Influence of the Subtropical High and Storm Track on Low-Cloud Fraction and Its Seasonality over the South Indian Ocean

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(Manuscript received 7 April 2017, in final form 16 February 2018)

ABSTRACT

The south Indian Ocean is characterized by enhanced midlatitude storm-track activity around a prominent sea surface temperature (SST) front and unique seasonality of the surface subtropical Mascarene high. The present study investigates the climatological distribution of low-cloud fraction (LCF) and its seasonality by using satellite data, in order to elucidate the role of the storm-track activity and subtropical high. On the equatorward flank of the SST front, summertime LCF is locally maximized despite small estimated inversion strength (EIS) and high SST. This is attributable to locally augmented sensible heat flux (SHF) from the ocean under the enhanced storm-track activity, which gives rise to strong instantaneous wind speed while acting to relax the meridional gradient of surface air temperature. In the subtropics, summertime LCF is maximized off the west coast of Australia, while wintertime LCF is distributed more zonally across the basin unlike in other subtropical ocean basins. Although its zonally extended distribution is correspondent with that of LCF, EIS alone cannot explain the wintertime LCF enhancement, which precedes the EIS maximum under continuous lowering of SST and enhanced SHF in winter. Basinwide cold advection associated with the wintertime westward shift of the subtropical high contributes to the enhancement of SHF, especially around 15°–25°S, while seasonally enhanced storm-track activity augments SHF around 30°S. The analysis highlights the significance of large-scale controls, particularly through SHF, on the seasonality of the climatological LCF distribution over the south Indian Ocean, which reflect the seasonality of the Mascarene high and storm-track activity.

1. Introduction

Low-level clouds strongly cool the earth as they reflect a large fraction of insolation while emitting longwave

Denotes content that is immediately available upon publication as open access.

Supplemental information related to this paper is available at the Journals Online website: <https://doi.org/10.1175/JCLI-D-17-0229.s1>.

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radiation nearly as much as the earth's surface (Hartmann and Short 1980). In fact, a small change in fractional coverage of low-level clouds could offset anthropogenic global warming (Randall et al. 1984). Since their formation is governed by small-scale turbulent processes, however, low-level clouds are not well represented in global climate models, introducing uncertainties into future climate projections (e.g., Bony and Dufresne 2005; Qu et al. 2014, 2015; Myers and Norris 2016). Thus, understanding how the properties of low-level clouds are determined is a fundamental issue in climate science.

Climatologically, low-level clouds are frequently observed over cool oceans, where deep convection is unlikely to occur. For example, stratocumulus clouds

DOI: 10.1175/JCLI-D-17-0229.1

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prevail over the eastern portion of each of the subtropical ocean basins (e.g., Klein and Hartmann 1993; Wood 2012), which is located east of a surface subtropical high that accompanies persistent midtropospheric subsidence and equatorward surface winds (e.g., Miyasaka and Nakamura 2005, 2010). The equatorward winds induce coastal upwelling and upper-ocean mixing in addition to surface evaporation, acting to maintain relatively low sea surface temperature (SST). Owing to their high albedo, those stratocumulus clouds also act to cool the ocean. Combination of the low SST and enhanced midtropospheric subsidence, which acts to warm the free troposphere, maintains a strong temperature inversion at the top of the boundary layer, inhibiting cloud-top entrainment of dry air. In fact, the temperature inversion can explain seasonality of low-cloud fraction (LCF) in the eastern portion of an ocean basin (Klein and Hartmann 1993; Wood and Bretherton 2006). Those equatorward surface winds yield cold advection, thus destabilizing the surface layer, and thereby facilitating shallow convection in the boundary layer, to further increase LCF.

Other regions of large LCF are found over the midlatitude–subpolar oceans, where not only stratocumulus but also fog and stratus are frequently observed (Norris 1998; Koshiro and Shiotani 2014). As typically observed in the eastern subtropics, cold air advection near the surface is favorable for stratocumulus formation. The occurrence of fog and stratus is, by contrast, more likely under warm advection that renders surface air temperature higher than SST underneath (Norris and Klein 2000; Norris and Iacobellis 2005; Tokinaga et al. 2009; Tanimoto et al. 2009; Koshiro et al. 2017). It has been shown that seasonal-mean LCF can be explained well by lower-tropospheric stability (Klein and Hartmann 1993; Wood and Bretherton 2006; Koshiro and Shiotani 2014).

In this study we focus on low-level clouds over the south Indian Ocean, where the subtropical Mascarene high exhibits a distinct seasonality from its counterpart in other ocean basins (Figs. 1a,d; Rodwell and Hoskins 2001; Lee et al. 2013). The Mascarene high resides over the eastern portion of the basin in summer (Fig. 1a), while it strengthens and shifts westward in winter (Fig. 1d). Correspondingly, the area of large LCF over the south Indian Ocean also shows westward extension from summer (Figs. 1a,d). Over each of the South Atlantic and Pacific, by contrast, the subtropical high resides in the eastern portion of the basin throughout the year, and thus such westward extension of a large LCF area into winter as seen over the south Indian Ocean is not apparent. Still, the seasonality of low-level clouds over the south Indian Ocean has not been examined in detail.

Another important factor that affects low-level clouds over the south Indian Ocean is a prominent oceanic front that forms in the confluent zone of the Agulhas Return Current and Antarctic Circumpolar Current (Figs. 1c,f). This midlatitude oceanic front extends nearly zonally around 45°S from 20° to 90°E and is characterized by pronounced SST gradient. It has been shown that this prominent SST front acts to maintain a surface baroclinic zone locally and thereby anchors a storm track along which migratory cyclones and anticyclones recurrently develop (Nakamura and Shimpf 2004; Nakamura et al. 2008; Nonaka et al. 2009; Hotta and Nakamura 2011). Subweekly fluctuations in surface winds associated with cyclones and anticyclones act to relax the surface air temperature gradient around the SST front, yielding cross-frontal gradient of surface sensible heat flux (SHF) (Nonaka et al. 2009; Hotta and Nakamura 2011). In fact, subweekly fluctuations in surface meridional wind velocity are enhanced around the particular SST front, especially in winter (Figs. 1b,e; Nakamura and Shimpf 2004).

In the following, we investigate the seasonality of the impacts of the Mascarene high, the storm track, and the SST front on LCF over the south Indian Ocean in detail, and the distinct seasonality is then highlighted in comparison with the corresponding seasonality over the South Atlantic and Pacific. Since in situ observations of low-level clouds have been scarce over the Southern Ocean, our investigation takes advantage of the Moderate Resolution Imaging Spectroradiometer (MODIS) on *Terra*, whose horizontal resolution is sufficient for capturing impacts of the oceanic front on low-level clouds. Accumulated from March in 2000, the MODIS data can be utilized for investigating the climatological distribution of LCF in detail. The rest of this paper is organized as follows. In section 2, the data used in this study are introduced. In section 3, climatological distribution of LCF and its seasonality over the south Indian Ocean are investigated in detail. Multiple linear regression model analysis on the seasonality of the subtropical LCF and comparison with the other ocean sectors are made in section 4. Conclusions are given in section 5.

2. Data and methodology

a. Cloud data

In this study, we use the collection 06 level-3 daily cloud product of MODIS on a $1^\circ \times 1^\circ$ grid (King et al. 2006; Hubanks et al. 2016). The horizontal resolution is adequate for extracting impacts of the oceanic fronts, as found in the SST climatologies after being

FIG. 1. (a) Climatological-mean SLP (contoured for every 4 hPa, with thick lines for 1020 hPa) and LCF (color shaded for every 10% as indicated on the right) over the southern oceans in austral summer (DJF). (b) Climatological-mean surface winds (m s^{-1} , arrows with reference on the right) and the variance of 8-day high-pass-filtered fluctuations of surface meridional wind (color shaded for every $6 \text{ m}^2 \text{ s}^{-2}$ as indicated on the right) for DJF. (c) Climatological-mean SST (contoured for every 3°C) and its meridional gradient [color shaded for every $0.4^\circ\text{C} (100 \text{ km})^{-1}$] for DJF. (d)–(f) As in (a)–(c), but for austral winter (JJA). The climatologies are for the 2002–15 period. LCF is based on MODIS and estimated under random overlap assumption, whereas the other variables are based on the ERA-Interim.

interpolated onto the MODIS grid (Figs. 1c and 1f). Only daytime (~ 1030 LT) data are analyzed for a 14-yr period from January 2002 to December 2015. We utilize a cloud mask, which is not subjected to the additional screening for the retrieval of cloud optical properties (called clear-sky restoration), and cloud-top pressure as well. In this study, those clouds whose top

pressures are higher than 680 hPa are regarded as “low-level clouds.” Since MODIS cannot detect low-level clouds if overlapped with mid- and/or high-level clouds, the random overlap assumption is used for reducing the influence of mid- and high-level clouds (e.g., Weare 2000). This is a reasonable assumption outside the areas of deep convection and landmass (McCoy et al. 2014;

Li et al. 2015). In this study, seasonality in LCF over the subpolar ocean is not discussed in detail, because large solar zenith angles (greater than 65° – 70°) are likely to introduce errors into the MODIS cloud retrievals (Grosvenor and Wood 2014).

b. Meteorological parameters

For meteorological parameters, we use the ERA-Interim global atmospheric reanalysis data (Dee et al. 2011). The $0.75^{\circ} \times 0.75^{\circ}$ gridded data have been interpolated linearly onto the $1^{\circ} \times 1^{\circ}$ MODIS grid. We use the data only at 0600 UTC for the south Indian Ocean, at 1200 UTC for the South Atlantic, and at 1800 UTC for the South Pacific (east of the international date line), so as to roughly correspond to the overpass time of *Terra* for each of the basins. Since the SST data prescribed at the lower boundary of the atmospheric model for data assimilation for the ERA-Interim have been significantly improved since January 2002 (Masunaga et al. 2015), we utilize the data from 2002 to 2015 of SHF, surface latent heat flux (LHF), 700-hPa vertical velocity (ω), 700-hPa relative humidity (RH), 2-m surface air temperature (SAT), 10-m surface winds, sea level pressure (SLP), and SST. It has been shown that air–sea flux products somewhat differ from one dataset to another (e.g., Smith et al. 2011; Liu et al. 2011; Yu et al. 2011), which suggests that SHF and LHF in the ERA-Interim also include some uncertainties. We have confirmed that qualitatively the same results are obtained when surface fluxes in Japanese Ocean Flux Data Sets with Use of Remote Sensing Observations (J-OFURO; <https://j-ofuro.scc.u-tokai.ac.jp/en/>) for the period of 2002–13 are used in place of the ERA-Interim data (see Figs. S1–S4 in the supplemental material). In addition, we use the SST data prescribed for the ERA-Interim. Near-surface temperature advection and wind divergence have been evaluated on a daily basis at the lowest level of the forecast model used for the ERA-Interim data, as in Masunaga et al. (2015).

In this study, the estimated inversion strength (EIS) defined by Wood and Bretherton (2006) is used as a measure of the strength of the inversion layer at the top of the boundary layer:

$$\text{EIS} = (\theta_{700} - \theta_{\text{sfc}}) - \Gamma_m^{850}(z_{700} - z_{\text{LCL}}), \quad (1)$$

where θ_{700} and θ_{sfc} denote potential temperature at 700 hPa and the surface, respectively, whereas z_{700} and z_{LCL} denote local altitudes of the 700-hPa surface and lifting condensation level, respectively. In (1), Γ_m^{850} signifies the moist adiabatic lapse rate at the 850-hPa level. EIS is calculated from the ERA-Interim data.

3. Influence of the Mascarene high and the storm track on LCF and its seasonality over the south Indian Ocean

a. Overview of climatological distribution of LCF

Before making detailed discussions on the relationship between LCF and meteorological conditions, we give an overview of climatological-mean distributions of LCF over the south Indian Ocean for summer (Fig. 2a) and winter (Fig. 2b). Across the subtropical basin (equatorward of $\sim 35^{\circ}\text{S}$), zonal inhomogeneity is evident in summertime LCF, with a distinct local maximum off the west coast of Australia around 105°E and rapid decline toward the west (Fig. 2a). By contrast, wintertime LCF is more zonally uniform across the subtropical basin (Fig. 2b). Both in summer and winter, LCF over the midlatitude and subpolar oceans is higher than over the subtropical ocean, and its distribution exhibits a high degree of zonal uniformity. Figures 2a and 2b indicate higher LCF in winter than in summer, although the wintertime MODIS retrievals are likely to suffer from errors under large solar zenith angles (Grosvenor and Wood 2014). In fact, cloud-top height-optical depth histograms based on the Multiangle Imaging Spectroradiometer (MISR) measurements (Marchand et al. 2010) show that climatological-mean LCF estimated under the random overlap assumption decreases into winter (not shown), which appears to be consistent with the zonal-mean LCF over the Southern Ocean (McCoy et al. 2014). In the following, we therefore avoid detailed investigation of the wintertime low-level clouds over the subpolar ocean.

Previous studies have indicated that climatological-mean distribution of LCF and its seasonality are well explained by lower-tropospheric stability, whose enhancement acts to increase LCF (e.g., Klein and Hartmann 1993; Wood and Bretherton 2006; Koshiro and Shiotani 2014). Climatological-mean distributions of EIS defined in (1) are thus shown in Figs. 2c and 2d for DJF and JJA, respectively. Across the summertime subtropical basin (Fig. 2c), EIS maximizes off the west coast of Australia around 105°E , in good correspondence with the spatial pattern of LCF (Fig. 2a). This local maximum of EIS corresponds to the longitudinal minimum in SST around 105°E (Fig. 2a). In fact, SST just off the Australian coast is slightly higher because of the southward Leeuwin Current, which is a unique feature of the south Indian Ocean from other ocean basins at equivalent latitudes (Smith et al. 1991; Kataoka et al. 2014). Compared to SST, 700-hPa potential temperature is more zonally uniform (Fig. 2c), but there is a slight zonal asymmetry. The θ_{700} at 30°S minimizes around

FIG. 2. Climatological distributions for austral summer (DJF) of maritime (a) LCF (color shaded for every 8%) and SST (contoured for every 3°C), (c) EIS (color shaded for every 2 K) and θ_{700} (contoured for every 3 K), (e) 700-hPa ω (color shaded for every 15 hPa day⁻¹) and 700-hPa RH (contoured for every 3 K), (g) near-surface temperature advection (color shaded for every 0.4 K day⁻¹), SST (contoured for every 3°C) and surface winds (m s⁻¹, arrows), (i) SHF (color shaded for every 4 W m⁻²; positive values for upward flux) and SST (contoured for every 3°C), and (k) LHF (color shaded for every 30 W m⁻²; positive values for upward flux) and SST (contoured for every 3°C). (right) As in (left), but for austral winter (JJA). In (b), the dotted line indicates the latitude poleward of which more than 30% of observations were made under (daily maximum) solar zenith angles >65°.

FIG. 3. Meridionally high-pass-filtered fields around the Agulhas SST front for austral summer (DJF) of (a) LCF (color shaded for every 0.8%), (b) EIS (color shaded for every 0.3 K), (c) 700-hPa ω (color shaded for every 2 hPa day⁻¹), (d) SHF (color shaded for every 2 W m⁻²), (e) 700-hPa RH (color shaded for every 0.5%), and (f) wind divergence near the surface (color shaded for every 0.5×10^{-6} s⁻¹). Superimposed with the contours is meridionally high-pass-filtered climatological SST (every 0.5°C; solid and dashed lines for positive and negative values, respectively; zero contours are omitted) for DJF. Local departures of each variable from its meridional nine-point running-mean values are regarded as the meridionally high-pass-filtered component.

85°E and increases gradually toward the east. At 29.5°S, for example, the SST difference from 105.5° to 85.5°E is -1.6 K, whereas the corresponding θ_{700} difference is 1.6 K. Thus, the increase in θ_{700} toward the east from 85°E also contributes to the local maximum of EIS off the west coast of Australia. This eastward warming is contributed to, in part, by the enhanced subsidence around 30°S, 110°E (Fig. 2e). The enhanced subsidence and associated surface divergence are required to keep the vorticity balance with strong equatorward winds in the eastern portion of the surface subtropical high (Fig. 2g; Rodwell and Hoskins 2001; Miyasaka and Nakamura 2010). Compared to the summertime situation, EIS in winter exhibits more zonally uniform distribution across the subtropical basin (Fig. 2d) in accordance with the spatial pattern of LCF (Fig. 2b). The high degree of zonal uniformity in wintertime EIS reflects a zonally uniform distribution of SST (Fig. 2b),

700-hPa potential temperature (Fig. 2d), and free-tropospheric subsidence (Fig. 2f). Poleward of 30°S, EIS in summer increases rapidly with latitude and maximizes around 50°S (Fig. 2c). The poleward increase in EIS is consistent with the latitudinal distribution of LCF (Fig. 2a). Overall, the spatial patterns of LCF over the south Indian Ocean in both summer and winter are in reasonable correspondence with those of EIS, as previous studies have indicated.

We point out, however, that there are several aspects that cannot be explained fully by EIS. For example, EIS increases rapidly across the midlatitude SST front along the Agulhas Return Current and maximizes on the poleward flank of the front (Fig. 2c), but the corresponding increase in LCF across the Agulhas SST front is weaker (Fig. 2a). In fact, over the subpolar ocean poleward of 50°S, EIS slightly decreases with latitude, while LCF still increases with latitude. Though not our

primary focus, this discrepancy is discussed briefly in [section 4](#). Other discrepancies in detailed aspects in seasonal cycles are discussed in [section 3c](#).

b. LCF around the Agulhas SST front

[Figure 3](#) shows meridionally high-pass-filtered fields of LCF, oceanic, and meteorological parameters around the Agulhas SST front for DJF. The meridional high-pass filtering has been applied, in order to highlight the impacts of the meridionally pronounced SST gradient across the front. Here local departures of a given variable from the meridional nine-point running-mean values are regarded as meridionally high-pass-filtered components. This procedure is equivalent to 9° latitudinal high-pass filtering. Comparison between [Figs. 3a and 3b](#) reveals opposing tendencies between LCF and EIS across the front. Specifically, LCF shows local maxima (minima) on the equatorward (poleward) flank of the SST front ([Fig. 3a](#)), albeit EIS exhibits local minima (maxima) on the equatorward (poleward) flank of the front ([Fig. 3b](#)) in accordance with the underlying SST distribution. [Figure 3d](#) shows the distribution of (upward) SHF, whose enhancement acts to increase LCF (e.g., [Xu et al. 2005](#); [Mauger and Norris 2010](#)). There is indeed local enhancement (reduction) of SHF on the equatorward (poleward) flank of the SST front, which overall coincides with the local maxima (minima) of LCF. On the warmer side of the SST front, the enhanced heat release (SHF) from the underlying ocean acts to lower SLP locally (not shown) and thereby induce wind convergence near the surface ([Fig. 3f](#)) and vice versa on the cooler side of the SST front with locally enhanced near-surface divergence. These cross-frontal contrasts are considered to be through the hydrostatic effect ([Lindzen and Nigam 1987](#); [Shimada and Minobe 2011](#)). The corresponding weakening of lower-tropospheric subsidence on the warmer side of the SST front ([Fig. 3c](#)) may be in part a signature of this hydrostatic effect, while no systematic signals are found in free-tropospheric relative humidity around the SST front ([Fig. 3e](#)).

A close inspection of [Fig. 3f](#) reveals, however, that the local maxima and minima of wind divergence near the surface tend to be shifted slightly eastward of the corresponding pattern of SST. This slight eastward shift is attributable to the advective effect by the prevailing near-surface westerlies and/or to enhanced (suppressed) turbulent mixing over warmer (cooler) SST and resultant downward transfer of the momentum of the prevailing westerlies within the boundary layer ([Wallace et al. 1989](#); [Chelton et al. 2004](#); [O'Neill et al. 2010](#)). The prevailing near-surface westerlies are characteristic of enhanced activity of synoptic-scale

atmospheric eddies migrating along the storm track (e.g., [Nakamura et al. 2008](#)), through their poleward heat transport and resultant downward transfer of wind momentum of the upper-tropospheric westerly jet. The enhanced storm-track activity is maintained in the presence of the Agulhas SST front ([Nakamura and Shimp0 2004](#); [Nakamura et al. 2008](#)). In association with the local minima and maxima of near-surface wind divergence ([Fig. 3f](#)), free-tropospheric subsidence tends to be weaker (stronger) over warmer (cooler) SST ([Fig. 3c](#)), but their spatial patterns do not coincide perfectly. This inconsistency may be due to synoptic-scale atmospheric eddies that dynamically induce subsidence climatologically on the warmer side of the SST front. Both observational studies (e.g., [Norris and Leovy 1994](#); [Clement et al. 2009](#); [Qu et al. 2014, 2015](#); [Seethala et al. 2015](#); [Myers and Norris 2016](#); [McCoy et al. 2017](#)) and modeling studies (e.g., [Rieck et al. 2012](#); [Brient and Bony 2013](#); [Bretherton and Blossey 2014](#)) suggested that LCF decreases with increasing SST. We then argue that direct impacts of EIS and SST should act to decrease LCF on the equatorward flank of the front, and that the local maximum of LCF is likely due to the locally enhanced SHF. Additionally, the locally suppressed wind divergence near the surface and subsidence aloft induced by enhanced SHF can also make additional contributions to the local increase in LCF.

[Figure 2g](#) shows near-surface temperature advection based on climatological-mean winds and temperature, which is generally weak along the warmer side of the Agulhas SST front. The sole noticeable exception is found around 43°S , 60°E , where the mean westerlies can yield modest cold advection as a result of a slight meridional tilt of the frontal axis. We therefore argue that the primary contributor to the enhanced SHF on the equatorward flank of the SST front must be cold advection by transient atmospheric eddies that recurrently develop along the storm track near the SST front ([Fig. 1b](#)). Although eddy-associated fluctuating winds across the SST front act to relax the pronounced cross-frontal SAT gradient, resultant instantaneous enhancement of SHF on the warmer side of the SST front acts to restore the cross-frontal SAT gradient efficiently ([Nonaka et al. 2009](#)), and their accumulated effects lead to the climatological enhancement in SST – SAT on the equatorward flank of the front ([Nakamura et al. 2008](#); [Hotta and Nakamura 2011](#)). In fact, the climatological-mean net cold advection caused by submonthly fluctuations in wind and temperature is recognizable along the warmer side of the SST front (not shown). Submonthly oceanic eddies around the Agulhas Return Current and Antarctic Circumpolar Current may also contribute to

FIG. 4. Dependence of daily (a) SHF (W m^{-2} ; right ordinate), (b) SST – SAT (K; right ordinate), and (c) surface scalar wind speed (m s^{-1} ; right ordinate) on surface meridional wind velocity in DJF. Gray bars are daily frequency (%) left ordinate) of surface meridional wind (m s^{-1}) at individual grid points within the domain on the equatorward flank of the Agulhas SST front (AF; 37.5° – 41.5°S , 40.5° – 54.5°E). Tick marks along the abscissa indicate bin boundaries. Leftmost and rightmost bars include all the samples of the strongest northerlies and southerlies, respectively. The black solid lines indicate climatological-mean values of the individual variables (right ordinate) for the individual bins with the 90% confidence intervals, while the climatological-mean values averaged over all the samples are represented by the black dashed horizontal lines. See [appendix A](#) for the details of the confidence level. The gray dashed vertical lines indicate climatological-mean meridional wind. Standard deviation of surface meridional wind is added on the top of (a). (d)–(f) As in (a)–(c), but for the dependence of the variables on surface zonal wind velocity.

the climatological-mean SHF ([Frenger et al. 2013](#)). Quantification of their impact is, however, difficult based on the data used in our study, as the horizontal resolution is inadequate for resolving the eddies.

The processes that influence SHF around the Agulhas SST front are investigated in detail based on analysis of the daily ERA-Interim data, and the results are summarized in [Fig. 4](#). Since SST – SAT increases almost linearly with the enhanced surface southerlies ([Fig. 4b](#)) and scalar wind speed is enhanced in case of strong southerlies and northerlies ([Fig. 4c](#)), SHF is strongly enhanced under the strong southerlies while the SHF decline under the enhanced northerlies is relatively modest ([Fig. 4a](#)). Likewise, since scalar wind speed also increases under both enhanced easterlies and westerlies ([Fig. 4f](#)), while SST – SAT is insensitive to zonal wind velocity ([Fig. 4e](#)), SHF increases under both enhanced westerlies and easterlies ([Fig. 4d](#)).

To evaluate the impacts of nonlinearity embedded in scalar wind speed on SHF more quantitatively, SHF is reconstructed as SHF_r by using a simple bulk formula:

$$\text{SHF}_r = CW(\text{SST} - \text{SAT}) \quad (2)$$

and

$$C = \rho c_p c_h, \quad (3)$$

where ρ and c_p denote air density and specific heat of air at constant pressure, respectively; and heat transfer coefficient c_h is set to 0.0015 ([Holton and Hakim 2012](#)). We decompose scalar wind speed at the surface, denoted as W in (2), for a given month into the wind speed $\hat{W}$ calculated from monthly mean zonal and meridional wind velocities, and the residual W^* that represents the contributions from submonthly wind fluctuations:

$$W = \hat{W} + W^*. \quad (4)$$

Note that W^* includes not only the contributions from subweekly fluctuations associated migratory transient eddies but also the contributions from more locally persistent fluctuations associated with quasi-stationary

FIG. 5. (a),(b) As in Fig. 2i, but for the reconstructed climatological-mean SHF ($\overline{\text{SHF}}_r$; color shaded for every 4 W m^{-2} as indicated at the bottom of this figure) through the decomposition in (5) in the text for austral summer (DJF) and winter (JJA). (c),(d) As in (a),(b), but for the term $C\bar{W}(\text{SST} - \text{SAT})$ in decomposition in (5). (e),(f) As in (a),(b), but for the term $C\bar{W}^*(\text{SST} - \text{SAT})$ in decomposition in (5). (g),(h) As in (a),(b), but for the term $C(W^*)'(\text{SST} - \text{SAT})'$ in decomposition in (5). In all the panels, SST is superimposed with the contours (every 3°C).

atmospheric eddies. If this decomposition is incorporated into (2), the seasonal-mean climatology of the reconstructed SHF (SHF_r) can be decomposed into

$$\begin{aligned} \overline{\text{SHF}}_r = & C\bar{W}(\overline{\text{SST} - \text{SAT}}) + C\bar{W}^*(\overline{\text{SST} - \text{SAT}}) \\ & + C(W^*)'(\overline{\text{SST} - \text{SAT}})', \end{aligned} \quad (5)$$

where overbars denote the climatological means and primes denote deviations from the climatological mean

(i.e., anomalies). Note that $\bar{W}^*$ is not necessarily zero because of the nonlinearity of scalar wind speed.

The climatological-mean SHF_r for summer (Fig. 5a) and winter (Fig. 5b) reproduces the spatial patterns of SHF (as a product of the ERA-Interim) for the respective seasons (Figs. 2i,j), including its distinct gradient across the Agulhas SST front, although SHF_r generally has slight negative biases across the basin for both seasons. Overall, the spatial pattern of SHF_r over the basin is accounted for primarily by the contribution

FIG. 6. Climatological-mean seasonal cycle of the longitudinal distributions of (a) LCF (color shaded for every 5%), (b) EIS (color shaded for every 1 K), (c) SST (color shaded for every 2°C), (d) near-surface temperature advection (color shaded for every 5 W m⁻²), (e) SHF (color shaded for every 15 hPa day⁻¹), (f) 700-hPa ω (color shaded for every 10%), and (g) 700-hPa RH (color shaded for every 10%) across the subtropical basin of the south Indian Ocean at 20.5°S. For each of the panels, the coloring convention is indicated at the bottom, and the contours indicate the corresponding seasonal cycle of the meridional component of surface winds (every 2 m s⁻¹; positive values for the southerlies). In (d), the climatological-mean temperature advection is calculated only from the climatological-mean winds and temperature.

from the climatological-mean surface wind velocities for each of the seasons (Figs. 5c,d). To the sharp decline of SHF_r poleward across the Agulhas SST front, by contrast, submonthly fluctuations in surface winds make a positive contribution (Figs. 5e,f), which is comparable to that from the climatological-mean wind velocities (Figs. 5c,d). The positive contribution from the submonthly fluctuations is obvious, especially toward the enhanced upward SHF on the equatorward flank of the Agulhas SST front. In winter (Fig. 5f), enhancement of upward SHF is also recognized on the equatorward flank of the zonally extending subtropical SST front around 25°–35°S. Finally, the contribution from the covariance term $C(W^*)'(SST - SAT)^T$ is found negligible for each of the seasons (Figs. 5g,h).

A comparison among Figs. 2, 3, and 5 reveals that, on the equatorward flank of the Agulhas SST front, transient atmospheric eddies that recurrently develop along the storm track increase the occurrence of those events where SST – SAT and/or wind speed are augmented, resulting in particularly large SHF and thereby acting to facilitate the formation of shallow convective clouds. The resultant increase in LCF contributes to the weaker gradient of LCF across the SST front than what is anticipated from the corresponding gradients of EIS

(Fig. 2c) or SST (Fig. 2a). Similarly, enhanced wind speed by submonthly atmospheric eddies also augment LHF (not shown), resulting in local maximum of LHF on the equatorward flank of the Agulhas SST front (Figs. 2k,l).

c. Seasonal cycle of LCF in the subtropical south Indian Ocean

To discuss the seasonality of LCF and its longitudinal distribution across the subtropical basin of the south Indian Ocean, longitude–time sections of climatological-mean LCF and the corresponding meteorological and oceanic variables at 20.5° and 28.5°S are shown in Figs. 6 and 7, respectively. The figures reveal complex relationships of LCF with meteorological parameters in the course of its seasonal cycle over the subtropical south Indian Ocean. At 20.5°S (Fig. 6b), EIS is larger in summer than in winter within the eastern subtropics (95°–110°E), which seems consistent with the winter–summer difference in LCF shown in Figs. 2a,b. As evident in Fig. 6a, however, in addition to its primary maximum in summer, LCF exhibits a secondary maximum in winter, which is not fully explained by EIS (Fig. 6b). In the western and central subtropics (50°–90°E), wintertime EIS is slightly higher than its summertime counterpart, which is again seemingly

FIG. 7. As in Fig. 6, but for 28.5°S.

consistent with the winter–summer difference in LCF (Figs. 2a,b). Despite the well-defined wintertime maximum in LCF, however, EIS maximizes in spring but not in winter. Almost all the characteristics of the seasonal cycle observed at 20.5°S can also be recognized at 28.5°S (Fig. 7), except that LCF exhibits its more prolonged maximum extending from July to October.

One of the possible factors that cause the aforementioned discrepancies between EIS and LCF in the subtropics may be SHF. Both at 20.5° and 28.5°S, SHF maximizes in winter (Figs. 6e and 7e), which can facilitate the formation of shallow convective clouds and thereby increase LCF in winter despite the spring maximum of EIS (Figs. 6b and 7b). Another contributor to the wintertime enhancement of LCF is probably SST. Climatologically SST is the lowest in August and September at both 20.5° and 28.5°S after wintertime deepening of the ocean mixed layer and resultant increase in thermal inertia. In the presence of negative dependence of LCF on SST (e.g., Rieck et al. 2012; Brient and Bony 2013; Bretherton and Blossey 2014; Qu et al. 2014, 2015; Seethala et al. 2015; McCoy et al. 2017), relatively cool SST in winter can contribute to an increase of LCF. At the same time, the decline of SST from late autumn into early spring acts to increase EIS continuously into early spring.

At 20.5°S, 700-hPa subsidence is enhanced in winter between 50° and 80°E (Fig. 6f) in accordance with the westward shift of the Mascarene high. As shown by Myers and Norris (2013), enhanced subsidence under

fixed EIS can contribute to the wintertime reduction of LCF, while acting to increase EIS by raising free-tropospheric temperature adiabatically. At 28.5°S, the seasonal cycle of subsidence (Fig. 7f) is similar to that in 20.5°S, although the summertime maximum off the west coast of Australia is more pronounced. We thus conjecture that the wintertime enhancement of LCF may be caused by the enhanced SHF under the lowering SST. Finally, free-tropospheric relative humidity tends to be lower in winter than in summer at both 20.5° and 28.5°S (Figs. 6g and 7g), which may also possibly act to reduce wintertime LCF (e.g., Bretherton et al. 2013; van der Dussen et al. 2015).

To discuss why SHF is enhanced over the almost entire subtropical basin in winter, near-surface temperature advection is calculated. In summer, the Mascarene high resides over the eastern portion of the basin, and enhanced cold advection is limited off the west coast of Australia (Figs. 2g, 6d, and 7d), resulting in locally enhanced upward SHF. Equatorward of 30°S, as the Mascarene high shifts westward into winter, cold advection occurs almost entirely across the basin (Figs. 2h, 6d, and 7d). In early winter, SST – SAT is still relatively high. Consequently, SHF is even larger in the eastern subtropics in winter than in summer (Figs. 2i,j, 6e, and 7e), despite the weaker cold advection off the west coast of Australia in winter than in summer (Figs. 2g,h, 6d, and 7d).

The wintertime enhancement of storm-track activity (Figs. 1b and 1e; Nakamura and Shimpo 2004) also

FIG. 8. (a) As in Fig. 4, but for JJA along the equatorward flank of the subtropical SST front (ST; 27.5°–29.5°S, 60.5°–109.5°E).

contributes to the enhanced SHF in the wintertime subtropics (Figs. 2i,j, 6e, and 7e). As shown in Fig. 1, an SST front forms in winter around 30°S extending zonally across the basin (Graham and De Boer 2013). Along the warmer flank of this subtropical SST front, the climatological-mean surface cold advection is augmented in winter (not shown) across the SST gradient, which is rather modest though, because of the climatological-mean southerlies associated with the Mascarene high (Fig. 2h) and temporally enhanced southerlies associated with submonthly atmospheric disturbances. Both the mean southerlies and submonthly eddy activity intensify climatologically into winter. Furthermore, fluctuations in surface winds also augment surface wind speed.

Dependence of wintertime SHF on fluctuating surface winds is shown in Fig. 8 along the equatorward flank of the subtropical SST front. Meridional winds strongly fluctuate with a standard deviation of 4.7 m s^{-1} around the climatological-mean southerlies ($+1.7 \text{ m s}^{-1}$), and scalar wind speed is enhanced temporally under both strong northerlies and southerlies (Fig. 8c), which is analogous to the situation along the warmer side of the Agulhas SST front in summer (Fig. 4c). Given the large positive value of SST – SAT under the enhanced southerlies and its strongly reduced values under the enhanced northerlies (Fig. 8b), upward SHF strongly increases under the enhanced southerlies while its reduction is

modest under the enhanced northerlies (Fig. 8a), resulting in the net positive contribution from transient atmospheric eddies to the climatological-mean SHF. Similarly, SHF is enhanced under both strong easterlies and westerlies (Fig. 8d) as a result of nonlinear dependence of scalar wind speed (Fig. 8f), combined with the corresponding weak dependence of SST – SAT on the zonal wind velocity (Fig. 8e). Note that LHF is also enhanced by daily fluctuations in surface winds through the nonlinear increase in scalar wind speed (not shown). This augmentation of SHF by submonthly fluctuations can be confirmed through Figs. 5b,d,f,h. Comparison between Figs. 5b and 5d reveals that the climatological-mean trades associated with the Mascarene high make the dominant contribution to the reconstructed climatological-mean SHF (SHF_r) around 15°–25°S. Around 30°S, almost the two-thirds of the climatological-mean SHF_r is accounted for by the submonthly wind fluctuations (Fig. 5f), as is already noted in section 3b. The corresponding augmentation of SHF by submonthly atmospheric disturbances is much weaker in summer than in winter (Figs. 5e,f) under the weaker eddy activity (Fig. 1b) and the poleward-shifted subtropical SST front (Fig. 1c).

4. Discussion

As shown in the preceding section, SHF, SST, and EIS act to increase LCF in winter across the subtropical

FIG. 9. (a) As in Fig. 6a for 20.5°S, but from which the annual-mean LCF within the domain has been removed. (b) As in (a), but for the corresponding LCF predicted by the multiple linear regression model described in appendix B. The correlation and root-mean-square error (RMSE) between (a) and (b) are shown below (a). (c)–(g) As in (b), but for the individual contributions from EIS, SST, SHF, 700-hPa ω , and 700-hPa RH to (b), respectively.

south Indian Ocean basin under the westward extension and intensification of the Mascarene high and the enhancement of storm-track activity. One may question the relative importance of the contributions from SHF, SST, and EIS to the wintertime enhancement of LCF.

To quantify the relative contributions, LCF is reconstructed through a multiple linear regression model as described in detail in appendix B. (The regression slope obtained for the LCF variations against each of the predictors of the model is given in Table B1.) Figures 9

FIG. 10. As in Fig. 9, but for 28.5°S.

FIG. 11. (a)–(d) As in Figs. 2a–d, but for the South Atlantic. (e)–(h) As in Figs. 2g–j, but for the South Atlantic.

and 10 show the longitude–time distributions of the predicted climatological seasonal cycle of LCF at 20.5° and 28.5°S, respectively. The multiple linear regression model explains 55% and 69% of the total variance of regionality of LCF and its seasonal cycle at 20.5° and 28.5°S, respectively, whereas the root-mean-square error between the observed LCF and the predicted LCF is 8% and 4% at 20.5° and 28.5°S, respectively. Most importantly, the model well reproduces the wintertime LCF maximum from July to September in the western and central subtropics (Figs. 9a and 9b for 20.5°S; Figs. 10a and 10b for 28.5°S). In the eastern subtropics (100°–110°E), the predicted summertime LCF maximum (Figs. 9c and 10c) is underestimated (Figs. 9c and 10c). Nevertheless, the zonal contrast in LCF in summer is still reproduced by the model. Reconstruction of LCF is not successful near the coasts, which is not our primary focus.

The reconstruction indicates that at the two subtropical latitudes EIS (Figs. 9c and 10c) and SST (Figs. 9d and 10d) make the greatest contributions to the wintertime enhancement of LCF (Figs. 9b and 10b), and

the contribution from SHF is also important (Figs. 9e and 10e). By contrast, 700-hPa relative humidity acts to reduce LCF in winter (Figs. 9g and 10g). The direct impact of the 700-hPa subsidence appears to be negligible (Figs. 9f and 10f), although the enhanced subsidence (Figs. 6f and 7f) may influence the wintertime increase in LCF indirectly through EIS. Meanwhile, the summertime LCF maximum off the west coast of Australia is attributable mostly to the enhanced EIS (Figs. 9b,c and 10b,c). Nevertheless, lower SST (Figs. 9d and 10d) and enhanced SHF (Figs. 9e and 10e) in the eastern portion of the basin compared with the western portion act to augment the zonal LCF contrasts in summer.

Over the south Indian Ocean poleward of 50°S, summertime EIS decreases slightly with latitude, probably because of the stronger latitudinal decrease in free-tropospheric temperature than in SST (Figs. 2a and 2c). No corresponding latitudinal decrease is observed in LCF with extremely large values retained over the subpolar oceans (Fig. 2a). One possible contributor may be the

FIG. 12. As in Fig. 11, but for the South Pacific.

poleward decrease in SST (Fig. 2a), and another possible contributor is the poleward increase in free-tropospheric relative humidity (Fig. 2e). The latter may weaken the entrainment drying and thus act to increase LCF (e.g., Bretherton et al. 2013; van der Dussen et al. 2015), although the origin of the high relative humidity is not clear. Note that the wintertime situation (Figs. 2b,d,f) is similar to the summertime counterpart, but the wintertime LCF is probably not reliable as discussed in section 3a.

Finally, the same analysis above is applied to the other basins, in comparison with the south Indian Ocean as shown in Fig. 2. Figures 11 and 12 show the climatological-mean distributions of LCF and related meteorological and oceanic variables in the South Atlantic and the South Pacific, respectively. In summer, the climatological-mean distributions of LCF in the South Atlantic (Fig. 11a) and the South Pacific (Fig. 12a) are similar to the south Indian Ocean (Fig. 2a). LCF in the South Atlantic and the South Pacific maximizes off the west coast of South Africa between 15° and 25°S and South America between 15° and 30°S, respectively, and LCF further increases with latitude poleward of 35°S. The LCF distributions

are well explained by EIS (Figs. 11c and 12c). The local maximum of EIS in the eastern portion of each of the subtropical basin coincides with the local minimum of SST (Figs. 11a and 12a) and the local maximum of θ_{700} (Figs. 11c and 12c). The latter is associated with locally enhanced 700-hPa subsidence (not shown). In both the South Atlantic (Fig. 11e) and Pacific (Fig. 12e), the near-surface cold advection in summer in the eastern portion over the subtropical basin is not as strong as in the south Indian Ocean (Fig. 2g), despite the strong southeasterly trade winds off the coast associated with the subtropical high (Figs. 11f and 12f). The weaker cold advection is due to the weaker gradient of underlying SST in the South Atlantic (Fig. 11a) and the South Pacific (Fig. 12a) than in the south Indian Ocean (Fig. 2a). Correspondingly, the local enhancement of summertime SHF off the west coast of South Africa (Fig. 11g) and South America (Fig. 12g) is less obvious than in the south Indian Ocean (Fig. 2i).

Although the spatial distributions of the climatological-mean LCF in summer overall correspond well with those of EIS for both the South Atlantic and the South Pacific

FIG. 13. As in Fig. 3, but for (a) LCF (%), (b) EIS (K), and (c) SHF (W m^{-2}) for DJF in the South Atlantic.

(Figs. 11 and 12, respectively), regions around oceanic frontal zones can be exceptions as discussed in section 3b for the south Indian Ocean (Figs. 2 and 3). We have shown that both the pronounced SST gradient and enhanced storm-track activity act to increase LCF along the warmer side of the Agulhas SST front.

As evident in Figs. 1c and 1f, areas of strong meridional SST gradients are also observed in the other basins. Though not as pronounced as the Agulhas SST front, the Brazil–Malvinas fronts are located in the western South Atlantic around 35°–50°W (Tokinaga et al. 2005). Over the warmer side of each of these two SST fronts, upward SHF is locally enhanced (Fig. 11g), which can be confirmed in the meridionally high-pass-filtered fields (Fig. 13c). The locally enhanced SHF coincides with the enhanced LCF (Fig. 13a), suggestive of the formation of shallow convective clouds facilitated under the enhanced SHF. By contrast, the corresponding local minimum of EIS over the local maximum of SST (Fig. 13b) acts to reduce LCF over the warmer side of each of the fronts. In fact, we have repeated the same analysis as shown in Fig. 5, to confirm that augmentation of SHF by submonthly atmospheric disturbances is essential for the large SHF over the warmer side of each SST front (not shown). The cross-frontal SHF gradient is also strong around 55°S, 140°–170°W in the South Pacific (Fig. 12g), although the corresponding impact on LCF is found to be rather modest (not shown). In summary, the combined impacts of the frontal SST gradient and submonthly atmospheric disturbances on LCF, as discussed in section 3b for the south Indian Ocean, can also be seen in the South Atlantic and the South Pacific, although the impacts are not as prominent as around the Agulhas SST front.

The right columns of Figs. 11 and 12 show the climatological-mean wintertime distributions of LCF and related meteorological and oceanic variables in the South Atlantic and South Pacific, respectively. Unlike in the subtropical south Indian Ocean (Fig. 2b), zonal asymmetry of LCF is evident in both the South Atlantic (Fig. 11b) and the South Pacific (Fig. 12b), as observed in summer (Figs. 11a and 12a). Correspondingly, EIS exhibits its maximum off the coasts of South Africa and South America (Figs. 11d and 12d), collocated with the local minimum of SST (Figs. 11b and 12b), as in summer

(Figs. 11c and 12c). Though less obvious than in summer, the local maximum of θ_{700} is also observed over the same coastal regions (Figs. 11d and 12d). Compared to the summertime situation, cold advection in the coastal regions associated with the subtropical high is enhanced in winter in each of the basins (Figs. 11f and 12f), resulting in local enhancement in SHF around 10°S, 10°W–10°E in the South Atlantic (Fig. 11h) and around 10°S, 90°W in the South Pacific (Fig. 12h). These features are all attributable to the subtropical highs that reside over the eastern portions of the South Atlantic and the South Pacific basins throughout the year, which is in sharp contrast with the south Indian Ocean (Figs. 1a,d).

Compared to summer, storm-track activity is enhanced in winter across the South Atlantic and South Pacific (Figs. 1b,e). In the South Pacific, the primary SST front in winter is located poleward of 50°S, far south compared with the South Atlantic and the south Indian Ocean (Fig. 1f). Still, the storm-track activity over the South Pacific is enhanced in winter around 30°S, where upper-level eddies traveling along the subtropical jet can couple with the near-surface baroclinic zone along the modest subtropical SST front in winter (Nakamura and Shimpou 2004). In addition, quasi-stationary atmospheric eddies also contribute to the submonthly fluctuations in surface winds (not shown). As also observed in the south Indian Ocean (Fig. 2), the wintertime enhancement of surface wind fluctuations acting on the subtropical SST front with modest SST gradients around 35°S in the South Atlantic and 30°S in the South Pacific in winter (Fig. 1f) can also contribute to the wintertime enhancement of SHF.

The seasonal cycles of LCF and related meteorological and oceanic variables across the subtropical basin for the South Atlantic and South Pacific are plotted in Fig. 14, which is compared with the corresponding plots in Fig. 7 for the south Indian Ocean. At 32.5°S in the South Atlantic, LCF maximizes from late winter into early spring (Fig. 14a), while EIS lags behind the wintertime maximum of LCF (Fig. 14b). The delay of the wintertime EIS increase relative to the LCF maximum is in common with the south Indian Ocean. By contrast, SHF is basically enhanced across the basin from late autumn to winter (Fig. 14e), which can act to facilitate the formation of shallow convective clouds and thereby to increase LCF in

FIG. 14. As in Fig. 7, but for (a) LCF (%), (b) EIS (K), (c) SST (°C), (d) near-surface temperature advection (K day^{-1}), and (e) SHF (W m^{-2}) at 32.5°S . (f)–(j) As in (a)–(e), but for 28.5°S in the South Pacific. In (d) and (i), the climatological-mean temperature advection is calculated only from climatological-mean winds and temperature.

winter. This wintertime enhancement of SHF is not explained well by near-surface temperature advection because of climatological-mean winds (Fig. 14d), which is quite weak except over the eastern portion of the basin. We have confirmed that wind fluctuations associated with submonthly atmospheric disturbances acting on the SST gradient is the main contributor to the wintertime enhancement of SHF, especially along the warmer flank of the wintertime subtropical SST front. In addition, we have repeated the same analysis as in Fig. 5 for the south Indian Ocean, and confirmed that enhancement of SHF by submonthly atmospheric eddies through augmentation of surface wind speed also contributes to the wintertime enhancement of SHF (not shown). As in the south Indian Ocean, SST also minimizes in late winter (Fig. 14c), which can also contribute to wintertime enhancement of LCF.

At 28.5°S in the South Pacific, the seasonal cycles of LCF, EIS, SST, and SHF (Figs. 14f–h,j) across the basin shows some resemblance to their counterpart for the South Atlantic (Figs. 14a–c,e). Across the subtropical South Pacific, LCF is characterized by its prolonged maximum from winter to early spring (Fig. 14f), whereas EIS maximizes from spring into early summer and SST is the lowest in late winter to early spring (Figs. 14g,h). SHF maximizes from late autumn into late winter (Fig. 14j), despite no clear wintertime enhancement of cold advection because of climatological-mean winds (Fig. 14i). Again, this enhancement of SHF is mainly due to submonthly atmospheric eddies through enhancement of net cold advections as well as the augmentation of surface wind speed (not shown), which may contribute to the wintertime enhancement of LCF. Overall, the unique

seasonality in the position of the subtropical high characterizes the distinct seasonality in LCF over the subtropical south Indian Ocean, while basinwide augmentation of LCF through SHF by submonthly atmospheric eddies can be seen in the other basins. Applying the multiple linear regression model (Table B1) to the South Atlantic and the South Pacific, we have confirmed that the wintertime enhancement of SHF plays a role in the seasonal cycle of the midlatitude and subtropical LCF (Figs. S6 and S8 in the supplemental material) while EIS dominates in the seasonal cycle of LCF farther equatorward (Figs. S5 and S7).

5. Conclusions

In the present study, influence of the subtropical high and storm-track activity on LCF and its seasonality over the south Indian Ocean is investigated by utilizing the MODIS satellite data. In austral summer, the climatological LCF has a local maximum off the west coast of Australia (Fig. 2a), as in the other subtropical basins (e.g., Klein and Hartmann 1993). Over the midlatitude and subpolar oceans, the mean LCF is greater and distributed more zonally across the basin (Fig. 2a). Across the subtropical basin, by contrast, atmospheric circulation is changed notably in association with the westward shift of the Mascarene high in winter, and correspondingly climatological LCF is distributed more zonally than in summer (Figs. 1a,d and 2a,b).

In both seasons, the spatial pattern of the climatological-mean LCF is overall explained by the EIS distribution. In summer, the SST distribution is characterized by a local subtropical minimum off the west coast of Australia

and by a pronounced meridional gradient across the midlatitude oceanic front anchored between the Agulhas Return Current and Antarctic Circumpolar Current (Figs. 1c,f and 2a,b). Owing to lower SST (Fig. 2a) and warmer free-tropospheric temperature (Fig. 2c) associated with enhanced subsidence (Fig. 2e), EIS has a local maximum off the west coast of Australia and increases with latitude in summer (Fig. 2c). In winter, EIS is horizontally uniform in the subtropics, which is also consistent with the zonal uniformity of LCF (Fig. 2d).

However, LCF distribution around the SST front is not fully explained by EIS. EIS strongly increases with latitude (Fig. 2c), whereas the corresponding increase in LCF is much less (Fig. 2a). We have found that enhanced SHF acts to increase LCF on the equatorward flank of the Agulhas SST front (Figs. 2i and 3d). This enhanced SHF is due to the enhanced storm-track activity in the presence of the SST front. Transient atmospheric eddies not only act to relax the SAT gradient but also increase instantaneous surface wind speed, leading to extremely large upward SHF over the warmer side of the SST front (Figs. 4 and 5). This rectifying effect of storm-track activity on climatological SHF seems prominent around the Agulhas SST front. Note that submonthly atmospheric eddies may also be important for climatological-mean low-cloud top, since climatological cross-frontal contrast of low-cloud top around the SST front along the Gulf Stream resembles the corresponding contrast under cold advection events (Liu et al. 2014).

Previous studies (e.g., Trenberth and Fasullo 2010; Grise et al. 2015) pointed out that many of the global climate models have a large bias in shortwave radiation over the Southern Ocean, which can possibly affect the atmospheric circulation via SST distribution (Frierson and Hwang 2012; Hwang and Frierson 2013; Ceppi and Hartmann 2015). Although underestimation of super-cooled liquid water in mixed-phase clouds in the models is considered to be the major cause of this bias (e.g., Senior and Mitchell 1993; Tsushima et al. 2006; Komurcu et al. 2014; McCoy et al. 2015b, 2016; Tan et al. 2016; Ceppi et al. 2016a; Bodas-Salcedo et al. 2016), our analysis suggests that bias in SST front and storm-track activity simulated in the models also leads to bias in LCF and thereby in the surface radiation budget. In fact, Masunaga et al. (2015) showed the importance of high-resolution SST in realistic representation of cloudiness around the North Pacific SST fronts in the ERA-Interim. The LCF increase locally over the warmer side of an oceanic front through SHF is also recognized in the South Atlantic (Fig. 13).

The seasonal cycle of subtropical LCF is not fully explained by EIS, either. EIS maximizes in spring, whereas LCF has a more prolonged maximum from early winter (Figs. 6a,b and 7a,b). We have found that this

wintertime augmentation of LCF is due to not only EIS but also enhanced SHF (Figs. 6e and 7e) and seasonal lowering of SST (Figs. 6c and 7c). Centered at the western portion of the basin in winter, the Mascarene high enhances near-surface cold advection almost entirely across the subtropical basin (Fig. 6d), resulting in large upward SHF (Figs. 5d and 6e). In addition, seasonally enhanced storm-track activity augments SHF by enhancing scalar wind speed around 30°S (Figs. 5f and 8), where the mean cold advection by the Mascarene high is weaker than around 15°–25°S (Figs. 6d and 7d). Thus, the westward shift of the subtropical high and the seasonally enhanced storm-track activity seem essential for the unique seasonality of LCF within the subtropical south Indian Ocean via SHF, which is also supported through our multiple linear regression analysis (Figs. 9 and 10).

The corresponding analysis of LCF in the South Atlantic and the South Pacific not only highlights the uniqueness of the south Indian Ocean arising from the seasonality of the Mascarene high but also reveals common features associated with submonthly fluctuations among the three ocean basins. Compared to the south Indian Ocean, the seasonality of the subtropical highs is much weaker in the South Atlantic and Pacific. They reside over the eastern portions of the basins in both summer and winter (Figs. 1a,d). Correspondingly, LCF maximizes off the west coast of South Africa (Figs. 11a,b) and South America (Figs. 12a,b) in both seasons in accordance with the EIS distributions (Figs. 11c,d and 12c,d). We have revealed, however, that there are similarities among the three ocean basins around 30°S, where LCF exhibits a zonally extended maximum from winter to spring (Figs. 14a,f) while EIS peaks in spring (Figs. 14b,g). Unlike in the south Indian Ocean, there is no basinwide enhancement of cold advection by climatological-mean southerlies in the South Atlantic and Pacific (Figs. 14d, i). Still, under the wintertime enhancement of submonthly atmospheric eddy activity around the subtropical SST fronts (Fig. 1e), SHF is enhanced in winter around 30°S across those basins (Figs. 14e,j), which can contribute to the wintertime enhancement of LCF across the subtropical basins (Figs. 14a,f) in the South Atlantic and Pacific as in the south Indian Ocean. In the Northern Hemisphere, two major storm tracks also form along the prominent oceanic frontal zones along the Gulf Stream and Kuroshio–Oyashio Extensions (Nakamura et al. 2004; Kwon et al. 2010), and the storm-track activity is much stronger in boreal winter than in summer, though somewhat suppressed in midwinter over the Pacific (Nakamura 1992). Impact of storm-track activity on LCF through SHF over those oceanic frontal zones in comparison with the subtropical North Pacific and Atlantic will be pursued in our future study.

As reviewed by Kamae et al. (2016) and Klein et al. (2017), near-surface temperature advection is recognized

as an important cloud controlling factor that represents how strongly large-scale atmospheric circulation enhances upward SHF in favor of the formation of low-level clouds. As shown in Figs. 2g–j, 6d,e, 7d,e, and 14d,e,i,j, however, the temperature advection calculated from monthly mean temperature and winds does not fully explain SHF because of the prominent contribution from wind fluctuations associated with submonthly atmospheric disturbances (Figs. 4, 5, and 8). In a warmed climate, a small negative feedback from enhanced cold advection is suggested by recent studies (Qu et al. 2015; Myers and Norris 2016), where the near-surface air temperature advection is mimicked as “SST advection” ($-\mathbf{V} \cdot \nabla \text{SST}$) by near-surface winds $\mathbf{V}$ that include no contribution from submonthly atmospheric eddies. We suggest that SHF may be used in place of near-surface temperature advection to obtain a more realistic observational constraint on LCF’s response to global warming. Unlike the SST advection, however, SHF is a locally determined boundary layer quantity that can likely be affected by low-level clouds themselves. Thus, incorporating SHF in place of temperature advection in a multiple-linear regression model brings about a causality issue, which should be carefully handled.

As pointed out by previous studies, strong covariability in interannual variations between EIS and SST can be an issue for projecting LCF’s response to global warming (Myers and Norris 2015), since both EIS and SST are likely to increase under warmed climatic conditions (Qu et al. 2014, 2015; Myers and Norris 2016; McCoy et al. 2017). Off the Peruvian coast, the EIS–SST anticorrelation is indeed strong in the course of the seasonal cycle (Fig. 12; Wang et al. 2011). However, the anticorrelation is rather weak in the eastern portion of the subtropical south Indian Ocean as shown through our analysis (Fig. 6) and off the California coast (Wang et al. 2011). Furthermore, the time lag between EIS and SST in the course of their seasonal cycle is apparent commonly over the subtropical and midbasin domains in each of the south Indian Ocean (Fig. 7), the South Atlantic, and Pacific (Fig. 14). We therefore argue that disentanglement of the dependence of LCF on EIS, SST, and SHF may be possible over the subtropical basins, including the south Indian Ocean. Thus, seasonal variations might be better suited than interannual variations for constructing a multiple linear regression model to constrain low-cloud response to the global warming.

As revealed through our analysis, the seasonality of the Mascarene high strongly affects low-level clouds over the subtropical south Indian Ocean. In fact, enhanced long-wave cooling associated with those low-level clouds acts to strengthen land–sea contrasts of diabatic heating (Wu and

Liu 2003), contributing to the formation of summertime subtropical highs (Miyasaka and Nakamura 2005, 2010). Nevertheless, the role of low-level clouds in the formation of the wintertime Mascarene high has not been investigated, but will be pursued in our future work.

Though beyond the scope of our study, satellite-based analysis of the variability in optical thickness of low-level clouds, in addition to their fraction (LCF), is necessary for deepening our understanding of albedo of low-level clouds and its variability over the midlatitude and subpolar oceans; however, such an analysis based on satellite data has not been performed until recently (e.g., Terai et al. 2016; Ceppi et al. 2016b). Furthermore, although our focus is on how meteorological and oceanic conditions affect LCF, cloud microphysical processes and aerosol properties may also play a key role in determining LCF, and oceanic aerosol productivity can also play an important role in determining cloud condensation nuclei (e.g., Lana et al. 2011; McCoy et al. 2015a). These aspects should also be explored in future studies.

Acknowledgments. The authors are grateful for the three anonymous reviewers for their sound criticism and constructive comments on the earlier versions of this paper, which have led to substantial improvement. This study is supported in part by the Japanese Ministry of Education, Culture, Sports, Science and Technology (MEXT) through Grants-in-Aid for Scientific Research 2205 (in Innovative Area) and 16H01844 and through the Arctic Challenge for Sustainability (ArCS) Program, by the Japanese Ministry of Environment through the Environment Research and Technology Department Fund 2-1503, and by the Japan Science and Technology Agency through Belmont Forum CRA “InterDec.” The figures were produced by GFD-DENNOU Library.

APPENDIX A

Estimation of Degrees of Freedom

To obtain confidence intervals, degrees of freedom for a given variable are needed. In this study, effective degrees of freedom N_e are estimated as

TABLE A1. List of (left to right): T_e (days), X_e (°), and Y_e (°) for SHF, SST – SAT, and scalar surface wind speed and (top to bottom) for the domains around the Agulhas SST front (AF) and the subtropical south Indian Ocean (ST).

	AF	ST
SHF	1, 9°, 5°	1, 13°, 3°
SST – SAT	1, 9°, 5°	2, 11°, 3°
Surface wind speed	1, 13°, 4°	1, 18°, 3°

TABLE B1. Regression slope for each of the predictors.

$\partial\text{LCF}/\partial\text{SHF}$ [% (W m ⁻²) ⁻¹]	$\partial\text{LCF}/\partial\text{EIS}$ (% K ⁻¹)	$\partial\text{LCF}/\partial\text{SST}$ (% K ⁻¹)	$\partial\text{LCF}/\partial\omega$ [% (hPa day ⁻¹) ⁻¹]	$\partial\text{LCF}/\partial\text{RH}$ (% % ⁻¹)
0.59	3.2	-1.5	-0.042	0.36

$$N_e = \frac{N}{T_e X_e Y_e}, \quad (\text{A1})$$

where N denotes the total sample size; and T_e , X_e , and Y_e denote temporal, zonal, and meridional decorrelation scales, respectively. In (A1), T_e is evaluated by

$$T_e = 1 + 2 \sum_{t=1}^{N_d} \left(1 - \frac{t}{N_d}\right) R(t), \quad (\text{A2})$$

where $R(t)$ is autocorrelation with time lag t . The quantities X_e and Y_e can be evaluated similarly. In our evaluation, N_d is set to 16 days for T_e , while 48° is assigned for both X_e and Y_e . Nevertheless Y_e is set to the meridional width of the domain if it is less than the above value. Table A1 shows the domain-averaged T_e , X_e , and Y_e for each of the variables (SHF, SST – SAT, and scalar surface wind speed) within the two sectors of the south Indian Ocean.

APPENDIX B

Analysis of Multiple Linear Regression Model

To quantify the relative importance of cloud controlling factors in seasonal variations of LCF, daily dependence of LCF on those factors are derived through multiple linear regression as used in recent studies of subtropical LCF (e.g., Qu et al. 2014, 2015; Seethala et al. 2015; Myers and Norris 2016; McCoy et al. 2017). In this study, a regression model is constructed from daily anomalies over the subtropical south Indian Ocean (20.5°–34.5°S, 55.5°–114.5°E), which are calculated at every grid point within the domain as daily departures from the seasonal-mean climatology (for each of DJF, MAM, JJA, and SON). As predictors for the daily LCF variations EIS, SST, SHF, and 700-hPa ω and RH are chosen, and their regression slopes against the LCF variations are then derived through inverting the covariance matrix. Our regression model is unlikely to suffer substantially from the covariability among the predictors, because correlations among the predictors do not exceed 0.5 in absolute values. As noted in McCoy et al. (2017), a multiple linear regression model cannot perfectly extract the impacts of individual large-scale forcing, since 1) it takes some time for clouds to respond

to the forcing, and 2) the clouds and boundary layer properties are advected spatially by the large-scale horizontal airflow. Nevertheless, the derived local dependence is useful for quantifying their local control on LCF.

The regression slope thus derived for each variable is shown in Table B1. We have calculated confidence intervals following the supporting information of Qu et al. (2015) and found that the 2.5%–97.5% confidence interval for each slope is smaller than the slope itself by one or two orders of magnitude as a result of the large sample size. The slope $\partial\text{LCF}/\partial\text{SHF}$ is positive in the subtropics, which is consistent with the notion that enhanced heat supply from the ocean destabilizes the surface layer to facilitate the shallow convection (e.g., Mauger and Norris 2010). The positive slope $\partial\text{LCF}/\partial\text{EIS}$ is also in agreement with previous studies (e.g., Wood and Bretherton 2006; Koshiro and Shiotani 2014), which showed that enhanced inversion strength leads to higher LCF. The mean slope in the subtropics is comparable to those obtained by Qu et al. (2014), Seethala et al. (2015), and McCoy et al. (2017) based on different satellite data and/or different time scales of variability. The strongly negative $\partial\text{LCF}/\partial\text{SST}$ seems consistent with large-eddy simulation (LES) experiments under the fixed relative humidity (Rieck et al. 2012; Bretherton et al. 2013), in which a decreasing tendency for subtropical LCF with warming SST is simulated. The mean slope of -1.5% K⁻¹ is also comparable to those obtained by Qu et al. (2014) and Seethala et al. (2015). The negative slope $\partial\text{LCF}/\partial\omega$ is also consistent with Myers and Norris (2013), who showed that weakening of subsidence under the fixed inversion strength leads to larger LCF over the subtropical oceans. Finally, $\partial\text{LCF}/\partial\text{RH}$ is positive, which is supported by the notion that entrainment of drier air reduces cloudiness (Bretherton et al. 2013; van der Dussen et al. 2015). The mean slope of 0.36% %⁻¹ is comparable to the value obtained by McCoy et al. (2017).

REFERENCES

- Bodas-Salcedo, A., P. G. Hill, K. Furtado, K. D. Williams, P. R. Field, J. C. Manners, P. Hyder, and S. Kado, 2016: Large contribution of supercooled liquid clouds to the solar radiation budget of the Southern Ocean. *J. Climate*, **29**, 4213–4228, <https://doi.org/10.1175/JCLI-D-15-0564.1>.
- Bony, S., and J.-L. Dufresne, 2005: Marine boundary layer clouds at the heart of tropical cloud feedback uncertainties in climate

- models. *Geophys. Res. Lett.*, **32**, L20806, <https://doi.org/10.1029/2005GL023851>.
- Bretherton, C. S., and P. N. Blossey, 2014: Low cloud reduction in a greenhouse-warmed climate: Results from Lagrangian LES of a subtropical marine cloudiness transition. *J. Adv. Model. Earth Syst.*, **6**, 91–114, <https://doi.org/10.1002/2013MS000250>.
- , —, and C. R. Jones, 2013: Mechanisms of marine low cloud sensitivity to idealized climate perturbations: A single-LES exploration extending the CGILS cases. *J. Adv. Model. Earth Syst.*, **5**, 316–337, <https://doi.org/10.1002/jame.20019>.
- Brient, F., and S. Bony, 2013: Interpretation of the positive low cloud feedback predicted by a climate model under global warming. *Climate Dyn.*, **40**, 2415–2431, <https://doi.org/10.1007/s00382-011-1279-7>.
- Ceppi, P., and D. L. Hartmann, 2015: Connections between clouds, radiation, and midlatitude dynamics: A review. *Curr. Climate Change Rep.*, **1**, 94–102, <https://doi.org/10.1007/s40641-015-0010-x>.
- , —, and M. J. Webb, 2016a: Mechanisms of the negative shortwave cloud feedback in middle to high latitudes. *J. Climate*, **29**, 139–157, <https://doi.org/10.1175/JCLI-D-15-0327.1>.
- , D. T. McCoy, and D. L. Hartmann, 2016b: Observational evidence for a negative shortwave cloud feedback in middle to high latitudes. *Geophys. Res. Lett.*, **43**, 1331–1339, <https://doi.org/10.1002/2015GL067499>.
- Chelton, D. B., M. G. Schlax, M. H. Freilich, and R. F. Milliff, 2004: Satellite measurements reveal persistent small-scale features in ocean winds. *Science*, **303**, 978–983, <https://doi.org/10.1126/science.1091901>.
- Clement, A. C., R. Burgman, and J. R. Norris, 2009: Observational and model evidence for positive low-level cloud feedback. *Science*, **325**, 460–464, <https://doi.org/10.1126/science.1171255>.
- Dee, D. P., and Coauthors, 2011: The ERA-Interim reanalysis: Configuration and performance of the data assimilation system. *Quart. J. Roy. Meteor. Soc.*, **137**, 553–597, <https://doi.org/10.1002/qj.828>.
- Frenger, I., N. Gruber, and M. Münnich, 2013: Imprint of Southern Ocean eddies on winds, clouds and rainfall. *Nat. Geosci.*, **6**, 608–612, <https://doi.org/10.1038/ngeo1863>.
- Frierson, D. M. W., and Y.-T. Hwang, 2012: Extratropical influence on ITCZ shifts in slab ocean simulations of global warming. *J. Climate*, **25**, 720–733, <https://doi.org/10.1175/JCLI-D-11-00116.1>.
- Graham, R. M., and A. M. De Boer, 2013: The dynamical subtropical front. *J. Geophys. Res. Oceans*, **118**, 5676–5685, <https://doi.org/10.1002/jgrc.20408>.
- Grise, K. M., L. M. Polvani, and J. T. Fasullo, 2015: Reexamining the relationship between climate sensitivity and the Southern Hemisphere radiation budget in CMIP models. *J. Climate*, **28**, 9298–9312, <https://doi.org/10.1175/JCLI-D-15-0031.1>.
- Grosvenor, D. P., and R. Wood, 2014: The effect of solar zenith angle on MODIS cloud optical and microphysical retrievals within marine liquid water clouds. *Atmos. Chem. Phys.*, **14**, 7291–7321, <https://doi.org/10.5194/acp-14-7291-2014>.
- Hartmann, D. L., and D. A. Short, 1980: On the use of earth radiation budget statistics for studies of clouds and climate. *J. Atmos. Sci.*, **37**, 1233–1250, [https://doi.org/10.1175/1520-0469\(1980\)037<1233:OTUOER>2.0.CO;2](https://doi.org/10.1175/1520-0469(1980)037<1233:OTUOER>2.0.CO;2).
- Holton, J. R., and G. J. Hakim, 2012: *An Introduction to Dynamic Meteorology*. Academic Press, 552 pp.
- Hotta, D., and H. Nakamura, 2011: On the significance of sensible heat supply from the ocean in the maintenance of the mean baroclinicity along storm tracks. *J. Climate*, **24**, 3377–3401, <https://doi.org/10.1175/2010JCLI3910.1>.
- Hubanks, P., S. Platnick, M. King, and B. Ridgway, 2016: MODIS Atmosphere L3 Gridded Product Algorithm Theoretical Basis Document (ATBD) and Users Guide. NASA GSFC, MODIS Tech. Doc. ATBD-MOD-30, 122 pp., https://modis-images.gsfc.nasa.gov/_docs/L3_ATBD_C6.pdf.
- Hwang, Y.-T., and D. M. W. Frierson, 2013: Link between the double-Intertropical Convergence Zone problem and cloud biases over the Southern Ocean. *Proc. Natl. Acad. Sci. USA*, **110**, 4935–4940, <https://doi.org/10.1073/pnas.1213302110>.
- Kamae, Y., T. Ogura, H. Shiogama, and M. Watanabe, 2016: Recent progress toward reducing the uncertainty in tropical low cloud feedback and climate sensitivity: A review. *Geosci. Lett.*, **3** (17), <https://doi.org/10.1186/s40562-016-0053-4>.
- Kataoka, T., T. Tozuka, S. Behera, and T. Yamagata, 2014: On the Ningaloo Niño/Niña. *Climate Dyn.*, **43**, 1463–1482, <https://doi.org/10.1007/s00382-013-1961-z>.
- King, M. D., S. Platnick, P. A. Hubanks, G. T. Arnold, E. G. Moody, G. Wind, and B. Wind, 2006: Collection 005 change summary for the MODIS cloud optical property (06_OD) algorithm, version 3.1. NASA Rep., 23 pp., https://modis-images.gsfc.nasa.gov/C005_Changes/C005_CloudOpticalProperties_ver311.pdf.
- Klein, S. A., and D. L. Hartmann, 1993: The seasonal cycle of low stratiform clouds. *J. Climate*, **6**, 1587–1606, [https://doi.org/10.1175/1520-0442\(1993\)006<1587:TSCOLS>2.0.CO;2](https://doi.org/10.1175/1520-0442(1993)006<1587:TSCOLS>2.0.CO;2).
- , A. Hall, J. R. Norris, and R. Pincus, 2017: Low-cloud feedbacks from cloud-controlling factors: A review. *Surv. Geophys.*, **38**, 1307–1329, <https://doi.org/10.1007/s10712-017-9433-3>.
- Komurcu, M., and Coauthors, 2014: Intercomparison of the cloud water phase among global climate models. *J. Geophys. Res. Atmos.*, **119**, 3372–3400, <https://doi.org/10.1002/2013JD021119>.
- Koshiro, T., and M. Shiotani, 2014: Relationship between low stratiform cloud amount and estimated inversion strength in the lower troposphere over the global ocean in terms of cloud types. *J. Meteor. Soc. Japan*, **92**, 107–120, <https://doi.org/10.2151/jmsj.2014-107>.
- , S. Yukimoto, and M. Shiotani, 2017: Interannual variability in low stratiform cloud amount over the summertime North Pacific in terms of cloud types. *J. Climate*, **30**, 6107–6121, <https://doi.org/10.1175/JCLI-D-16-0898.1>.
- Kwon, Y.-O., M. A. Alexander, N. A. Bond, C. Frankignoul, H. Nakamura, B. Qiu, and L. Thompson, 2010: Role of the Gulf Stream and Kuroshio–Oyashio systems in large-scale atmosphere–ocean interaction: A review. *J. Climate*, **23**, 3249–3281, <https://doi.org/10.1175/2010JCLI3343.1>.
- Lana, A., and Coauthors, 2011: An updated climatology of surface dimethylsulfide concentrations and emission fluxes in the global ocean. *Global Biogeochem. Cycles*, **25**, GB1004, <https://doi.org/10.1029/2010GB003850>.
- Lee, S.-K., C. R. Mechoso, C. Wang, and J. D. Neelin, 2013: Interhemispheric influence of the northern summer monsoons on southern subtropical anticyclones. *J. Climate*, **26**, 10 193–10 204, <https://doi.org/10.1175/JCLI-D-13-00106.1>.
- Li, J., J. Huang, K. Stamnes, T. Wang, Q. Lv, and H. Jin, 2015: A global survey of cloud overlap based on CALIPSO and CloudSat measurements. *Atmos. Chem. Phys.*, **15**, 519–536, <https://doi.org/10.5194/acp-15-519-2015>.
- Lindzen, R. S., and S. Nigam, 1987: On the role of sea surface temperature gradients in forcing low-level winds and convergence in the tropics. *J. Atmos. Sci.*, **44**, 2418–2436, [https://doi.org/10.1175/1520-0469\(1987\)044<2418:OTROSS>2.0.CO;2](https://doi.org/10.1175/1520-0469(1987)044<2418:OTROSS>2.0.CO;2).

- Liu, J., T. Xiao, and L. Chen, 2011: Intercomparisons of air–sea heat fluxes over the Southern Ocean. *J. Climate*, **24**, 1198–1211, <https://doi.org/10.1175/2010JCLI3699.1>.
- Liu, J.-W., S.-P. Xie, J. R. Norris, and S.-P. Zhang, 2014: Low-level cloud response to the Gulf Stream front in winter using CALIPSO. *J. Climate*, **27**, 4421–4432, <https://doi.org/10.1175/JCLI-D-13-00469.1>.
- Marchand, R., T. Ackerman, M. Smyth, and W. B. Rossow, 2010: A review of cloud top height and optical depth histograms from MISR, ISCCP, and MODIS. *J. Geophys. Res.*, **115**, D16206, <https://doi.org/10.1029/2009JD013422>.
- Masunaga, R., H. Nakamura, T. Miyasaka, K. Nishii, and Y. Tanimoto, 2015: Separation of climatological imprints of the Kuroshio Extension and Oyashio fronts on the wintertime atmospheric boundary layer: Their sensitivity to SST resolution prescribed for atmospheric reanalysis. *J. Climate*, **28**, 1764–1787, <https://doi.org/10.1175/JCLI-D-14-00314.1>.
- Mauger, G. S., and J. R. Norris, 2010: Assessing the impact of meteorological history on subtropical cloud fraction. *J. Climate*, **23**, 2926–2940, <https://doi.org/10.1175/2010JCLI3272.1>.
- McCoy, D. T., D. L. Hartmann, and D. P. Grosvenor, 2014: Observed Southern Ocean cloud properties and shortwave reflection. Part I: Calculation of SW flux from observed cloud properties. *J. Climate*, **27**, 8836–8857, <https://doi.org/10.1175/JCLI-D-14-00287.1>.
- , S. M. Burrows, R. Wood, D. P. Grosvenor, S. M. Elliott, P.-L. Ma, P. J. Rasch, and D. L. Hartmann, 2015a: Natural aerosols explain seasonal and spatial patterns of Southern Ocean cloud albedo. *Sci. Adv.*, **1**, e1500157, <https://doi.org/10.1126/sciadv.1500157>.
- , D. L. Hartmann, M. D. Zelinka, P. Ceppi, and D. P. Grosvenor, 2015b: Mixed-phase cloud physics and Southern Ocean cloud feedback in climate models. *J. Geophys. Res. Atmos.*, **120**, 9539–9554, <https://doi.org/10.1002/2015JD023603>.
- , I. Tan, D. L. Hartmann, M. D. Zelinka, and T. Storelvmo, 2016: On the relationships among cloud cover, mixed-phase partitioning, and planetary albedo in GCMs. *J. Adv. Model. Earth Syst.*, **8**, 650–668, <https://doi.org/10.1002/2015MS000589>.
- , R. Eastman, D. L. Hartmann, and R. Wood, 2017: The change in low cloud cover in a warmed climate inferred from AIRS, MODIS, and ERA-Interim. *J. Climate*, **30**, 3609–3620, <https://doi.org/10.1175/JCLI-D-15-0734.1>.
- Miyasaka, T., and H. Nakamura, 2005: Structure and formation mechanisms of the Northern Hemisphere summertime subtropical highs. *J. Climate*, **18**, 5046–5065, <https://doi.org/10.1175/JCLI3599.1>.
- , and —, 2010: Structure and mechanisms of the Southern Hemisphere summertime subtropical anticyclones. *J. Climate*, **23**, 2115–2130, <https://doi.org/10.1175/2009JCLI3008.1>.
- Myers, T. A., and J. R. Norris, 2013: Observational evidence that enhanced subsidence reduces subtropical marine boundary layer cloudiness. *J. Climate*, **26**, 7507–7524, <https://doi.org/10.1175/JCLI-D-12-00736.1>.
- , and —, 2015: On the relationships between subtropical clouds and meteorology in observations and CMIP3 and CMIP5 models. *J. Climate*, **28**, 2945–2967, <https://doi.org/10.1175/JCLI-D-14-00475.1>.
- , and —, 2016: Reducing the uncertainty in subtropical cloud feedback. *Geophys. Res. Lett.*, **43**, 2144–2148, <https://doi.org/10.1002/2015GL067416>.
- Nakamura, H., 1992: Midwinter suppression of baroclinic wave activity in the Pacific. *J. Atmos. Sci.*, **49**, 1629–1642, [https://doi.org/10.1175/1520-0469\(1992\)049<1629:MSOBWA>2.0.CO;2](https://doi.org/10.1175/1520-0469(1992)049<1629:MSOBWA>2.0.CO;2).
- , and A. Shimpo, 2004: Seasonal variations in the Southern Hemisphere storm tracks and jet streams as revealed in a reanalysis dataset. *J. Climate*, **17**, 1828–1844, [https://doi.org/10.1175/1520-0442\(2004\)017<1828:SVITSH>2.0.CO;2](https://doi.org/10.1175/1520-0442(2004)017<1828:SVITSH>2.0.CO;2).
- , T. Sampe, Y. Tanimoto, and A. Shimpo, 2004: Observed associations among storm tracks, jet streams and midlatitude oceanic fronts. *Earth's Climate: The Ocean–Atmosphere Interaction*, *Geophys. Monogr.*, Vol. 147, Amer. Geophys. Union, 329–345.
- , —, A. Goto, W. Ohfuchi, and S.-P. Xie, 2008: On the importance of midlatitude oceanic frontal zones for the mean state and dominant variability in the tropospheric circulation. *Geophys. Res. Lett.*, **35**, L15709, <https://doi.org/10.1029/2008GL034010>.
- Nonaka, M., H. Nakamura, B. Taguchi, N. Komori, A. Kuwano-Yoshida, and K. Takaya, 2009: Air–sea heat exchanges characteristic of a prominent midlatitude oceanic front in the South Indian Ocean as simulated in a high-resolution coupled GCM. *J. Climate*, **22**, 6515–6535, <https://doi.org/10.1175/2009JCLI2960.1>.
- Norris, J. R., 1998: Low cloud type over the ocean from surface observations. Part II: Geographical and seasonal variations. *J. Climate*, **11**, 383–403, [https://doi.org/10.1175/1520-0442\(1998\)011<0383:LCTOTO>2.0.CO;2](https://doi.org/10.1175/1520-0442(1998)011<0383:LCTOTO>2.0.CO;2).
- , and C. B. Leovy, 1994: Interannual variability in stratiform cloudiness and sea surface temperature. *J. Climate*, **7**, 1915–1925, [https://doi.org/10.1175/1520-0442\(1994\)007<1915:IVISCA>2.0.CO;2](https://doi.org/10.1175/1520-0442(1994)007<1915:IVISCA>2.0.CO;2).
- , and S. A. Klein, 2000: Low cloud type over the ocean from surface observations. Part III: Relationship to vertical motion and the regional surface synoptic environment. *J. Climate*, **13**, 245–256, [https://doi.org/10.1175/1520-0442\(2000\)013<0245:LCTOTO>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<0245:LCTOTO>2.0.CO;2).
- , and S. F. Iacobellis, 2005: North Pacific cloud feedbacks inferred from synoptic-scale dynamic and thermodynamic relationships. *J. Climate*, **18**, 4862–4878, <https://doi.org/10.1175/JCLI3558.1>.
- O'Neill, L. W., S. K. Esbensen, N. Thum, R. M. Samelson, and D. B. Chelton, 2010: Dynamical analysis of the boundary layer and surface wind responses to mesoscale SST perturbations. *J. Climate*, **23**, 559–581, <https://doi.org/10.1175/2009JCLI2662.1>.
- Qu, X., A. Hall, S. A. Klein, and P. M. Caldwell, 2014: On the spread of changes in marine low cloud cover in climate model simulations of the 21st century. *Climate Dyn.*, **42**, 2603–2626, <https://doi.org/10.1007/s00382-013-1945-z>.
- , —, —, and A. M. DeAngelis, 2015: Positive tropical marine low-cloud cover feedback inferred from cloud-controlling factors. *Geophys. Res. Lett.*, **42**, 7767–7775, <https://doi.org/10.1002/2015GL065627>.
- Randall, D. A., J. A. Coakley Jr., C. W. Fairall, R. A. Kropfli, and D. H. Lenschow, 1984: Outlook for research on subtropical marine stratification clouds. *Bull. Amer. Meteor. Soc.*, **65**, 1290–1301, [https://doi.org/10.1175/1520-0477\(1984\)065<1290:OFROSM>2.0.CO;2](https://doi.org/10.1175/1520-0477(1984)065<1290:OFROSM>2.0.CO;2).
- Rieck, M., L. Nuijens, and B. Stevens, 2012: Marine boundary layer cloud feedbacks in a constant relative humidity atmosphere. *J. Atmos. Sci.*, **69**, 2538–2550, <https://doi.org/10.1175/JAS-D-11-0203.1>.
- Rodwell, M. J., and B. J. Hoskins, 2001: Subtropical anticyclones and summer monsoons. *J. Climate*, **14**, 3192–3211, [https://doi.org/10.1175/1520-0442\(2001\)014<3192:SAASM>2.0.CO;2](https://doi.org/10.1175/1520-0442(2001)014<3192:SAASM>2.0.CO;2).

- Seethala, C., J. R. Norris, and T. A. Myers, 2015: How has subtropical stratocumulus and associated meteorology changed since the 1980s? *J. Climate*, **28**, 8396–8410, <https://doi.org/10.1175/JCLI-D-15-0120.1>.
- Senior, C. A., and J. F. B. Mitchell, 1993: Carbon dioxide and climate: The impact of cloud parameterization. *J. Climate*, **6**, 393–418, [https://doi.org/10.1175/1520-0442\(1993\)006<0393:CDACTI>2.0.CO;2](https://doi.org/10.1175/1520-0442(1993)006<0393:CDACTI>2.0.CO;2).
- Shimada, T., and S. Minobe, 2011: Global analysis of the pressure adjustment mechanism over sea surface temperature fronts using AIRS/Aqua data. *Geophys. Res. Lett.*, **38**, L06704, <https://doi.org/10.1029/2010GL046625>.
- Smith, R. L., A. Huyer, J. S. Godfrey, and J. A. Church, 1991: The Leeuwin Current off Western Australia, 1986–1987. *J. Phys. Oceanogr.*, **21**, 323–345, [https://doi.org/10.1175/1520-0485\(1991\)021<0323:TLCOWA>2.0.CO;2](https://doi.org/10.1175/1520-0485(1991)021<0323:TLCOWA>2.0.CO;2).
- Smith, S. R., P. J. Hughes, and M. A. Bourassa, 2011: A comparison of nine monthly air-sea flux products. *Int. J. Climatol.*, **31**, 1002–1027, <https://doi.org/10.1002/joc.2225>.
- Tan, I., T. Storelvmo, and M. D. Zelinka, 2016: Observational constraints on mixed-phase clouds imply higher climate sensitivity. *Science*, **352**, 224–227, <https://doi.org/10.1126/science.aad5300>.
- Tanimoto, Y., S.-P. Xie, K. Kai, H. Okajima, H. Tokinaga, T. Murayama, M. Nonaka, and H. Nakamura, 2009: Observations of marine atmospheric boundary layer transitions across the summer Kuroshio Extension. *J. Climate*, **22**, 1360–1374, <https://doi.org/10.1175/2008JCLI2420.1>.
- Terai, C. R., M. D. Zelinka, and S. A. Klein, 2016: Constraining the low-cloud optical depth feedback at middle and high latitudes using satellite observations. *J. Geophys. Res. Atmos.*, **121**, 9696–9716, <https://doi.org/10.1002/2016JD025233>.
- Tokinaga, H., Y. Tanimoto, and S.-P. Xie, 2005: SST-induced surface wind variations over the Brazil–Malvinas confluence: Satellite and in situ observations. *J. Climate*, **18**, 3470–3482, <https://doi.org/10.1175/JCLI3485.1>.
- , —, —, T. Sampe, H. Tomita, and H. Ichikawa, 2009: Ocean frontal effects on the vertical development of clouds over the western North Pacific: In situ and satellite observations. *J. Climate*, **22**, 4241–4260, <https://doi.org/10.1175/2009JCLI2763.1>.
- Trenberth, K. E., and J. T. Fasullo, 2010: Simulation of present-day and twenty-first-century energy budgets of the southern oceans. *J. Climate*, **23**, 440–454, <https://doi.org/10.1175/2009JCLI3152.1>.
- Tsushima, Y., and Coauthors, 2006: Importance of the mixed-phase cloud distribution in the control climate for assessing the response of clouds to carbon dioxide increase: A multi-model study. *Climate Dyn.*, **27**, 113–126, <https://doi.org/10.1007/s00382-006-0127-7>.
- van der Dussen, J. J., S. R. de Roode, S. D. Gesso, and A. P. Siebesma, 2015: An LES model study of the influence of the free tropospheric thermodynamic conditions on the stratocumulus response to a climate perturbation. *J. Adv. Model. Earth Syst.*, **7**, 670–691, <https://doi.org/10.1002/2014MS000380>.
- Wallace, J. M., T. P. Mitchell, and C. Deser, 1989: The influence of sea-surface temperature on surface wind in the eastern equatorial Pacific: Seasonal and interannual variability. *J. Climate*, **2**, 1492–1499, [https://doi.org/10.1175/1520-0442\(1989\)002<1492:TIOSST>2.0.CO;2](https://doi.org/10.1175/1520-0442(1989)002<1492:TIOSST>2.0.CO;2).
- Wang, L., Y. Wang, A. Lauer, and S.-P. Xie, 2011: Simulation of seasonal variation of marine boundary layer clouds over the eastern Pacific with a regional climate model. *J. Climate*, **24**, 3190–3210, <https://doi.org/10.1175/2010JCLI3935.1>.
- Weare, B. C., 2000: Near-global observations of low clouds. *J. Climate*, **13**, 1255–1268, [https://doi.org/10.1175/1520-0442\(2000\)013<1255:NGOOLC>2.0.CO;2](https://doi.org/10.1175/1520-0442(2000)013<1255:NGOOLC>2.0.CO;2).
- Wood, R., 2012: Stratocumulus clouds. *Mon. Wea. Rev.*, **140**, 2373–2423, <https://doi.org/10.1175/MWR-D-11-00121.1>.
- , and C. S. Bretherton, 2006: On the relationship between stratiform low cloud cover and lower-tropospheric stability. *J. Climate*, **19**, 6425–6432, <https://doi.org/10.1175/JCLI3988.1>.
- Wu, G., and Y. Liu, 2003: Summertime quadruplet heating pattern in the subtropics and the associated atmospheric circulation. *Geophys. Res. Lett.*, **30**, 1201, <https://doi.org/10.1029/2002GL016209>.
- Xu, H., S.-P. Xie, and Y. Wang, 2005: Subseasonal variability of the southeast Pacific stratus cloud deck. *J. Climate*, **18**, 131–142, <https://doi.org/10.1175/JCLI3250.1>.
- Yu, L., Z. Zhang, S. Zhong, M. Zhou, Z. Gao, H. Wu, and B. Sun, 2011: An inter-comparison of six latent and sensible heat flux products over the Southern Ocean. *Polar Res.*, **30**, 10167, <https://doi.org/10.3402/polar.v30i0.10167>.