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Key Points:
- Variability in tropical tropopause height imposes potential influence on variations of tropical cyclone (TC) intensity
- Weaker stratification of the lower stratosphere tends to correspond to an increase in the power dissipation index of intense TCs
- Large-scale impacts of the lower stratosphere on variability in TC intensity differ between ocean basins

Supporting Information:
- Supporting Information S1

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Large-scale control of the lower stratosphere on variability of tropical cyclone intensity

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Abstract Recent studies of tropical cyclones (TC) have suggested intricate impacts of the lower stratosphere layer (LSL) on TC development at the high-intensity limit. This study examines the potential realization of the impacts of the LSL interannual variability on TC intensity. By minimizing the effects of sea surface temperature and outflow temperature based on the potential intensity framework, partial correlation analyses show a negative correlation between the power dissipation index (PDI) for intense TCs and the tropopause height in the northwestern Pacific basin but a weaker relationship in the north Atlantic basin. Similar analyses for the LSL stratification also reveal signals of negative correlations between the LSL stratification and PDI in both basins, corroborating recent modeling studies about the impacts of the LSL on TC development. Due to the complexity of the relationships and data limitations, however, all correlations are statistically insignificant, thus rendering the impacts of LSL on TC intensity inconclusive at present.

Plain Language Summary This study presents analyses of the statistical relationship between tropical cyclone intensity and the interannual variability of the lower stratosphere. By isolating the dominant impacts of sea surface temperature and outflow temperature, it is shown that there is a signal of inverse relationship between the tropopause height and the power dissipation index (PDI) for intense TCs. This inverse relationship is however more apparent in the northwestern Pacific basin than in the North Atlantic basin. Likewise, similar inverse relationship is seen between the lower stratospheric stratification and the PDI, corroborating recent modeling studies about the impacts of the LSL on TC development. While all relationships are not statistical significance at present due to the multivariate nature of the PDI and limited data record, the results presented in this study suggest that the long-term variability of the lower stratosphere could be a potential factor influencing the variability of TC intensity that has not been fully understood.

1. Introduction

The variability in tropical cyclone (TC) intensity is sensitively influenced by their intricate interaction with large-scale environments such as sea surface temperature (SST), vertical wind shear, humidity of the lower and middle troposphere, or dry-air intrusion [e.g., *Gray*, 1968; *Goldenberg et al.*, 2001; *Mann and Emanuel*, 2006; *Wang et al.*, 2010; *Murakami et al.*, 2011]. Despite extensive research and study of these large-scale factors, a consensus about future projections of TC activity remains elusive due to our inadequate understanding of multiscale physical processes and associated nonlinear feedbacks [*Knutson et al.*, 2007, 2010; *Oouchi et al.*, 2006; *Bengtsson et al.*, 2007; *Hill and Lackmann*, 2011; *Emanuel et al.*, 2013]. In addition, significant uncertainty associated with inhomogeneous data, different baseline periods used to define the climatology, or different metrics used to quantify the statistical trends also contributes to the difficulty of quantifying the TC intensity variability [*Landsea*, 2007]. At present, most modeling studies of TCs under different future climate change scenarios suggest an overall decrease in TC frequency but an increase in the frequency of the most intense TCs (Category 4 and higher) [e.g., *Knutson et al.*, 2007, 2013; *Bender et al.*, 2010; *Vecchi et al.*, 2013].

Among the many factors that can influence the climatology of TC intensity, the impacts of SST have been most extensively examined. Modeling and observational studies such as those by *Knutson et al.* [1998], *Emanuel* [2005], and *Ramsay* [2013] have shown that global and regional trends in SST are the most important influence on TC intensity variability in the North Atlantic (NATL) basin. In contrast, the northwestern Pacific (WPAC) basin has a somewhat less certain connection between TC intensity and SST variability likely due to the opposing influences of the SST warming trend and adverse atmospheric conditions for TC development [*Lin and Chan*, 2015; *Emanuel et al*., 2013, hereafter E13].

Recent studies by E13, *Ramsay* [2013], *Wang et al*. [2014], and *Wing et al*. [2015, hereinafter W15] have revealed, however, a possible role of the tropical tropopause layer in modulating the climatology of TCs. Their studies found that tropical tropopause layer cooling is correlated to recent increases in TC potential intensity due to increased thermodynamic efficiency associated with a larger difference between SST and outflow temperature in the NATL basin. These results suggest that upper level conditions such as tropopause temperature could be a valuable predictor of TC activity, especially in the NATL basin where the interannual variability of the tropical tropopause layer could account for 30–70% of the potential intensity variation (W15).

Except for the direct influence of tropopause temperature on TC variability derived from Emanuel's potential intensity (PI) theory, the tropopause and the lower stratosphere layer (LSL) are often considered as a passive lid that plays a little dynamical role in TC development. Recent real-time intensity forecasts in the WPAC basin along with aircraft observations, however, have captured a peculiar structure for many intense TCs, suggesting possible impacts of the lower stratosphere on TC structure and development beyond the PI framework [*Kieu and Tallapragada*, 2014; *Ohno and Satoh*, 2015; *Stern and Doyle*, 2016; *Kieu et al*., 2016]. Specifically, at Category 3 and above, the TC inner core often develops a secondary warm core near the tropopause (~100 hPa) along with a thin layer of upper level inflow in the LSL, thus forming a double-warm core structure in TC inner core at the high-intensity limit. The development of this upper level inflow and the double-warm core structure is frequently followed by another episode of intensification. While the existence of these features is not conclusively verified due to the rarity of observations of TC inner-core above 100 hPa, the fact that the development of a double-warm core structure is always accompanied by an upper level inflow above the typical outflow level indicates that the LSL may play a role in the development of intense TCs that has not been previously addressed.

Given the potential importance of the LSL in the development of intense TCs, the objective of this study is to quantify the extent to which the stratification in the lower stratosphere (N2) and the tropical tropopause height (TTH) can affect the variability of the TC intensity. Our hypothesis is that a lower TTH or a weaker N2 favors the formation of an upper level warm core and an inflow in the LSL, thus allowing TCs to further strengthen beyond the PI limit. This hypothesis has been recently examined in a set of idealized experiments by *Moon and Kieu* [2016], which showed noticeable impacts of the LSL on the development of intense TCs in the WPAC basin. To what extent this feature is realized in long-term observational data is unknown and, therefore, presented here.

2. Data and Methods

2.1. Input Data

Because the main focus of this study is on the interannual variability of the tropopause and lower stratosphere characteristics, it is important to have a data set that contains sufficient information about the tropopause structure and its long-term variability. As a result, the National Centers for Environmental Prediction (NCEP)/National Center for Atmospheric Research (NCAR) reanalysis [*Kalnay et al*., 1996] data are used for our analysis. This data set consists of three-dimensional distributions of geopotential height and temperature at standard pressure levels, as well as two-dimensional distributions of tropopause pressure and temperature, recorded four times daily from 1948 to 2015 at a spatial resolution of $1 \times 1°$. Because the NCEP reanalysis may have potential issues with tropical tropopause layer temperature trends [e.g., *Randel et al*., 2009, E13; *Vecchi et al*., 2013], the interim European Center Medium-Range Forecast Reanalysis (ERA-Interim) [*Dee et al*., 2011] from 1979 to 2015 are also employed to complement the NCEP/NCAR reanalysis. While the length of the ERA-Interim data set is relatively short as compared to the NCEP reanalysis, both data sets contain pertinent information for the calculation of TTH and its related interannual variability that this study examines.

Regarding the TC data, the best track data set provided by Joint Typhoon Warning Center in the WPAC basin and the most updated HURricane DATabases [*Landsea et al*., 2014] data set in the NATL basin are used, which contain all required information about the TC intensity from 1948 to 2015. For sea surface temperature (SST), the monthly mean data from the Hadley Centre Sea Ice and Sea Surface Temperature data set [*Rayner et al*., 2003] are used such that the correlation analysis between power dissipation index (PDI) and SST is consistent with *Emanuel* [2005].

2.2. Tropopause Height Calculation

While tropopause temperature and height are often inversely related, practical calculations of the TTH based on the hydrostatic equation require a temperature profile of the entire troposphere rather than just the tropopause temperature. In this study, we follow *Reid and Gage* [1981], *Wu and Pauluis* [2013], and *Fueglistaler et al.* [2009] and define TTH using the conventional definition of the thermodynamic tropopause as the point above the 500 hPa level at which (i) the temperature lapse rate decreases to 2 K km^{-1} or less and (ii) the temperature is lowest.

Because neither the NCEP/NCAR reanalysis nor ERA-Interim explicitly provides TTH, we examine tropopause height variations by using the thickness of the 1000 hPa tropopause layer derived from the hydrostatic equation as follows:

$$h_t = z_{1000} - \frac{R}{g}\int_{1000}^{p_t} \frac{T}{p}\mathrm{d}p, \tag{1}$$

where z_{1000} is geopotential height at level p = 1000 hPa, p_t is the tropopause pressure, and T is the absolute temperature on the standard pressure levels. Although h_t is not exactly TTH, it suffices to characterize variability of TTH on the seasonal time scale because we are concerned more with interannual variations in the basin-wise average of tropopause height instead of its absolute value. Verification of interannual variability of h_t and the actual TTH output from the NCEP Final Analysis (FNL) confirms indeed the consistency of h_t in describing variability of the TTH during 1997–2015 period where the FNL data are available (see Figure S1 in the supporting information). This consistency gives us confidence regarding the use of h_t to extrapolate TTH variability back to 1948 in the NCEP/NCAR reanalysis for the longest data history possible. Note that because the ERA-Interim lacks information on tropopause temperature and tropopause pressure, we use an additional diagnostic to search for the pressure level of the tropopause and the corresponding tropopause temperature at each grid point, using the aforementioned definition of the tropopause. This procedure is applied for all grid points, resulting in two-dimensional distributions of the tropopause pressure p_t and tropopause temperature necessary for the tropopause height calculation as given by equation (1).

2.3. Correlation Analysis

Analysis of the statistical relationship between TC intensity and TTH is challenging because TC intensity is a complex function of various environmental factors and because of the discrete and intermittent nature of TC intensity variability. There are a number of different metrics that can be used to characterize the variability of TC intensity such as Emanuel's theoretical PI limit (V_{PI}), the accumulated cyclone energy index, the power dissipation index (PDI), or the observed maximum 10 m surface wind speed (V_m). In this study, the variability of the TC intensity is examined in terms of the PDI, which can effectively represent not only the strength of a TC but also their lifetime and frequency. By definition, the PDI is computed as

$$\mathrm{PDI} = \int_0^T {V_m}^3 \mathrm{d}\tau, \tag{2}$$

where T is the lifetime of a TC. As mentioned in *Emanuel* [2005], PDI is a better indicator of TC threat than TC frequency or V_m alone. In addition, the cubic power allows for a sharper representation of the extreme intensities that this study wishes to explore.

Assuming that PDI is proportional to V_{PI}^3, the PI framework indicates that PDI should be strongly dependent on SST, outflow temperature, and the enthalpy exchange at the ocean surface. The latter is in turn a function of wind speed and enthalpy disequilibrium at the ocean surface. With potential impacts of tropopause height on the TC inner-core structure as suggested from recent modeling studies, we hypothesize that the PDI of intense TCs should depend further on tropopause height beyond the outflow temperature or SST. To minimize the influence of the outflow temperature as well as the direct impacts of SST on the PDI of intense TCs, the contribution from these two dominant variables will first be subtracted from the total PDI as follows:

$$\mathrm{RPDI} = \ln(\mathrm{PDI}^*) - \beta_1 \ln(\mathrm{SST} - T_0) - \beta_2 \ln(T_0), \tag{3}$$

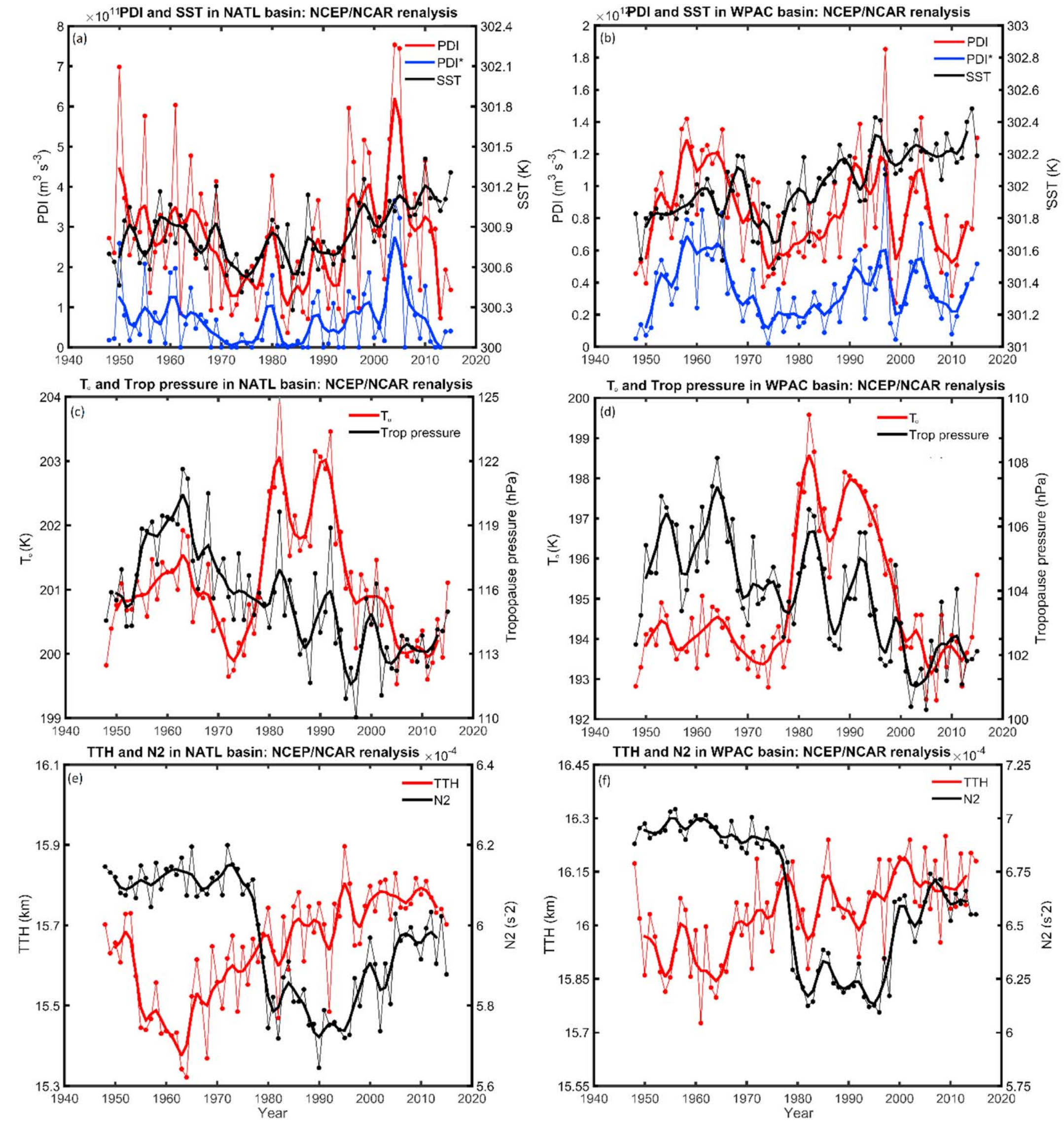

Figure 1. Time series of (a and b) the PDI (red dots), intense PDI (blue dots), and SST (black dots), (c and d) tropical tropopause temperature T_0 (red) and tropopause pressure (black), and (e and f) tropopause height TTH (red) and the lower stratospheric stratification N2 (black) during 1948–2015 for the North Atlantic (Figures 1a, 1c, and 1e) and the WPAC basin (Figures 1b, 1d, and 1f). Solid lines denote smoothed curves, which are obtained by applying a 1-2-1 moving-average operator to the time series.

where RPDI denotes the natural logarithm of the residual PDI of intense TCs, T_0 denotes the tropical tropopause temperature that approximates the outflow temperature, β_1 and β_2 are regression coefficients, and PDI* is the PDI for intense TCs defined as follows:

$$\mathrm{PDI}^* = \int_{T_1}^{T_2} {V_m}^3 \mathrm{d}\tau, \tag{4}$$

where $[T_1, T_2]$ is the time interval during which a TC reaches Category 4 and above (see Figure S2 in the supporting information for time series of cycles that TCs reached Category 4 and above during the 1948–2015 period). The basis for the RPDI calculation given by equation (3) follows from Emanuel's PI limit in which V_{PI} is proportional to (SST-T_o)/ T_o. As such, a natural log transforms the V_{PI} into a linear form, thus facilitating the linear regression.

Note that the PDI* given by equation (4) is evaluated only for intense TCs of Category 4 and above, because it is expected that TCs will most effectively interact with the LSL at the high-intensity limit. On average, PDI*accounts for about 30% of the total PDI during 1948–2015 in the NATL basin and about 40% in the WPAC basin (cf. Figure 1). While the RPDI derived from PDI* as given by equation (3) does not completely remove the influence of SST (as SST also enters through the enthalpy disequilibrium term in the V_{PI} formulation), it minimizes the direct influence of SST. The partial correlation analysis between RPDI and TTH is thus expected to better expose the functional dependence of PDI on TTH than analyses that directly use the raw PDI given by equation (2).

Along with the tropopause height, the atmospheric stratification of the lower stratosphere is another measure that could well represent the variability of the lower stratosphere. In this study, the lower stratospheric stratification (N2) is defined as follows:

$$N^2 = \frac{1}{\theta}\frac{\partial\theta}{\partial p}, \tag{5}$$

for the layer from 100 to 75 hPa, where θ is potential temperature and p is pressure. Given the typical location of the tropopause around 125 hPa–100 hPa level in the tropical region, the temperature structure in the 100–75 hPa layer is expected to capture major interannual signals of the lower stratospheric stratification on the basin scale. Similar to the correlation analysis between RPDI and TTH, all partial correlations of RPDI with N2 are analyzed using both the NCEP/NCAR reanalysis during 1948–2015 and the ERA-Interim during 1979–2015 period. To ensure the large-scale representativeness, all time series of N2 and TTH are computed by taking a basin-wise average over fixed domains: [6–18°N] × [20–60°W] for the NATL basin and [5–15°N] × [130–180°E] for the WPAC basin, similar to W15's study.

3. Results

3.1. Interannual Variability of the Tropical Tropopause

To first establish the interannual variability of TC intensity and the LSL characteristics, Figures 1 and 2 show time series of PDI, SST, tropical tropopause temperature T_0, and TTH for both the WPAC and the NATL basins that are obtained from the NCEP/NCAR and ERA-Interim reanalyses. Consistent with *Emanuel* [2005], there is a strong correlation between PDI and SST in the NATL basin, which is most robust during the 1950–2004 period, thus suggesting a dominant control of SST on the TC intensity in the NATL basin. In the WPAC basin, the correlation between SST and PDI is noticeably weaker, despite substantially warmer SST and higher PDI (see the supporting information for the correlation analysis between SST and PDI for the NATL and the WPAC basins). As discussed in *Wang and Chan* [2002], *Murakami et al*. [2011], and *Lin and Chan* [2015], the weaker SST-PDI relationship in the WPAC is likely due to influence of inimical atmospheric factors such as stronger vertical wind shear or larger tropospheric stratification, which appear to offset the effects of warmer SST in this ocean basin. Note that the PDI index includes not only the TC intensity but also the TC lifetime and frequency. Therefore, the PDI-SST correlation is not just a function of SST but also depends on other large-scale factors that govern the TC frequency as discussed in *Bengtsson et al*. [2007], *Camargo et al.* [2009], *Murakami et al*. [2011], and *Lin and Chan* [2015].

Of specific importance to this study are long-term variations of the LSL characteristics including tropopause temperature T_0, tropopause height TTH, and the lower stratospheric stratification N2. It is intriguing to observe that both the NATL and WPAC basins experience an increase in T_0 between 1970 and 1980 in the NCEP reanalysis, which corresponds to a large dip in both PDI and PDI*, albeit the variations are not entirely in phase. Such a surge of warming tropopause is not seen in ERA-Interim, which is available only after 1979 (Figure 2), and it is therefore hard to conclude if this warming surge in the NCEP/NCAR reanalysis data is due to changes in the global model, the NCEP data assimilation system, or it indeed captures a real shift in climate during this period.

Along with the anomalous tropopause warming during 1970–1980, it is noted that there appears to be no noticeable correlation between TTH and T_0 in either basin despite the expectation that higher TTH would correspond to lower T_0 (the exception being the period from 1990 to 2015 during which the correlation appears to be slightly inverse in the NATL basin). The weak relationship between TTH and T_0 is also seen in

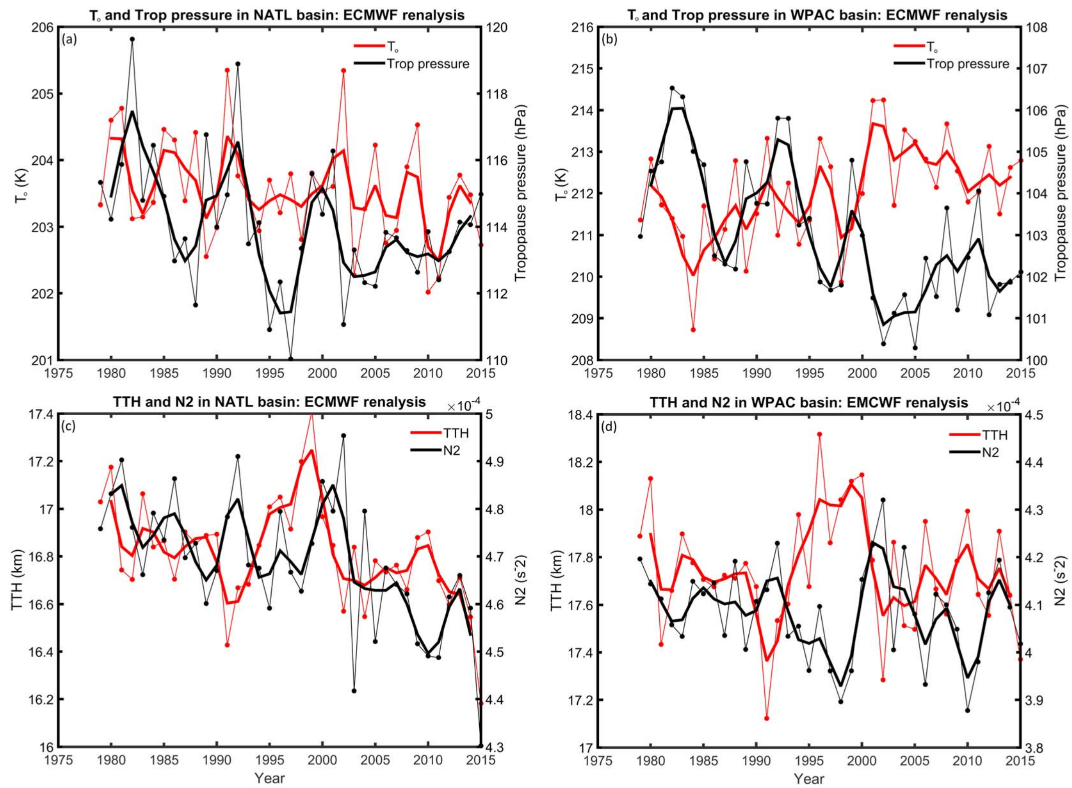

Figure 2. Similar to Figures 1c and 1f but obtained from the ERA-Interim reanalysis during 1979–2015 period.

ERA-Interim from 1979 to 2015, which is attributed to the fact that TTH is determined not only by the tropopause temperature but also by the entire tropospheric temperature layer as shown via the hydrostatic equation (1). Although such a weak TTH-T_0 relationship is not conclusive from either NCEP/NCAR (partial correlations of ~0.21 for the NATL and ~0.09 for the WPAC at 95% significance level) or ERA reanalysis (partial correlations of about 0.1 for the NATL and 0.23 for the WPAC at 95% significance level) due to the different characteristics of the reanalysis data set, the consistent behavior of area-averaged T_0 and TTH in both ocean basins indicates that TTH variation is not a simple function of T_0 as often assumed in previous studies.

Unlike the weak TTH-T_0 connection, T_0 and N2 show stronger inverse correlations in both NATL and WPAC ($|r| > 0.80$ in both ocean basins with the NCEP/NCAR reanalysis and roughly -0.53 and -0.30 in the NATL and WPAC basins with the ERA reanalysis). Specifically, the surge in T_0 corresponds to a large decrease of N2 during the period from 1970 to 2000, which is evident in both ocean basins (cf. Figures 1c and 1d and 1e and 1f). This negative correlation between T_0 and N2 is consistent with the basic thermodynamic structure of the upper troposphere, i.e., an increase in tropopause temperature will directly reduce the temperature difference between the tropopause level and the level above it, assuming the same temperature of the level right above the tropopause. As such, the warmer T_0 would imply a smaller temperature difference in the LSL, thus explaining the weaker N2 as seen in Figures 1c–1f and 2,

Given the weak correlation between T_0 and TTH, and between the TTH and N2, it is thus reasonable to assume that TTH and N2 can be considered as independent predictors of TC intensity beyond the traditional variables of SST and outflow temperature T_0, for which our partial correlation analyses turn next.

3.2. Influence of Tropopause Height on TC Intensity

Although the strong correlation between the overall PDI and the PDI for intense TCs seen in Figures 1a and 1b implies that any influence of tropopause level on intense TCs should be apparent in all TCs as well, this section presents long-term variability of PDI and tropopause height only for intense TCs to isolate the signal of the tropopause variability on the most intense TCs. All of the results and discussion below that mention

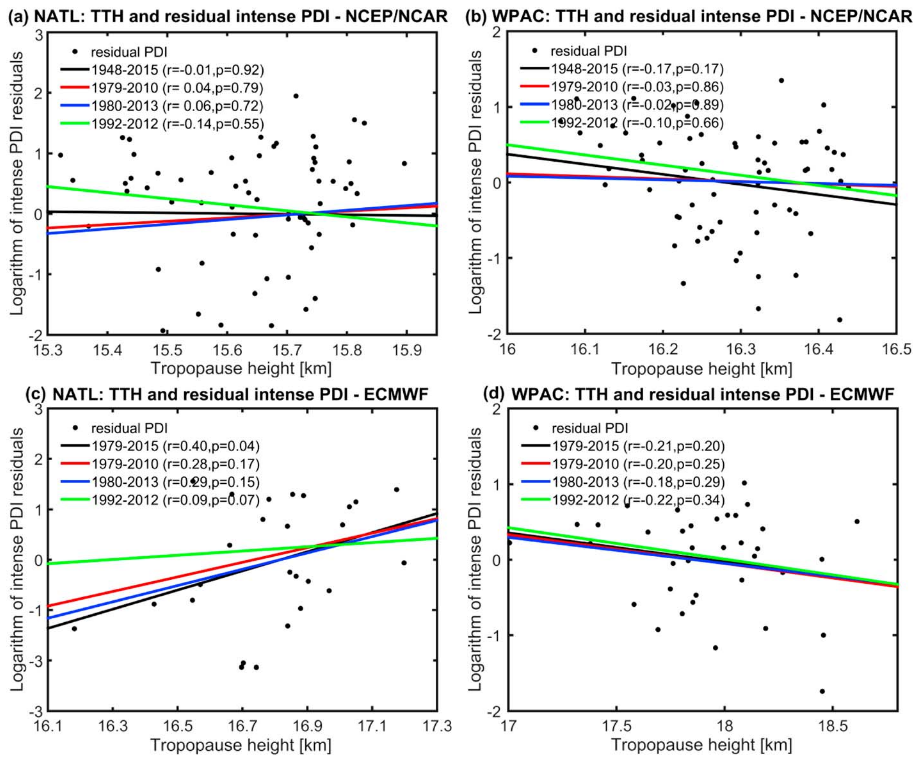

Figure 3. Partial correlation of the tropopause height (TTH, unit: km) and the natural logarithm of PDI for intense TCs after removing the effects of SST and the tropical tropopause temperature T_0 for (a) the NATL basin and (b) the WPAC basin, using four different baseline periods 1948–2015 (black), 1979–2010 (red) [*Emanuel et al.*, 2013], 1992–2012 (green) [*Lin and Chan*, 2015], and 1980–2013 (blue) [*Wing et al.*, 2015]. Black dots denote all data from 1948 to 2015. The correlation coefficients *r* for all periods are indicated in each panel. (c and d) Similar to Figures 3a and 3b but obtained from the ERA-Interim during the 1979–2015 period.

"correlation" refer to the partial correlation of PDI after reducing the direct effects of SST and T_0 based on equation (3), which we refer to as residual PDI (RPDI).

Figure 3 shows scatterplots of RPDI versus TTH obtained from the NCEP/NCAR reanalysis from 1948 to 2015 and ERA-Interim from 1979 to 2015 for the NATL and the WPAC basins, along with linear regressions obtained from several different baseline periods in previous studies including 1979–2010 [*Emanuel*, 2003], 1992–2012 [*Lin and Chan*, 2015], and 1980–2013 (W15). Except for a weak negative correlation for the 1948–2015 period, one notices in Figures 3a–3c that there is essentially no significant RPDI-TTH correlation in the NATL basin with both the NCEP/NCAR reanalysis and ERA-Interim data, regardless of the baseline periods. While none of these correlations are statistically significant (see the *p* values reported in Figure 3), this result suggests that the strongest signal of the interannual variability of the TC intensity in the NATL basin appears to be attributed to the dominant factors SST and T_0 as found in previous studies [*Emanuel*, 2005; W15]. Thus, further inclusion of the tropopause height does not add explanatory value to the interannual variability of TC intensity in the NATL basin.

In the WPAC basin, there is, however, a more consistent negative correlation between TTH and RPDI obtained from both the NCEP/NCAR reanalysis and the ERA-Interim, i.e., the lower the tropopause height, the higher the RPDI (Figures 3b and 3d). This negative correlation is seen for all baseline periods, with the strongest relationship being for the entire 1948–2015 period. At face value, the above negative TTH-RPDI correlation implies that the lower tropopause height would correspond to either higher intensity or more prolonged period of high-intensity stages. Although no correlation is statistically significant in either reanalysis data set, such a modest dependence of the TC intensity on the tropopause height is consistent with recent modeling study of the TC-lower stratosphere interaction in the WPAC basin by *Moon and Kieu* [2016], which captured an increase of the TC intensity when lowering the tropopause level in their idealized simulations. Physically, a

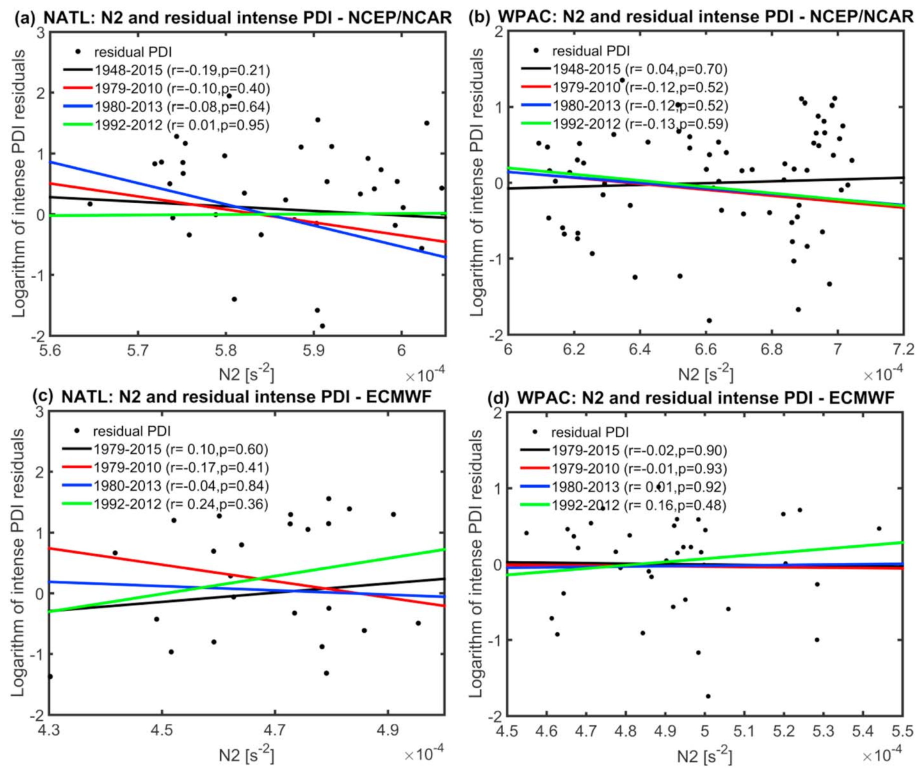

Figure 4. Similar to Figure 3 but for the atmospheric stratification of the lower stratosphere layer between 100–75 hPa (N^2, unit: s^{-2}) and the natural logarithm of the residual PDI for intense TCs.

lower TTH would imply two direct consequences: (i) a change in the tropospheric stratification (assuming that the outflow temperature and SST are kept unchanged) and (ii) a change of the LSL stratification. For example, a stronger tropospheric stratification tends to suppress deep convection and inhibits TC development [*Hill and Lackmann* 2011; *Lin and Chan*, 2015], whereas the weaker LSL stratification allows TCs to more effectively interact with the LSL. Idealized simulations by *Moon and Kieu* [2016] suggest that the interaction of TCs with the LSL appears to play a more important role in development of the double-warm core structure in intense TCs in the WPAC basin than that in the NATL basin, thus accounting for the more frequent development of intense TCs in the WPAC basin. In this regard, the negative TTH- RPDI correlation captured in the WPAC basin could be a manifestation of the importance of the LSL that has not been well understood.

If the role of the LSL in modulating the PDI for intense TCs is considerable, one would expect to also see realizable impacts of the LSL stratification on the long-term variability of the PDI for intense TCs, as will be next discussed.

3.3. Impacts of the Lower Stratosphere Stratification

Physically, the tropopause level examined in the previous section simply indicates the elevation at which TCs can interact with the LSL, i.e., a variation of the tropopause level can affect the effectiveness of the TC-LSL interaction. How this interaction takes place should, nonetheless, depend further on the stratification of the lower stratosphere. With the noticeable negative correlation between TTH and RPDI in the WPAC basin, it is natural to expect that the LSL stratification could also affect the RPDI variability in this basin.

Figure 4 shows the correlation analysis between N2 and RPDI obtained from the NCEP/NCAR reanalysis and ERA-Interim together with three other baseline periods similar to those in Figure 3. With the exception of the WPAC basin for the entire 1948–2015 period, both ocean basins show a generally negative correlation between RPDI and N2, which varies from -0.02 to -0.19 for all baseline periods. While these correlations have *p* values that range widely, the consistency of the sign of the relationship between the two basins and among different baseline periods as obtained from the NCEP/NCAR reanalysis suggests that weaker

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LSL stratification tends to correspond to stronger TC intensity as expected from the interaction of the LSL and intense TCs. It should be noted however that the signal of the negative RPDI-N2 correlation is not as clear from the ERA-Interim as from the NCEP/NCAR reanalysis (Figures 4c and 4d), which varies strongly with p values between 0.24 and 0.93. Although one cannot reject the null hypothesis of no correlation between RPDI and N2, a mechanism underlying the potential LSL-TC interaction seems to be related to how intense TCs could induce an upper level inflow in the LSL and enhance the warm core structure of TCs as presented in *Kieu et al*. [2016]. Specifically, the ability to induce an upper level inflow would depend on the LSL stratification, which can affect the TC intensity at the high-intensity limit. Of course, details of this mechanism cannot be simply deduced from the statistical analysis of the variability of the PDI and N2. The results presented here, however, provide a consistent picture with those obtained from model simulations that indicate the possible role of the LSL that has not been previously examined.

It should be noted that the negative RPDI-N2 correlation may appear at odds with the inverse relationship between V_{PI} and T_0 based on Emanuel's PI theory (E13 and W15). This contrast can be seen directly in Figure 1, which shows a negative correlation between $T_0$and N2. If one takes this inverse N2-T_0relationship into account, then the warmer T_0would correspond to weaker N2 and consequently higher PDI according to the negative RPDI-N2 correlation, whereas W15's results show that warmer T_0 would correspond to weaker V_{PI}. This conundrum is explainable if we recall that PDI represents the TC actual intensity computed from the best track, whereas V_{PI} is the theoretical limit derived from a given thermodynamic condition. Furthermore, the real-data PDI integrates not only information about the TC intensity but also their lifetime and frequency and possibly other implicit factors such as vertical wind shear or tropospheric stratification that are not accounted for by the potential intensity theory. As a result, a higher V_{PI} is not always translated to higher PDI and vice versa.

4. Discussion and Conclusion

In this study, relationships among the power dissipative index (PDI) for intense TCs, the tropopause height (TTH), and the stratification of the lower stratosphere (N2) were analyzed for the WPAC and NATL basins, using the NCEP/NCAR and ERA-Interim reanalyses. By minimizing the dominant dependence of the PDI for intense TCs on sea surface temperature (SST) and the outflow temperature, it was found that there is a signal of a negative correlation between the residual PDI for intense TCs (RPDI) and tropopause height in the WPAC basin. Although the p values for the RPDI-TTH correlation are too large to be conclusive, this signal of the negative correlation between RPDI and tropopause height in the WPAC basin suggests that lower tropopause height would allow TCs to more effectively interact with the LSL, thus implying more intense TCs. In the NATL basin, the RPDI-TTH correlation is less apparent, possibly due to the cooler SSTs in this ocean basin that prevent TCs from attaining a high-intensity limit (Category 4 and above) necessary for TCs to interact with the lower stratosphere.

Further examination of the variability of the RPDI and N2 showed again a modest negative correlation between these two variables, which suggests that a weaker LSL stratification would correspond to either higher-intensity or more strong storms. Unlike the RPDI-TTH relationship, the negative correlation between RPDI and N2 is fairly consistent in both the WPAC and the NATL basins regardless of the time periods used to calculate the correlation, albeit statistical significance is again not achieved. Given the expectation of the negative correlation among N2 and RPDI, one could in principle conduct one-tailed hypothesis testing that could improve the statistical significance level of their correlation. Our examination of these one-tailed testings showed that the p values are still fairly large for all time periods, and so the statistical relationship between these variables is indeed not significant with the current data set. While the correlations between RPDI and TTH or N2 obtained, so far, from both the NCEP/NCAR reanalysis and ERA-Interim do not meet the usual thresholds for statistical significance due to the relatively short time series and large interannual variability of the PDI index, the relationships between these variables appear to be qualitatively consistent with the recent cloud-resolving modeling studies of TC development at the high-intensity limit, at least in the WPAC basin.

That the interaction between intense TCs and the LSL appears to be more strongly manifested in the WPAC basin than the NATL basin raises an important question regarding physical processes underlying this difference. At present, such opposing statistics is likely related to several differences in large-scale conditions between these two basins such as SST, vertical wind shear, tropospheric stratification, or midtroposphere moisture contents. In fact, E13 speculated that the NATL basin is governed more by thermodynamic conditions such as SST or tropopause cooling than in other ocean basins. For example, the 2–3 K colder SST in the NATL basin may result in less intense TCs, thus preventing the NATL storms to interact efficiently with the LSL and to further strengthen as compared to the WPAC basin. Such a basin-wise characteristic is in accord with more frequent emergence of the double-warm core structure in the WPAC basin as reported in *Kieu et al*. [2016], whereas the double-warm core structure is rarely seen in the NATL basin despite the use of the same forecasting model. These differences in large-scale conditions account for different relationships not only between RPDI and TTH or RPDI and N2 but also other relationships such as the SST-PDI relationship [e.g., *Lin and Chan*, 2015] or the PDI-T_0 relationship (E13 and W15).

The fact that both European Centre for Medium-Range Weather Forecasts and NCEP/NCAR reanalysis products yield a null statistical result regarding the impacts of TTH or LSL stratification on PDI of intense TCs may point to potential issues either with the data sets or the effectiveness of the (linear) partial correlation analysis within the PI framework presented in this study. This issue is generally difficult to address due to large uncertainties in the lower stratosphere variability in the current reanalysis data sets. However, the consistency among recent airborne upper level observations in the TC inner-core region, modeling studies, and analyses of the variability of PDI with TTH and N2 in this study suggests that variations in tropopause height and lower stratospheric stratification may be important climatological factors influencing the interannual variability of TC intensity. In this regard, these factors should be taken into account when examining long-term intensity variation of intense TCs.

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