vertically coherent from the midtroposphere into the UTLS.

In TC Yasa, 100-hPa σ_{PV} , σ_{ζ} , and σ_T each exhibits a diurnal cycle, with a sunset minimum and a peak overnight (Fig. 5b). The correlation coefficient for σ_{PV} and σ_{ζ} is 0.79, for σ_{PV} and σ_T is 0.44, and for σ_{ζ} and σ_T is 0.38. Variations in σ_{PV} are more closely aligned with variations in σ_{ζ} . The term involving σ_{ζ} is expected to contribute much more than the term involving σ_T to σ_{PV} , as shown in the scale analysis of (4).

To understand why there is a maximum in σ_{PV} in the UTLS, we consider the static stability factor, $S = 1/\rho_\theta$, where ρ_θ is isentropic density. Figure 6 shows an estimate of the time mean profile of mean static stability factor $[S]$, following TC Yasa, using three different methods. The first estimate was obtained by averaging the ratio of mean PV to mean absolute vorticity in (3) over the full domain: $[1/\rho_\theta] = [(1/\rho_\theta)(\partial\theta/\partial z)] = [-g(\partial\theta/\partial p)] = [PV]/[f + \zeta]$. The mean static stability parameter values at 100 and 200 hPa in Fig. 6 are $[S] \sim 0.063$ and $0.006 \text{ K m}^2 \text{ kg}^{-1}$, with mean values for isentropic density of $[\rho_\theta] \sim 16$ and $161 \text{ kg m}^{-2} \text{ K}^{-1}$.

A second estimated profile of $[S]$ was obtained by averaging the ratio of mean σ_{PV} to σ_{ζ} : $[S] = [\sigma_{PV}]/[\sigma_{\zeta}]$. At 100 hPa, $[\sigma_{PV}]/[\sigma_{\zeta}] \sim 0.063$, which agrees with the first method (Fig. 6). Agreement between these two different estimates of $[S]$ verifies that variations in ECMWF-calculated PV at 100 hPa are due primarily to variations in ζ , multiplied by mean static stability.

The third estimate was obtained by using the mean temperature profile and $\theta = T(1000/p)^{R/c_p}$ to calculate $[-g(\partial\theta/\partial p)]$ using centered differences on the coarse vertical grid of the archived data. Although this profile generally agrees with the other two estimates, the 300–400-hPa layer exhibits static instability (Fig. 6). This is likely due to error arising from coarse resolution, e.g., $\Delta z = -\Delta p/(\rho g) \sim 4 \text{ km}$ at 100 hPa,

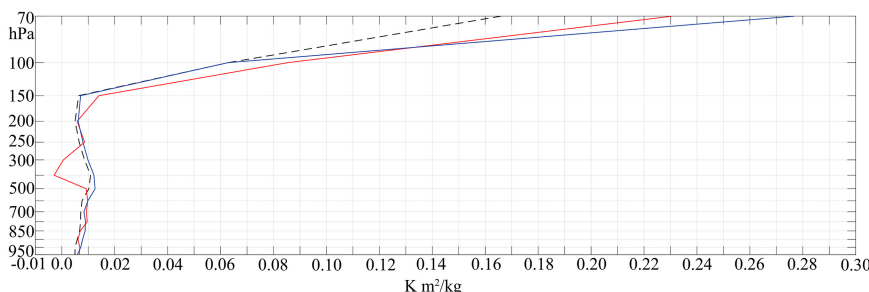


FIG. 6. The time mean vertical profile of the static stability factor, S , estimated by three different methods: from the ratio $[PV]/[f + \zeta]$ (solid blue), from the ratio $[\sigma_{PV}]/[\sigma_{\zeta}]$ (dashed black), and from finite difference calculations on the coarse vertical grid using $[-g(\partial\theta/\partial p)]$ (solid red) for the full $6^\circ \times 6^\circ$ domain following the track of TC Yasa during the 6-day period 0000 FJT 12 Dec–1800 FJT 17 Dec 2020.

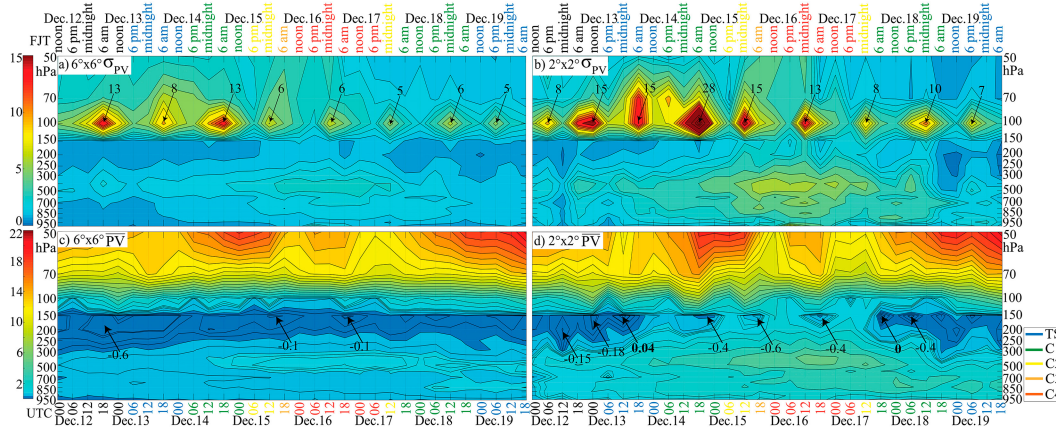


FIG. 7. Time–altitude sections of (a),(b) σ_{PV} and (c),(d) $\overline{PV}$ following the evolution of TC Yasa for (left) the full $6^\circ \times 6^\circ$ domain and (right) the $2^\circ \times 2^\circ$ inner domain. In (a) and (b), σ_{PV} contours are shown every 0.5 PVU from 0 to 3, every 1 PVU from 3 to 15, and every 2 PVU from 22 to 30. In (c) and (d), $\overline{PV}$ contours are shown every 0.02 PVU from 0.04 to -0.1 , every 0.1 PVU from -0.1 to -0.4 , every 0.2 PVU from -0.4 to -1.0 , every 1 PVU from -1 to -3 , and every 2 PVU from -3 to -21 . Selected maxima and minima are labeled. Times are color coded by storm strength (inset legend).

compared to ~ 300 -m resolution of the ECMWF model grid. The temperature profile in the 300–400-hPa layer is close to adiabatic, so that coarse vertical resolution can be a source of error. This profile with negative $[S]$ near 400 hPa (Fig. 6) may indicate that this region is often unstable, as found by Downs and Kieu (2020). Further stability analysis including moisture would be required to investigate this fully. The first estimate, $\overline{S} = [\overline{PV}/(f + \zeta)]$, is probably the best estimate of the domain-averaged static stability profile.

The mean static stability factor is ~ 10 times larger near 100 hPa than at 200 hPa in TC Yasa (Fig. 6). It is thus expected that ECMWF σ_{PV} will exhibit a sharp maximum at the base of the stratosphere relative to σ_ζ , due to the abrupt increase in static stability and finite penetration of evanescent mesoscale vorticity anomalies into the lower stratosphere. Near 100 hPa, with $\rho_\theta \sim 16 \text{ kg m}^{-2} \text{ K}^{-1}$ and $\sigma_\zeta \sim 1 \times 10^{-4} \text{ s}^{-1}$, $\sigma_{PV} \sim \pm 6 \text{ PVU}$, consistent with average values in Fig. 5b. This is also consistent with the relation $\sigma_{PV} \approx \overline{S}\sigma_\zeta$ obtained from scale analysis of (4).

e. Time–height evolution of dynamical variables

The evolution of σ_{PV} and $\overline{PV}$ is shown for the $6^\circ \times 6^\circ$ and $2^\circ \times 2^\circ$ domains in Fig. 7. A strong maximum in σ_{PV} occurs at 100 hPa for Yasa, with a minimum in the 150–200-hPa layer, in both the full and inner domains (Figs. 7a,b). This sharp maximum in σ_{PV} at 100 hPa is due to the increase in static stability at the tropopause (Fig. 6). Maximum values of the diurnal cycle in σ_{PV} at 100 hPa are ~ 5 – 15 PVU in the full domain and ~ 10 – 30 PVU in the inner domain (Figs. 7a,b). Each domain exhibits a reduction in amplitude by $\sim 50\%$ after reaching C2 (Figs. 7a,b).

Relative maxima in σ_{PV} are also seen in the midtroposphere (Figs. 7a,b), where σ_{PV} grows and decays in proportion to storm intensity, as indicated by $\overline{PV}$ and strength category

(Figs. 7c,d). At 500 hPa, σ_{PV} values reach ~ 2 – 3 PVU in the full domain and ~ 3 – 6 PVU in the inner domain.

A cyclonic maximum (negative $\overline{PV}$) is found in the midtroposphere, with increasing values spreading both upward and downward from $\sim 500 \text{ hPa}$ as Yasa strengthened through 17 December (Figs. 7c,d). One may distinguish two midtropospheric $\overline{PV}$ maxima, one near 450 hPa and one near 800 hPa (Figs. 7c,d), somewhat similar to TC Lan (Tsujino et al. 2021). While Yasa was C2–C4 (15–17 December), $\overline{PV}_{450\text{hPa}} \sim -1 \text{ PVU}$ in the $6^\circ \times 6^\circ$ domain and $\sim -4 \text{ PVU}$ in the $2^\circ \times 2^\circ$ domain (Figs. 7c,d). These values for the symmetric vortex are similar in magnitude to those diagnosed for TC Danny by Molinari et al. (1998).

A layer of very small negative $\overline{PV}$ is observed in the upper troposphere prior to reaching C2, with $\overline{PV} > -0.1$ in the 200–150-hPa layer in the $6^\circ \times 6^\circ$ domain (Fig. 7c). During amplification, the upward extension of cyclonic $\overline{PV}$ encroaches on this layer of small $\overline{PV}$ (Figs. 7c,d). Nevertheless, during 13–18 December, there is a diurnal cycle in 150 hPa $\overline{PV}$, with the smallest values occurring overnight, being even slightly positive near 0600 FJT 13 December and 0600 FJT 18 December in the $2^\circ \times 2^\circ$ domain (Figs. 7c,d). Consistent with reduced static stability and reduced inertial stability, these minima in $\overline{PV}$ are compatible with the hypothesis that cloud-radiative forcing destabilizes the upper troposphere and increases anticyclonic outflow in the UTLS overnight.

To explore relationships between dynamical variables, time–height sections of σ_ζ , $\sigma_{V \cdot V}$, and σ_ω for the $6^\circ \times 6^\circ$ domain, together with 100-hPa time series, are shown in Fig. 8. The diurnal cycle in σ_{PV} (Figs. 5a and 7a) is strongly related to the diurnal cycle in σ_ζ (Figs. 5b and 8a,b). Note how the diurnal cycle in σ_ζ diminishes in amplitude into the stratosphere (Fig. 8a). In comparison, σ_{PV} exhibits an abrupt upward increase at 100 hPa, with values ~ 10 times those at 200 hPa (Fig. 7a), due to the increase in the static stability factor (Fig. 8).

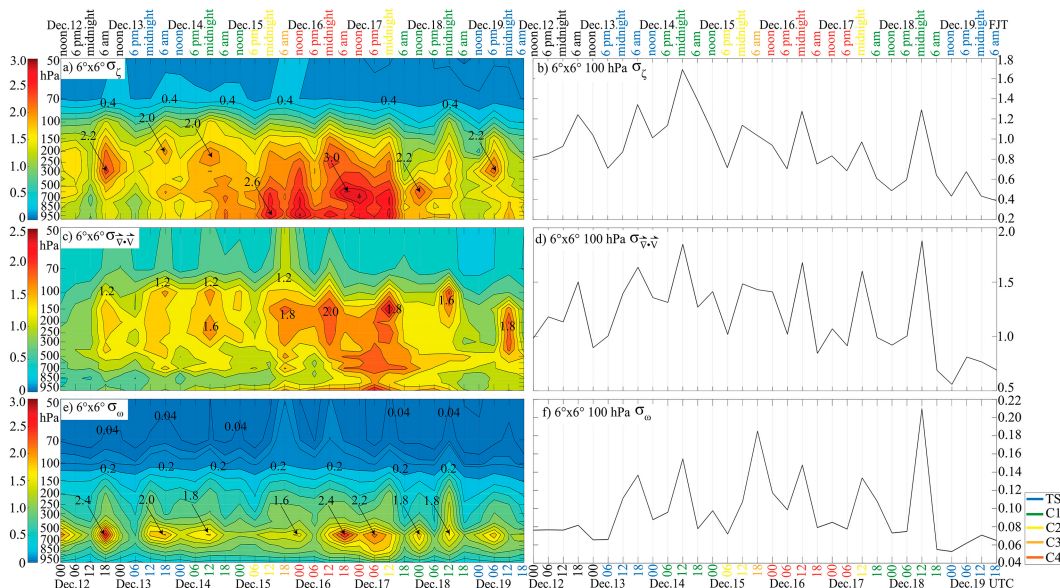


FIG. 8. Time–altitude sections following the evolution of TC Yasa for the $6^\circ \times 6^\circ$ domain for (a) σ_z (10^{-4} s^{-1}), (c) $\sigma_{\mathbf{v} \cdot \mathbf{v}}$ (10^{-4} s^{-1}), and (e) σ_ω (Pa s^{-1}), with contours every 0.02 from 0.04 to 0.08 and every 0.2 from 0.2 to 3.0. Time series at 100 hPa for (b) σ_z , (d) $\sigma_{\mathbf{v} \cdot \mathbf{v}}$, and (f) σ_ω .

The time mean profile of the ratio $\sigma_{\text{PV}}/\sigma_z$ in Fig. 6 is nearly equal to the profile of the static stability factor.

The time series of σ_{PV} and σ_z at 100 hPa have a correlation coefficient of 0.79. This is compatible with scale analysis of (4), which suggested that σ_{PV} can be primarily characterized by eddy variations in relative vorticity, multiplied by the mean static stability profile.

During the first 3 days (13–15 December), a 0600 FJT maxima in σ_z occurred in the 150–300-hPa layer (Fig. 8a). During 16–17 December, this regime with an elevated upper-tropospheric maximum in σ_z is overwhelmed by lower-tropospheric maxima which extend upward from below 500 hPa (Fig. 8a). This evolution is consistent with cloud-radiative theories of overnight destabilization of the upper troposphere, coupling with overall amplification of deep convection in the midtroposphere in proportion to storm strength.

There is a diurnal variation in σ_ω centered in the midtroposphere, with maxima extending into the UTLS overnight (Figs. 8e,f), consistent with the convective origin of 100-hPa σ_{PV} features. A similar diurnal variation is seen in $\sigma_{\mathbf{v} \cdot \mathbf{v}}$, which is intimately involved in the response to deep convective updrafts, the formation of mesoscale PV structures, and the radiation of gravity waves in the adjustment process. A diurnal maximum in $\sigma_{\mathbf{v} \cdot \mathbf{v}}$ occurs at 0000–0600 FJT in the mid- and upper troposphere (Figs. 8c,d), which extends into the UTLS and covaries with σ_{PV} at 100 hPa (Figs. 5 and 7a).

The time series of σ_z and $\sigma_{\mathbf{v} \cdot \mathbf{v}}$ at 100 hPa (Figs. 8b,d) have a correlation coefficient of 0.87, which is consistent with deep convective forcing of relative vorticity anomalies in the UTLS. Diurnal maxima in σ_{PV} are related to maxima in divergence variability and the production of relative vorticity anomalies in the UTLS. This gives rise to large PV anomalies at the tropopause, where the basic-state static stability

abruptly increases. The diurnal cycle in σ_{PV} at 100 hPa may be interpreted as the result of updrafts which penetrate the 100-hPa level more frequently overnight, due to enhanced cloud-radiative forcing and decreased convective stability in the upper troposphere.

The evolution of σ_{PV} is different in the midtroposphere, where the amplitude varies in proportion to storm strength. At 500 hPa, $\overline{\text{PV}}$ in the inner core tracks closely with storm intensity, reaching -5 PVU while C4 (Fig. 7d). At 500 hPa, σ_{PV} (Fig. 7a) and σ_z (Fig. 8a) vary closely with $\overline{\text{PV}}$ (Fig. 7c). The peak in σ_{PV} is ~ 2.8 PVU, and the peak in σ_z is $\sim 2.9 \times 10^{-4} \text{ s}^{-1}$. The average strength of relative vorticity anomalies exceeds the Coriolis parameter, consistent with $\text{Ro} > 1$ and a system undergoing dynamical adjustment. At 500 hPa, the correlation coefficient between σ_{PV} and σ_z is 0.96. At 500 hPa, the correlation between σ_{PV} and $\overline{\text{PV}}$ is -0.58 , which indicates that a stronger cyclonic vortex (more negative $\overline{\text{PV}}$) is associated with enhanced PV variability in the midtroposphere.

4. Climatology of UTLS mesoscale PV variability in South Pacific TCs

a. Storm tracks and time series of ECMWF σ_{PV} near the tropopause

Individual storm tracks illustrate the range of locations for the 13 C1–C2 TCs (Fig. 9). A diurnal cycle in 100-hPa σ_{PV} is evident for the first 6 days in most of the TCs (Fig. 10). The amplitude of the diurnal cycle tends to decrease after the key time, but not in all C1–C2 TCs (Fig. 10). Almost all of the C1–C2 TCs exhibited a maximum near sunrise for days 1–3. There is more variability in the timing of the maximum among the TCs during days 4–6, with many occurring at midnight (Fig. 10).

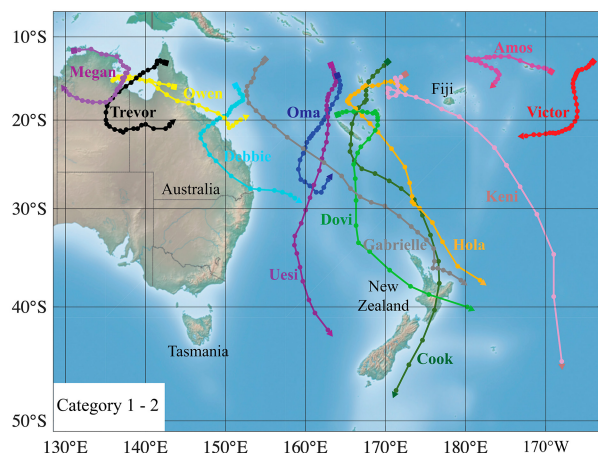


FIG. 9. IBTrACS storm tracks for 13 TCs which satisfied our selection criteria and reached C1–C2 in the South Pacific during 2016–24.

The range in position of storm tracks for the 10 TCs which reached C3–C5 is more compact than the C1–C2 group (Figs. 9 and 11) and more centered over the warm pool. A diurnal cycle is also evident for the eight C3–C5 TCs with

maximum amplitudes at 100 hPa (Fig. 12a). After the key date, the amplitude of the diurnal cycle decreases, and there is more variation in the timing of the maximum (Fig. 12a). Almost all of the C3–C5 TCs which maximized near 100 hPa exhibited a 0600 LT peak for days 1–3 (Fig. 12a). During days 4–6, there is more variability in the timing of the maximum, with many C3–C5 TCs exhibiting a midnight maximum (Fig. 12a).

For TCs Winston and Kevin, the signal at 70 hPa dominated (Figs. S6a,b), so their 70-hPa σ_{PV} time series are shown separately in Fig. 12b. These two very strong TCs exhibited irregularity in the timing of diurnal maxima (Fig. 12b and Figs. S6a,b). However, there is a reduction in amplitude after the key date (Fig. 12b). Note also the gradual growth and decay of the mid-tropospheric maximum in σ_{PV} associated with storm strength (Figs. S6a,b), similar to TC Yasa (Fig. 7a). TC Harold was also a very strong storm and exhibited a complex life cycle, with periods of redevelopment (Fig. S6c).

b. Climatology of the diurnal cycle in tropopause-level σ_{PV}

Values of tropopause-level ECMWF σ_{PV} for all 7 days for 20 TCs, binned by time of day, together with the mean for each time bin, are shown for groups C1–C2, C3–C5, and C1–C5 in

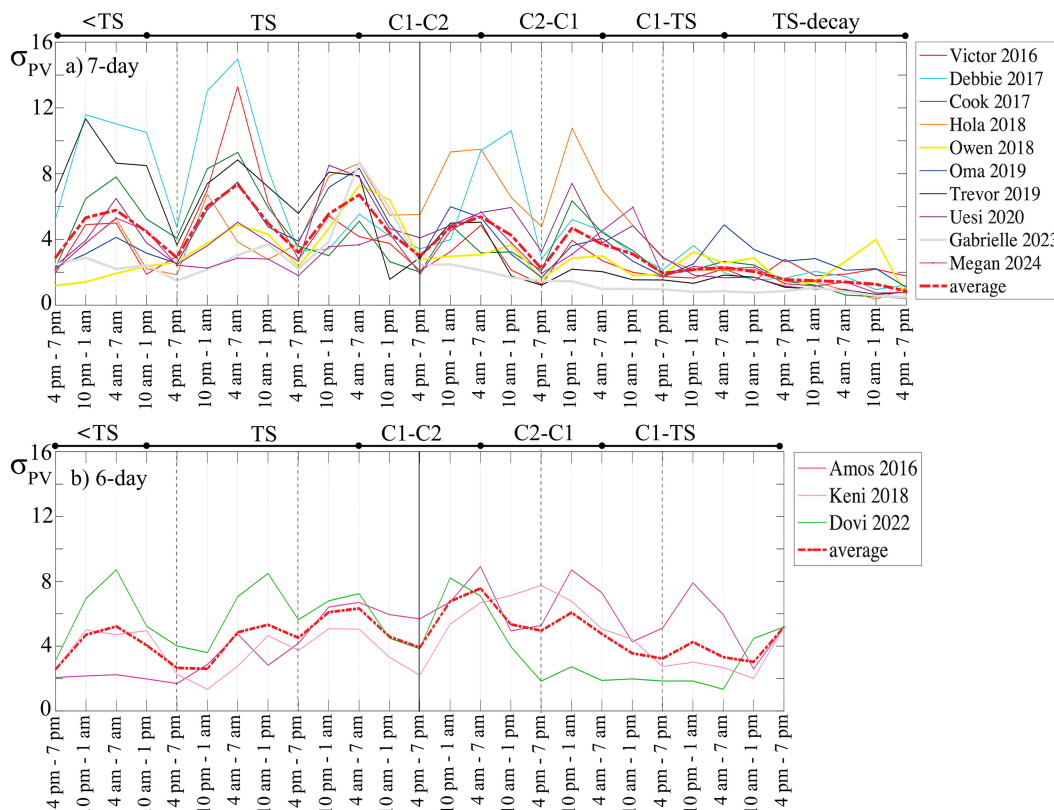


FIG. 10. Time series of ECMWF 100-hPa σ_{PV} for (a) each of the 10 TCs which reached C1–C2 and lasted 7 days (legend inset), with their average (dashed), and (b) each of the three TCs which reached C1–C2 and lasted 6 days (legend inset), with their average (dashed). Sunset (1600–1900 LT) is indicated for each day by a thin dashed vertical line. The key time is highlighted by a solid vertical line. Storm strength is shown as a function of time at the top.

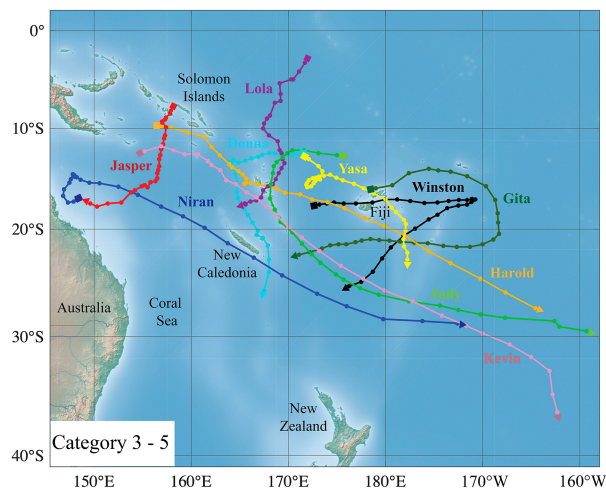


FIG. 11. IBTrACS storm tracks for 10 TCs which satisfied our selection criteria and reached C3–C5 in the South Pacific during 2016–24.

Fig. 13. In each group, midnight and sunrise values exceed sunset values, and there is a moderate preference for sunrise over midnight, which is more evident in the C3–C5 group.

To test H1, we asked whether the mean at 0600 LT (T4) is significantly higher than the mean at 1800 LT (T2) and analyzed all 20 TCs that had a full 7-day record (Figs. 10a and 12). The p value of 10^{-7} in Table 3 suggests that the diurnal cycle observed in these 23 South Pacific TCs is highly statistically significant.

We explored H2 and H3 by using the first 6 days from all 23 TCs. Figure 14 and Table 3 show that there is a reduction in amplitude from days 1–3 to days 4–6 and a preference for a sunrise maximum during days 1–3, but not during days 4–6.

To test H2, we first asked whether the mean value of σ_{PV} for days 1–3 is significantly larger than the mean value for days 4–6. For the C1–C2 group (Fig. 14a), the p value is $\sim 10^{-6}$ (Table 3). These reductions in mean amplitude of 23% for the C1–C2 group of (Fig. 14a) and 24% for the C3–C5 group (Fig. 14b) are highly significant, with p values $\sim 10^{-6}$ (Table 3).

For H2, we also asked whether the diurnal range is reduced, where $[X]_1$ equals the mean for period T4 (0600 FJT) minus the mean for period T2 (1800 FJT) for days 1–3 and $[X]_2$ is the mean for T4 minus the mean for T2 for days 4–6. For the C1–C2 group (Fig. 14a), the p value of $\sim 10^{-5}$ (Table 3) shows that there is a reduction in the range of the diurnal cycle of $\sim 45\%$ which is highly significant. For the C3–C5 group (Fig. 14b), with a p value of $\sim 10^{-7}$ (Table 3), there is a larger, more significant reduction in the range of the diurnal cycle, by $\sim 60\%$.

To test H3, we asked whether the mean value of σ_{PV} for T4 (0600 FJT) is significantly higher than the mean for period T3 (0000 FJT). For the C1–C2 group and days 1–3 (Fig. 14a), the t score is 1.9, with a p value of 0.025 (Table 3). This is significant at the 97.5% level and thus supports H3 that there is a preference for a 0600 FJT maximum during days 1–3 for C1–C2 storms. For the C1–C2 group and days 4–6, the t score is 0.5, with a p value of 0.3 (Table 3). This indicates that there is no statistically significant difference between 0000 and 0600 FJT for a maximum during days 4–6 for C1–C2 storms. This

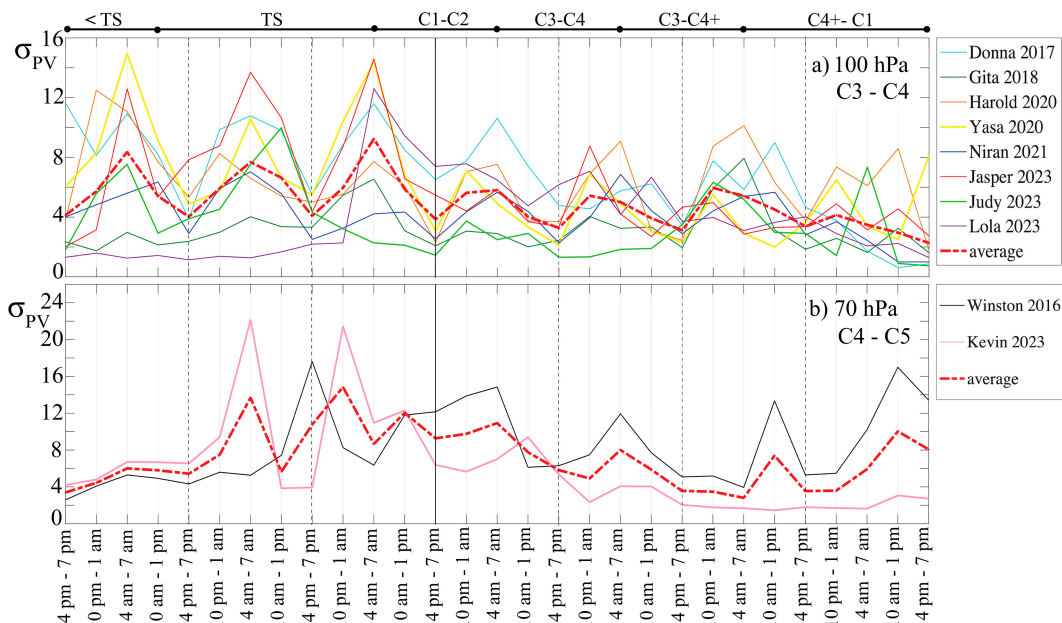


FIG. 12. Time series of ECMWF σ_{PV} for TCs which reached C3–C5 for (a) the eight TCs which maximized at 100 hPa (legend inset) and (b) the two TCs which maximized at 70 hPa (legend inset). Sunset (1600–1900 LT) is indicated for each day by a thin dashed vertical line. The key time is highlighted by a solid vertical line. Storm strength is shown as a function of time at the top.

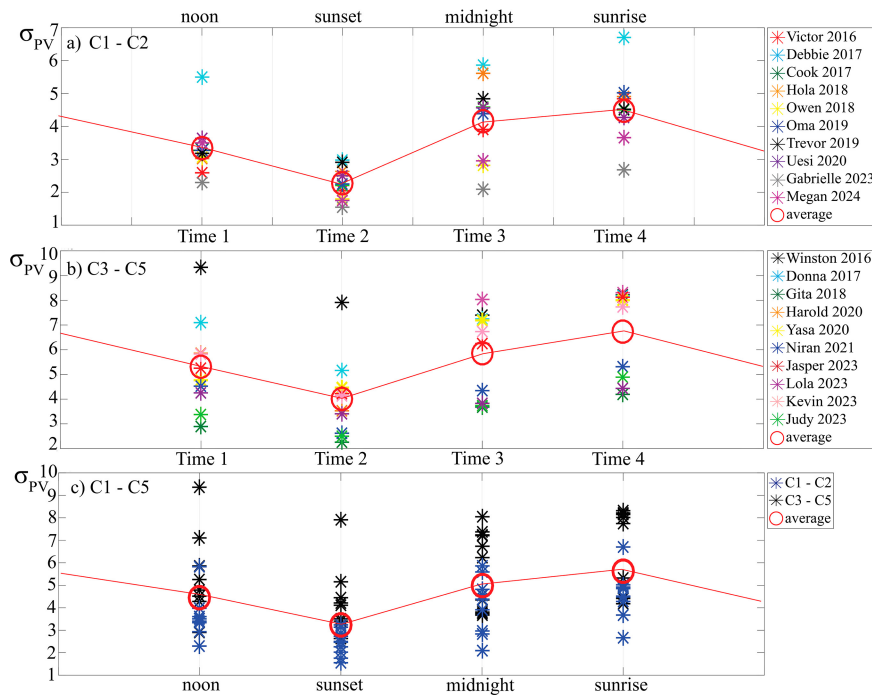


FIG. 13. Values of ECMWF tropopause-level σ_{PV} binned by time of day for 1000–1300 FJT (noon), 1600–1900 FJT (sunset), 2200–0100 FJT (midnight), and 0400–0700 FJT (sunrise), for all 7 days, for (a) the 10 TCs which reached C1–C2 (asterisks; legend inset), b) the 10 TCs which reached C3–C5 (asterisks; legend inset), and (c) all TCs, C1–C5 (C1–C2 indicated by a blue asterisk; C3–C5 by a black asterisk). The average value in each time bin is shown as an open circle, connected by solid line segments. Time 1 is noon (0000 UTC), time 2 is sunset (0600 UTC), time 3 is midnight (1200 UTC), and time 4 is sunrise (1800 UTC).

also supports H3, which predicts a reduced preference for 0600 FJT.

For the C3–C5 group and days 1–3 (Fig. 14b), the p value is 10^{-7} (Table 3), showing a highly statistically significant preference for a 0600 FJT maximum. For the C3–C5 group and days 4–6, the t score is -0.007 , with a p value of 0.5 (Table 2), showing an insignificant preference for a midnight maximum. These results are consistent with H3, which predicts a preference for a sunrise maximum which goes away in the mature stage.

Statistically significant results include a reduction in mean amplitude from days 1–3 to days 4–6 of $\sim 25\%$ for both C1–C2 and

C3–C5 TCs and a reduction in the diurnal range of $\sim 45\%$ for C1–C2 and of $\sim 60\%$ for C3–C5 TCs. The preference for a 0600 FJT maximum during days 1–3 goes away during days 4–6.

5. Discussion

a. Evolution of the diurnal cycle in 100-hPa σ_{PV}

Choosing a key time of 0600 LT within 24 h after reaching C1, and requiring at least a 6-day record, provides the opportunity to study the life cycle of TCs, as manifested in the

TABLE 3. Results of statistical tests for hypotheses H1–H3, showing the times of day and days used, group (C1–C2, C3–C5, C1–C5), two mean values of 100-hPa σ_{PV} (PVU) used in the test, number of independent samples, t score, and p value. Time 1 (T1) is noon, T2 is sunset, T3 is midnight, and T4 is sunrise. The key time is 1800 LT day 3, with D4–6 indicating the 3 days after reaching C1.

Hyp.	Times	Group	$[X]_1$	$[X]_2$	N	t score	p value
H1	T4–T2, D1–7	C1–C5	5.6	3.4	140	9.2	$\sim 10^{-7}$
H2	Mean, D1–3 – D4–6	C1–C2	4.8	3.7	156	4.2	$\sim 10^{-6}$
H2	Mean, D1–3 – D4–6	C3–C5	6.5	5.1	156	4.4	$\sim 10^{-6}$
H2	T4–T2, D1–3 – D4–6	C1–C2	2.9	1.6	78	3.6	$\sim 10^{-5}$
H2	T4–T2, D1–3 – D4–6	C3–C5	4.2	1.7	60	5.5	$\sim 10^{-7}$
H3	T4–T3, D1–3	C1–C2	6.0	5.3	78	1.9	0.025
H3	T4–T3, D4–6	C1–C2	4.4	4.2	78	0.5	0.3
H3	T4–T3, D1–3	C3–C5	8.8	6.5	60	6.2	$\sim 10^{-7}$
H3	T4–T3, D4–6	C3–C5	5.7	5.8	60	-0.007	0.5

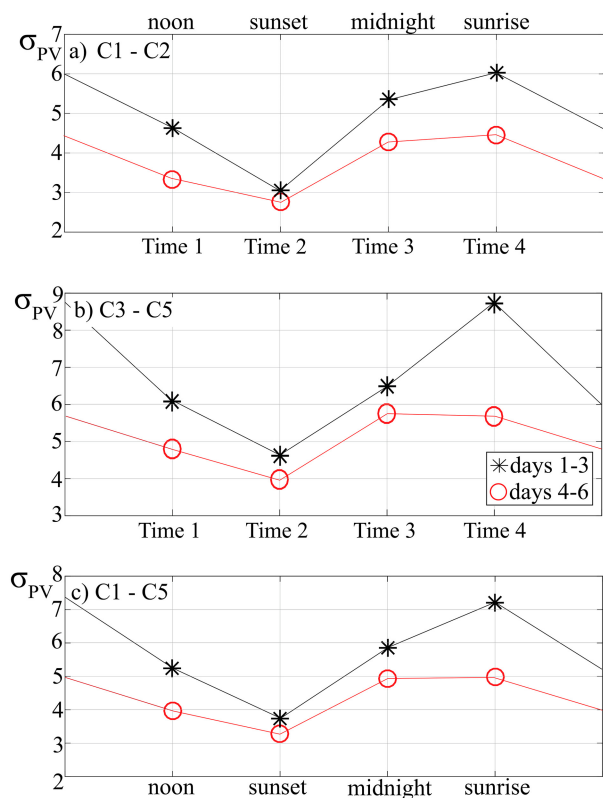


FIG. 14. Values of tropopause-level σ_{PV} binned by time of day for 1000–0100 FJT (noon), 1600–1900 FJT (sunset), 2200–0100 FJT (midnight), and 0400–0700 FJT (sunrise), averaged separately for days 1–3 (*) and days 4–6 (o), using 6-day records, for (a) the 13 storms that reached C1–C2, (b) the 10 storms that reached C3–C5, and (c) all 23 storms, C1–C5. Time 1 is noon (0000 UTC), time 2 is sunset (0600 UTC), time 3 is midnight (1200 UTC), and time 4 is sunrise (1800 UTC). The period between sunrise and noon is shown at both ends for visual continuity.

diurnal cycle in tropopause-level σ_{PV} , including the development, maturation, and decay phases. During development (days 1–3), South Pacific TCs exhibit a strong diurnal cycle with a preference for a peak at 0600 LT. During the mature phase (days 4–6), there is a reduction in amplitude of $\sim 45\%$ for C1–C2 and $\sim 60\%$ for C3–C5, with a reduced 0600 LT preference. This reduction in amplitude of the diurnal cycle in 100-hPa σ_{PV} is consistent with reduced dynamical adjustment after achieving a symmetric core in the mature phase. An incoherent diurnal cycle in 100-hPa σ_{PV} is observed when a TC has lost its core circulation and is decaying, which occurs on day 7 in many of the TCs. These three distinct qualitative phases in the diurnal cycle in 100-hPa σ_{PV} may be useful information for considering the nature of dynamical adjustment in TCs.

b. Comparison with cloud-radiative forcing of the diurnal cycle in TCs

The changes observed in 100-hPa σ_{PV} from days 1–3 to days 4–6 are consistent with expectations from cloud-radiative forcing theories and hypotheses H2 and H3. The larger range

and preference for a 0600 LT peak during days 1–3 are consistent with the dynamical cloudy-clear mechanism during the development phase, while the reduced range and loss of 0600 LT preference during days 4–6 are consistent with the lapse rate mechanism and a uniform central vortex during the mature phase. The early morning minima in $\overline{PV}$ in the 150–200-hPa layer for TC Yasa are also compatible with cloud-radiative forcing, which destabilizes the upper troposphere in TCs overnight. The extra component of clear/cloudy heating contrast during days 1–3 may be important for achieving a critical burst of convection, which creates a strong, axisymmetric vortex through dynamical adjustment toward balance (Smith et al. 2020).

6. Conclusions

A statistical method for characterizing PV variability in TCs in 9-km-resolution ECMWF forecast analyses was described, where the spatial mean and standard deviation of quantities are calculated at each altitude for selected domains moving with the storm. It was shown that the amplitude maximum in σ_{PV} at the tropopause is caused by the convolution of the profile of σ_g , which decreases into the stratosphere, with the profile of the mean static stability factor, which increases abruptly at the tropopause.

We investigated the evolution and diurnal cycle of mean PV and mesoscale variability in PV and other dynamical variables in 9-km ECMWF forecasts for 23 TCs in the South Pacific basin during 2016–24. A detailed comparison of the distribution of mesoscale PV structures with *Himawari-8* satellite cloud-top temperatures in TC Yasa showed that they occur where cold temperatures indicate the presence of deep convection and that the diurnal cycle of 100-hPa σ_{PV} is in phase with the cold cloud-top fractional area < 202 K. Statistical analysis of 13 C1–C2 TCs and 10 C3–C5 TCs shows that there is a 45%–60% reduction in the range of the diurnal cycle in 100-hPa σ_{PV} and a reduced preference for a 0600 LT maximum after becoming a TC. These results are consistent with expectations from cloud-radiative forcing theories.

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Data availability statement. The 0.08°-resolution ECMWF IFS CY41r2 High-Resolution Operational Forecasts are maintained at the Research Data Archive at the National Center for Atmospheric Research, Computational and Information Systems Laboratory (<https://doi.org/10.5065/D68050ZV>). Fields of *Himawari-8* cloud-top temperature at 2-km resolution were obtained from National Oceanic and Atmospheric Administration at <https://noaa-himawari8.s3.amazonaws.com/index.html>. Maps of SST were obtained from NOAA at <https://www.ncei.noaa.gov/>. Storm track data were obtained from the Cooperative Institute for Meteorological Satellite Studies at <https://tropic.ssec.wisc.edu/misc/adt/info.html>, the Joint Typhoon Warning Center at <https://www.metoc.navy.mil/jtwc/jtwc.html?southern-hemisphere>, and the International Best Track

Archive for Climate Stewardship, available at <https://www.ncei.noaa.gov/products/international-best-track-archive>. Strength and duration values were cross-checked with the NOAA Historical Hurricane Track database at <https://oceanservice.noaa.gov/news/historical-hurricanes/>. Images in this manuscript were produced using PANOPLY from the National Aeronautics and Space Administration Goddard Institute for Space Studies, available at <https://www.giss.nasa.gov/tools/panoply/>, MATLAB, and Adobe Illustrator, available at <https://www.adobe.com>. MATLAB version R2022b (MATLAB 2022), available at <https://www.mathworks.com/>, was used to calculate means and standard deviations of gridded data using the locations and times from IBTrACS.

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