Aspects of interannual and intraseasonal variability of the tropopause and lower stratosphere

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SUMMARY

Radiosonde and National Centers for Environmental Prediction/National Center for Atmospheric Research reanalysis data are utilized to consider aspects of large-scale variability in tropopause height, temperature and pressure. This variability is related to coherent dynamical fluctuations in the troposphere and lower stratosphere through the use of linear correlation and regression. On interannual time-scales, significant global-scale tropopause fluctuations are tied to variability in sea surface temperature (SST) associated with the El Niño/Southern Oscillation phenomenon. When SST is anomalously high in the central tropical Pacific, tropopause height (pressure) is high (low) throughout the Tropics, with largest perturbation amplitudes in the subtropical Pacific. At the same time, the tropopause is cold over the tropical and subtropical Pacific sector but warm elsewhere in the Tropics. Over the extratropics, wave-like perturbations in the tropopause are seen, with anomalous cyclonic flow corresponding to a lower tropopause height and higher tropopause temperature and pressure, and vice versa. The sign of the temperature anomalies in the lower stratosphere tends to match that at the tropopause over much of the globe, with opposite-signed anomalies in the upper troposphere. The vertical structure of these perturbations is consistent with the expected potential-vorticity anomalies induced by quasi-stationary Rossby waves and vertically propagating gravity waves forced by displacements of tropical convection. Similar relationships are associated with the eastward propagation of tropical convection due to the Madden–Julian Oscillation on intraseasonal time-scales.

KEYWORDS: Interannual variability Intraseasonal variability Stratosphere Tropical convection Tropopause

1. INTRODUCTION

Despite extensive study, the tropopause remains one of the most enigmatic features of the atmosphere. While many aspects of the tropopause are understood, at least in principle, a comprehensive explanation of its detailed climatological distribution and variability is still lacking (Thuburn and Craig 1997, 2000). Documentation of the detailed behaviour of the tropopause is crucial to a more complete understanding of the cycling of atmospheric constituents, since much of the slow, steady exchange of air and trace chemicals between the troposphere and stratosphere on seasonal time-scales is now believed to take place through the tropical tropopause (e.g. Rosenlof and Holton 1993; Holton et al. 1995; Mote et al. 1995, 1996; Tuck et al. 1997). At higher frequencies, more episodic stratosphere–troposphere exchange is known to occur within deep tropical convection (e.g. Danielsen 1993; Reid and Gage 1996) and in regions of ‘cut off’ circulations, tropopause folding, and other synoptic-scale events in the extratropics (e.g. Shapiro 1980).

One recent development aiding the study of the tropopause is the availability of objective tropopause analyses from the National Centers for Environmental Prediction (NCEP) and National Center for Atmospheric Research (NCAR) reanalysis project (Kalnay et al. 1996). The NCEP/NCAR reanalysis contains grids of tropopause temperature and pressure at the same global spatial and temporal resolution as the other analysed parameters. While these tropopause fields are considered to be directly observed quantities (so-called ‘class A’ variables), they are nevertheless significantly influenced
by the first-guess model output in regions where there are few radiosonde data, especially in the period before satellite measurements became available. Tropopause characteristics have also been derived from analyses of temperature and pressure using European Centre for Medium-Range Weather Forecasts (ECMWF) data (Hoerling et al. 1991; Hoinka 1998, 1999; Highwood and Hoskins 1998; Simmons et al. 1999). Overall, it has been found that these objective analyses are extremely useful, especially when statistical processing results in the removal of at least some of the systematic biases and unrealistic random fluctuations inherent in these data.

We are here concerned with certain aspects of interannual and intraseasonal variations in the global tropopause that can be tied to large-scale dynamical forcing by tropical convection. Since this variability also appears as fluctuations within the troposphere and lower stratosphere, we include, where appropriate, analyses of associated perturbations in these regions as well.

At interannual time-scales, variability in the tropical tropopause and lower stratosphere is known to be linked to both the quasi-biennial oscillation, or QBO (e.g. Angell and Korshover 1964; Reid and Gage 1985; van Loon and Labitzke 1987; Randel et al. 2000) and the solar cycle (e.g. Gage and Reid 1981; van Loon and Labitzke 1998). By far the largest signal, though, is that due to warm and cold events of the El Niño/Southern Oscillation (ENSO) phenomenon (Quiroz 1983; Reid and Gage 1985; Gage and Reid 1986, 1987; van Loon and Labitzke 1987; Reid et al. 1989; Reid 1994; Yulaeva and Wallace 1994; Randel et al. 2000). We examine the behaviour of such ENSO variability in section 3. In section 4, we show that over the west Pacific warming pool, important tropopause fluctuations occur in association with the Madden–Julian Oscillation (MJO). The spatial structure of this variability is planetary in scale, and so we include an examination of accompanying signals in the extratropics as well.

2. DATA AND METHODOLOGY

Our approach relies heavily on records of radiosonde ascents obtained from NCAR, which are used to obtain information on the detailed vertical structure of temperature, pressure and wind at various tropical stations. The vertical resolution of these data varies widely over time and from station to station, but in general is much better than the reanalysis data. Some of these records are quite long, extending back to the early 1950s in the case of some Micronesian stations, and are nearly complete, with less than 5% of the observations reported as missing. Despite drawbacks in the quality of these data due to instrument changes and other biases (Gaffen et al. 2000), radiosonde data are by far the most reliable means for studying the climatology and variability in tropopause characteristics, as recently discussed by Randel et al. (2000) and Seidel et al. (2001).

The NCEP/NCAR reanalyses provide four-times-daily grids at 2.5° horizontal resolution and 17 pressure levels in the vertical from 1000 to 10 hPa. The datasets also contain analyses of the tropopause temperature ($T_{trop}$) and tropopause pressure ($P_{trop}$) at the same resolution. Outgoing long-wave radiation (OLR), also on the same 2.5° grid as the reanalysis, is utilized as a proxy for deep tropical convection. These data were interpolated in time and space to fill in regions of missing observations (Liebmann and Smith 1996). The impact of such interpolation is minimal for statistical work, where the OLR has been used successfully to represent subseasonal anomalies in convection (e.g. Kiladis and Weickmann 1997).

The World Meteorological Organization definition of the tropopause is used to produce daily tropopause parameters from the radiosonde data. This lapse-rate definition locates the tropopause as 'the lowest level at which the lapse rate decreases to
2 degC km\(^{-1}\) or less, provided also the average lapse rate between this level and all higher levels within 2 km does not exceed 2 degC km\(^{-1}\). In addition to these criteria, the tropopause is not allowed to occur at pressures greater than 400 hPa or less than 85 hPa. This definition is also used to locate the tropopause in the reanalysis dataset (M. Kanamitsu, personal communication), except that in this case the lapse rates are derived from sigma-level data, and the \(T_{\text{trop}}\) and \(P_{\text{trop}}\) are obtained by interpolation. Other tropopause definitions have been used in past studies (see Hoerling et al. 1991; Seidel et al. 2001). In particular, the 'cold point' tropopause may relate more directly to the exchange of water vapour between the troposphere and stratosphere (Mote et al. 1996), and its location can differ substantially from the lapse-rate tropopause as used here (Seidel et al. 2001). The lapse-rate tropopause is used throughout this study, however, since this is the definition used in the NCEP/NCAR reanalysis dataset.

In order to isolate the large-scale features of the tropopause and lower stratosphere which are related to various parameters of the climate system, we utilize a cross-correlation and regression technique. This approach has been used in many previous studies to examine relationships between large-scale circulations and Pacific sea surface temperature (SST) on interannual time-scales (e.g. Wallace et al. 1998) and tropical convection on intraseasonal time-scales (e.g. Kiladis and Weickmann 1997). Briefly, the linear relationship between an independent variable, such as tropical Pacific SST, and various dependent variables, such as tropopause height or temperature, is established by regression and correlation. In the case of gridded analyses used as dependent variables, the regression is computed separately at each individual grid point. The correlation coefficient of the linear relationship is obtained along with a regressed value of the dependent variable for an arbitrary perturbation in the independent variable. The correlation coefficient can be used to determine the statistical significance of the linear relationship between the two variables, once the temporal auto-correlation of each is determined and used to calculate the effective degrees of freedom (Livezey and Chen 1983). In this study, a signal is considered to be significant if it tests at better than the 95% confidence level.

In previous studies the tropopause height \((Z_{\text{trop}})\) has also been analysed (Gage and Reid 1982, 1986, 1987; Reid and Gage 1985). For this study we have created a gridded \(Z_{\text{trop}}\) dataset, obtained by simple linear interpolation using the tropopause pressure and geopotential-height data from the reanalyses.

In Fig. 1, the daily time series of \(T_{\text{trop}}\) at Koror (7.3\(^{\circ}\)N, 134.5\(^{\circ}\)E) and at the nearest reanalysis grid point (7.5\(^{\circ}\)N, 135.0\(^{\circ}\)E) are plotted from 1 January 1966 to 31 December 1998 (note that the reanalysis series has been offset by \(-20\) degC compared to the radiosondes). A similar seasonal cycle is apparent in both datasets, and the overall impression is that the reanalysis captures much of the variability seen in the radiosonde data. The correlation between the two series is \(+0.68\), but the mean temperature in the reanalysis (\(-79.2\) \(^{\circ}\)C) is too high by 2.7 degC compared to the radiosonde mean (\(-81.9\) \(^{\circ}\)C).

Despite the biases in the reanalysis dataset, both series in Fig. 1 have identical daily standard deviations (3.04 degC), although the impression from the plot is that the radiosonde data have more day-to-day variability, which is confirmed once the seasonal cycle, interannual variability, and trends are removed by high-pass filtering the data with a 120-day cut-off. In that case the daily standard deviation of the radiosonde data drops to 2.15 degC, compared to 1.35 degC for the reanalysis, and the correlation falls to \(+0.45\). However, we note that there is no a priori reason to expect very high correlations between these two series, since the daily correlation of less than 120-day filtered radiosonde \(T_{\text{trop}}\) between Koror and Yap, 470 km away, is only \(+0.49\). This
Figure 1. Daily time series of tropopause temperature (°C) for the period 1966–98 from (top) radiosonde observations at Koror (7.3°N, 134.5°E), and (bottom) the nearest reanalysis grid point to Koror (7.5°N, 135°E). The reanalysis data have been offset by −20 degC for comparison.

indicates that there are substantial local day-to-day fluctuations of the tropopause over the warm pool, probably accounting for well over half of the overall daily variability in the region (Reid and Gage 1996). We will not be concerned in this study with this higher-frequency local variability, but instead will concentrate on fluctuations which are coherent over large regions and can be tied to planetary-scale dynamics.

The warm bias in NCEP reanalysis tropopause parameters has been documented in detail by Pawson and Fiorino (1998a,b, 1999) and Randel et al. (2000), who found that it was present throughout the Tropics. These authors also found that the bias is not constant over time but increases abruptly in 1979. This change is evident in the reanalysis time series of Fig. 1. The cause of this increase has been shown to be related to the assimilation of satellite data into the reanalysis starting in 1979 (Mo et al. 1995), which results in an erroneous weighting of the temperatures in the lower stratosphere towards those in the (warmer) middle stratosphere. The erroneous temperatures also contribute to a bias in the reanalysis tropopause pressure. For example, the mean tropopause pressure at Koror is 5.3 hPa too high in the reanalysis for the 1966–98 period, although in this case the bias is much smaller during northern summer than in winter. A similar analysis of $T_{\text{trop}}$ and $P_{\text{trop}}$ as well as stratospheric temperature at several other tropical locations reveals similar biases, although the amplitude of the 1979 discontinuity decreases rapidly both above and below the level of the tropopause (Pawson and Fiorino 1998a,b, 1999).

To some extent the 1979 discontinuity can be considered to be a step change in the value of the systematic bias between the reanalysis and radiosonde (and other) data sources. In order to make use of the reanalysis data over an extended period, we have followed a procedure similar to that of Randel et al. (2000), and produced monthly anomaly datasets for the 1957–99 period by using monthly climatological values calculated separately for the 1957–78 and 1979–99 periods. In addition, further data processing through correlation and regression of many years of reanalysis data
ameliorates to some extent the effect of random noise in the data. While these procedures may not completely remove other, less well-documented sources of systematic bias, we will demonstrate later that the reanalysis can still be used to derive valuable quantitative and qualitative information on the large-scale behaviour of the global tropopause and lower stratosphere.

3. INTERANNUAL VARIABILITY

(a) Results from tropical radiosondes

Reid and Gage (1985) and Gage and Reid (1987) documented the interannual behaviour of the tropical tropopause over the Pacific and found a strong dependence on equatorial SST, such that over the west Pacific warm pool the tropopause is higher when SST is anomalously high in the eastern Pacific, as during an El Niño or ‘warm event’. This correlation is highest when SST leads tropopause height by three or four months.

We utilize the ‘Niño 3.4’ SST index to define the state of ENSO over the tropical Pacific. This index is obtained by monthly averaging the SST between 5°N–5°S, 170°W–120°W, and subtracting these values from a 1961–90 climatology (see Trenberth 1997). For reference, the linear dependence of OLR on the SST index is shown in Fig. 2. This map was obtained by regressing OLR against the Niño 3.4 SST anomaly (SSTA), using the period June 1974 to December 1998. The values are plotted for an arbitrary +2.0 degC value of SSTA, which is a typical value of the index during a mature ENSO warm event. The plot shows enhanced convection (negative OLR anomalies) over the western Indian Ocean, and along the equator east of 150°E over the entire tropical Pacific, with maximum amplitude just south of the equator near the date line. A region of suppressed convection is seen over the eastern Indian Ocean and Indonesia, extending into the South Pacific Convergence Zone (SPCZ).

In Fig. 3(a) the monthly mean time series of anomalies in Koror Z_{trop} are plotted along with the Niño 3.4 SSTA. Since the month-to-month variability of the tropopause at Koror shows much less persistence than the SSTA, we have smoothed both series using
a five-month running mean. The positive relationship between these two parameters shown by Gage and Reid (1987) has continued into the most recent decade, although the signal of the stratospheric QBO superimposes additional higher-frequency variability onto the tropopause, as will be discussed in more detail later. Despite this, the correlation between the series in Fig. 3(a) is +0.42, maximized when SST is lagging $T_{\text{trop}}$ by three months, and is significant at the 95% level. Likewise, the correlation between Niño 3.4 SST and Koror $T_{\text{trop}}$ is +0.40 for five-month running means. These relationships are consistent with those obtained by Reid et al. (1989).

A similar plot for $Z_{\text{trop}}$ at Lihue, Hawaii (22.0°N, 159.4°W) is shown in Fig. 3(b). At this subtropical location, where the influence of the QBO is minimal, there is a much higher correlation of +0.66 between Niño 3.4 SST and $Z_{\text{trop}}$, again occurring with SST leading by three months. However, $T_{\text{trop}}$ at Lihue is out of phase with Niño 3.4 SST, with a correlation of −0.51 at zero lag. Thus Koror and Lihue show a concerted response in tropopause height to equatorial Pacific SST variability, but an inverse response in tropopause temperature. In this regard, it is worth noting that there is a negligible OLR signal related to ENSO around Hawaii in Fig. 2, and no substantial local SST or surface temperature signal (Kiladis and Diaz 1989). This is immediately suggestive of a dynamical rather than a local radiative mechanism controlling the Lihue tropopause on these time-scales, the origin of which will be explored further later.

By using the long records of Koror and Lihue radiosonde ascents, the evolution of $T_{\text{trop}}$ and $Z_{\text{trop}}$ at these sites with respect to Niño 3.4 SST can be calculated, and is shown in Figs. 4(a) and (b) (top panels). Also shown are time–height plots of the lagged month–by-month air temperature from 1000 to 10 hPa versus the SST index. These were obtained by regressing the quantities shown versus Niño 3.4 SST and plotting the perturbations associated with a given SST at the various lags. By design, SST peaks at lag zero and is scaled to a +2.0 degC perturbation. As a reference, the average $P_{\text{trop}}$ ($Z_{\text{trop}}$) is 101.0 hPa (16.6 km) at Koror and 117.5 hPa (15.7 km) at Lihue during the sample period.

The in-phase evolution between $Z_{\text{trop}}$ and $T_{\text{trop}}$ at Koror is evident in Fig. 4(a), as opposed to the out-of-phase relationship at Lihue in Fig. 4(b). As discussed previously, $Z_{\text{trop}}$ peaks at both stations three months following the peak SST. At this time, at both stations, most of the troposphere is warm above approximately 800 hPa, with SST leading the temperature at tropopause level by around two months and by a few more months at lower levels. Slight cooling is noted at the lowest levels at Koror, and cooling is seen at both stations in the lower stratosphere, with the largest amplitude at or slightly below 70 hPa. However, this cooling extends well below the level of the tropopause at Lihue, down to around 150 hPa, but is present only in a thin layer down to around 80 hPa at Koror, not quite as low as the tropopause.

Another interesting feature in Fig. 4(a) is the apparent downward phase propagation of the stratospheric temperature perturbations at Koror, where the signals at 70 hPa appear to originate above 10 hPa more than a year earlier. The length of time between the two negative temperature perturbations at 20 hPa in Fig. 4(a) is 27 months, suggesting a relationship with the QBO. Figures 5(a) and (b) show similar plots between Niño 3.4 SST and Singapore (1.4°N, 104.0°E) temperature and zonal wind, extended out to 36 months of lead and lag. Since Singapore is close to the equator, the amplitude of the QBO signal is maximized there, and indeed a little more than two cycles of the QBO are captured by these plots. The downward phase speed and quadrature space and time relationship between the temperature and zonal wind is consistent with that of the QBO (e.g., Andrews et al. 1987). The relationship between Niño 3.4 SST and especially the QBO temperature is statistically significant well beyond two years on either side
Figure 3. (a) Monthly time series of Koror tropopause-height anomalies (heavy line) and Niño 3.4 sea-surface-temperature anomaly (SSTA) (light line) from 1958 to 1998. SSTA are in degC, and tropopause heights are in km multiplied by a factor of 3 for scaling purposes. All values have been smoothed using a five-month running mean, and tropopause-height values have been shifted in time so that they represent the value two months after those of the SSTA for a given month. (b) As (a), except for Lihue, Hawaii.
Figure 4. Vertical cross-section of lagged temperature perturbations at 29 pressure levels (hPa) from radiosonde data at (a) Koror and (b) Lihue, regressed onto Niño 3.4 sea-surface-temperature anomaly (SSTA), using monthly data from 1957 to 1999. Also shown are (top) lagged tropopause height perturbations (km), and (middle) temperature perturbations (degC). Values are plotted for a +2.0 degC perturbation in SSTA. Contour interval is 0.2 degC, with positive (negative) perturbations shown as solid (dashed) contours. Heavy (light) shading denotes statistically significant (at the 95% level) positive (negative) temperature perturbations. Lags are in months from the simultaneous relationship (lag 0).
of an ENSO event, and is large amplitude all the way up to 10 hPa and presumably above that level. Since the QBO is known to affect the tropopause (e.g. Reid and Gage 1985; Randel et al. 2000), Fig. 5 indicates that such links are not entirely independent of ENSO.

Statistical relationships between the QBO and ENSO have been reported in many past studies (e.g. Yasunari 1989; Angell 1992; Collimore et al. 1998), although the precise nature of the physical link remains obscure. Gray et al. (1992) argued that changes in the zonal wind shear near the tropopause due to the QBO might either favour or suppress convective activity near the equator, leading to ENSO fluctuations through the modulation of convection over the west Pacific warm pool. An alternate possibility is that the phase locking of ENSO and the QBO to the annual cycle results in aliasing of the signals of SSTA to the QBO temperature and wind signals. The strong seasonal phase locking of ENSO ensures that the regressed values at lags around zero in Fig. 5 are dominated by the months towards the latter and beginning portion of the year, when Niño 3.4 SSTA tends to have its maximum amplitude. The QBO, however, is much more weakly phase locked to the annual cycle, although the transition from the westerly to the easterly phase at 50 hPa is favoured to occur during northern spring and summer (Dunkerton 1990). This is consistent with the fact that in Fig. 5(b) the 50 hPa easterly to westerly transition occurs seven months after the peak SSTA, which is on average at highest amplitude during December for the period under consideration. Yet another explanation for the SSTA/QBO relationship might involve the forcing of the QBO by vertically propagating waves originating in the troposphere (e.g. Holton and Lindzen 1972; Horinouchi and Yoden 1998). If, for example, vertically propagating Kelvin waves, which are believed to contribute to the momentum flux convergence leading to the transition from QBO easterlies to westerlies, were more favoured during warm events, then these transitions would occur more frequently during those warm episodes and lead to a projection of the QBO phase transition onto the SSTA. In any case, the strong relationship between Pacific SSTA and lower-to-middle stratospheric temperature is an interesting problem that certainly warrants further investigation.

(b) Results from global reanalysis

A more complete picture of the response of the global tropopause to tropical Pacific SST is shown in Fig. 6, which maps the simultaneous regression of reanalysis \( Z_{\text{trop}} \), \( T_{\text{trop}} \) and \( P_{\text{trop}} \) versus Niño 3.4 SSTA for the period from June 1958 to May 1999. As before, the tropopause perturbations are shown for a +2.0 degC Niño 3.4 SSTA. Before discussing these plots in detail, it seems appropriate to justify the methodology used to produce them. In addition to the regression plots shown in Fig. 6, we have also formed seasonal composite maps of global tropopause parameters for the various warm and cold event years identified in Kiladis and Diaz (1989) back to 1958 (not shown). These composites show a remarkable degree of linearity, in that the cold-event patterns are nearly opposite to those of the warm events as far as all of the large-scale features in the plots of Fig. 6 are concerned. Likewise, composites and regressions were also calculated separately for the periods 1957–78 and 1979–99, and very similar patterns were obtained. When the composites and regressions are computed by stratifying according to season, there is little variation in the location of the main features, although their amplitudes are generally higher over the winter hemisphere. There is typically little variation in the signals lagged over time against the SST index, and although there are slightly higher correlations and amplitudes with SST leading by a month or two, this varies considerably from place to place, so here we map only the simultaneous correlations using all of the data, regardless of season. However, it should
Figure 5. As in Fig. 4, except at 16 pressure levels for (a) Singapore temperature (contour interval 0.2 degC) and (b) Singapore zonal wind (contour interval 2 m s$^{-1}$) for the period 1957–90.
Figure 6. Maps of (a) tropopause height, (b) tropopause temperature, and (c) tropopause pressure, regressed onto Niño 3.4 sea-surface-temperature anomalies (SSTA), using monthly data from 1957–99. Values are plotted for a ±2.0 degC perturbation in SSTA. Contour interval is 50 m in (a), 0.2 degC in (b), and 1 hPa in (c). Positive (negative) perturbations are shown as solid (dashed) contours. Heavy (light) shading denotes statistically significant positive (negative) perturbations.
be pointed out that these maps are slightly more representative of the months September to April, when there is a tendency for Niño 3.4 SSTA to have its highest amplitude. Based on these comparisons, the large-scale statistically significant features common to all of these maps are considered to be robust and are discussed later.

Throughout most of the Tropics between about 20°N and 20°S, a higher (lower) than normal tropopause accompanies warm (cold) events in Niño 3.4 SSTA (Fig. 6(a)), confirming the results of Gage and Reid (1986, 1987). This signal reaches its highest positive amplitude in the subtropical Pacific at 20°N and 20°S, and around 140°W longitude, with nearly equal amplitude negative perturbations polewards of these locations at 40°N and 40°S. To the west, the negative \( Z_{\text{trop}} \) regions shift equatorwards to 30°N and 30°S, extending from the Atlantic across the Indian sector to eastern Asia and Australia. Two other extratropical features of note are the high \( Z_{\text{trop}} \) regions over north-western North America and over the Bellingshausen Sea to the west of the Antarctic peninsula.

The \( T_{\text{trop}} \) and \( P_{\text{trop}} \) patterns corresponding to Fig. 6(a) are shown in Figs. 6(b) and (c). While there are strong zonally symmetric tropical \( Z_{\text{trop}} \) and \( P_{\text{trop}} \) signals related to ENSO, \( T_{\text{trop}} \) has a much more zonally varying structure at low latitudes, as was also shown by Randel et al. (2000). However, away from the equator, \( T_{\text{trop}} \) and \( P_{\text{trop}} \) tend to have the same sign, so that when the tropopause is cold its pressure is anomalously low, and vice versa. A comparison of Figs. 6(a), (b) and (c) shows that almost everywhere in the extratropics there is a strong negative correlation between \( Z_{\text{trop}} \) and \( T_{\text{trop}} \), such that when the tropopause is high (low) it is also cold (warm) and its pressure is lower (higher). This is also true throughout the subtropical and equatorial central and eastern Pacific. However, despite the higher than normal tropopause over the western Pacific, \( T_{\text{trop}} \) there is actually anomalously high, as was observed in the radiosonde data of Fig. 4(a) and by Reid et al. (1989). The geographical extent of the region where the tropopause is high and also warm corresponds closely to the region of positive OLR anomaly in Fig. 2 centred on Indonesia. This suggests that the origin of this tropopause warming might be increased upward long-wave radiation from the surface, since this area is less cloudy than normal during ENSO warm phases (Kiladis and Diaz 1989). This mechanism is also consistent with the results of Thuburn and Craig (2000), who predict an in-phase relationship between the height of the tropopause and upward long-wave radiation.

In the previous section it was shown that ENSO influences extend well into the stratosphere at Koror, Libue, and Singapore. Global-scale stratospheric ENSO signals have been demonstrated in many other studies (e.g. van Loon and Labitzke 1987; Reid et al. 1989; Reid 1994; Yulaeva and Wallace 1994). Figure 7(a) shows a map of the temperature perturbation at 70 hPa corresponding to the same Niño 3.4 SSTA as shown in previous figures. Once again, large regions of the globe are covered by statistically significant signals. The pattern correlation between 70 hPa temperature and \( Z_{\text{trop}} \) in Fig. 6(a) is negative and strong, with a cold stratosphere where \( Z_{\text{trop}} \) is high, and vice versa. Likewise, the correlation is strongly positive with \( T_{\text{trop}} \) in Fig. 6(b), so that at this level the temperature perturbation has the same sign as the \( T_{\text{trop}} \) anomaly in most regions. Significant global-scale temperature signals continue to extend upwards into the middle stratosphere (not shown).

The behaviour of some of the gross features of the tropopause signals in Fig. 6 can be understood as a dynamical response to the displacement of convection during ENSO episodes, which has been studied extensively (e.g. Trenberth et al. 1998). Figure 7(b) shows the 100 hPa geopotential height and wind corresponding to the tropopause perturbation fields of Fig. 6. This pressure surface lies on average close to the tropopause throughout most of the Tropics (Highwood and Hoskins 1998; Seidel
Figure 7. As in Fig. 6 except for (a) temperature at 70 hPa, and (b) geopotential height and statistically significant wind at 100 hPa. Contour intervals are 0.2 degC in (a) and 5 m in (b). The largest wind vector in (b) represents about a 5 m s\(^{-1}\) perturbation.

\textit{et al.} 2001), while in the extratropics this level is well into the lower stratosphere. The extratropical geopotential-height pattern in Fig. 7(b) is remarkably equivalent barotropic, with virtually all of the main features present from 1000 hPa all the way up to 10 hPa, allowing for a small amount of westward tilt with height (not shown). Thus the extratropical circulation pattern captured by Fig. 7(b) is also representative of the tropopause level at those latitudes.

The zonally symmetric signal of positive geopotential-height perturbations in the Tropics is a reflection of the anomalously warmer (cooler) than normal tropical tropospheric temperature during warm (cold) events, present at virtually all longitudes from the surface (e.g. Kiladis and Diaz 1989) throughout the depth of the troposphere (Newell and Weare 1976; Reid \textit{et al.} 1989; Yulaeva and Wallace 1994; Kiladis and Mo 1998). In models of the tropopause based on radiative–convective constraints (Held 1982; Thuburn and Craig 1997, 2000), \(Z_{\text{trop}}\) increases with both surface temperature and with a decrease in the tropospheric lapse rate. The elevated \(Z_{\text{trop}}\) throughout the
Tropics during a warm event might then be considered to be a consequence of this combination of factors. For example, in Fig. 4(a) there is little change in surface temperature at Koror, or even slight cooling, but the large increase in temperature close to 100 hPa would result in a net decrease in the tropospheric lapse rate. For the Tropics as a whole, the zonal mean vertical structure of temperature (not shown) shows that there is little overall change in $T_{\text{trop}}$, as might be inferred from Fig. 6(b), but that there is a substantial increase in surface temperature. These arguments, however, appear to break down at Lihue (Fig. 4(b)), where a large increase in $Z_{\text{trop}}$ is accompanied by a net increase in tropospheric lapse rate resulting from the strong cooling at tropopause levels and little change in surface temperature. This suggests that simple radiative–convective models assuming a constant tropospheric lapse rate may have to be modified to account for the vertical structure of temperature as seen in Fig. 4.

In Fig. 7(b), local maxima in geopotential-height anomalies occur in the subtropical Pacific of both hemispheres at 140°W, the same locations as the maxima in both $Z_{\text{trop}}$ and $T_{\text{trop}}$, and minima in $P_{\text{trop}}$, in Fig. 6. This is the sector where the anomalous convection over the Pacific sector (Fig. 2) is associated with the most pronounced enhancement of the local Hadley circulation, leading to a spin-up of the Pacific subtropical anticyclones. This signal also appears as anticyclonic potential vorticity (PV) anomalies maximized close to tropopause level (not shown), which are a response to enhanced upper-tropospheric mass outflow from the anomalous equatorial convection (Sardeshmukh and Hoskins 1985; Mo and Rasmussen 1993).

An increase in the subtropical geopotential-height gradient occurs at all longitudes, tracing the mean position of the climatological westerly maxima of both hemispheres. The increase in the geostrophic wind over the central and eastern Pacific reflects an extension of the Asian and Australian subtropical jet systems, along with an equatorward shift in the jet axes over the Indian sector. Midlatitude cyclonic anomalies near the Aleutians and over the South Pacific, and anticyclones over central Canada and the Bellinghausen Sea, are consistent with Rossby responses to the subtropical Pacific Rossby-wave sources (Trenberth et al. 1998; Kiladis and Mo 1998). There is a high overall pattern correlation between the geopotential-height features of Fig. 7(b) and the tropopause fields of Fig. 6, although there are many interesting exceptions in certain regions.

The vertical structure of the subtropical tropopause and temperature fields is quite consistent with the upper-tropospheric anticyclonic PV perturbations generated by the anomalous convection. As discussed by Hoskins et al. (1985), anticyclonic PV perturbations in the upper troposphere lead to an upward displacement in the tropopause, and require negative temperature anomalies above and positive temperature anomalies below them to maintain hydrostatic and geostrophic balance. This is just the structure seen, for instance, in the Lihue vertical temperature field of Fig. 4(b), where cross-sections of PV perturbations calculated from reanalysis data show maximum amplitude at the level of the zero line in the temperature-anomaly field at around 150 hPa (not shown). An alternative but entirely consistent view is that the subtropical jet exit region, the axis of which would lie at around 150 hPa just to the north of Lihue, would have anomalous subsidence below and rising motion above it along its equatorward flank (e.g. Murray and Daniels 1953).

In the extratropics, a comparison of Fig. 7(b) with the fields in Fig. 6 shows that, overall, cyclonic upper-level perturbations are associated with a low $Z_{\text{trop}}$ and high $T_{\text{trop}}$ and $P_{\text{trop}}$ (and vice versa). As has been known for a long time, similar tropopause behaviour is seen on the synoptic scale, for example, accompanying the passage of an upper-level cut off cyclone (e.g. Palmén 1951), and is in accord with PV considerations.
(see Hoskins et al. (1985) for examples). Under the assumption of quasi-geostrophy, Juckes (1994) showed that the tropopause potential temperature should be very nearly proportional to its geopotential height, with a constant of proportionality dependent upon the stratospheric and tropospheric lapse rates. A calculation of the tropopause potential temperature analogous to the fields shown in Fig. 6 (not shown) reveals a high pattern correlation with \( Z_{\text{trop}} \) in Fig. 6(a) in the extratropics, in line with the existence of this local relationship.

Hydrostatic and PV arguments also provide a convenient framework for considering the tropopause perturbations closer to the equator. Highwood and Hoskins (1998) studied the large-scale response of the tropical tropopause to convective heating through the use of a dry primitive-equation model, in which a diabatic heating was imposed along the equator. This produced a Gill-type response, with subtropical anticyclonic gyres to the west and a Kelvin response with equatorial westerlies to the east of the heat source at upper levels, much like that seen in Fig. 7(b). The vertical structure featured increased (decreased) pressure in the upper (lower) troposphere, with subsidence and warming of the middle and lower troposphere and upward motion and cooling extending through the upper troposphere, tropopause, and into the lower stratosphere.

It seems evident that, on interannual time-scales at least, the lower stratospheric temperature and tropopause perturbations at most locations can be explained by purely dynamical arguments. However, we can not discount the possibility that diabatic or 'downward control' principles may be of some importance to the details of the distribution, especially in the Tropics. Further study of the fine structure of these perturbations may help in quantifying the radiative and other dynamical effects on the tropopause.

4. AN EXAMPLE OF INTRASEASONAL TROPOPAUSE VARIABILITY

In this section we briefly examine some aspects of intraseasonal fluctuations in the tropical tropopause, through an examination of the convective and dynamical signals associated with \( T_{\text{trop}} \) fluctuations over the west Pacific warm pool. In this case, daily \( T_{\text{trop}} \) at Koror was high-pass filtered with a 120-day cut-off to remove the seasonal cycle and interannual variability. This time series was then correlated and regressed against various dynamical fields of the NCEP reanalysis dataset, similar to the technique using Niño 3.4 SSTA in the previous section. In this approach no a priori assumptions about the preferred time-scale of the fluctuations were made, except that they are subseasonal with periods of less than 120 days.

Figure 8 shows the simultaneous regressed values of global \( T_{\text{trop}} \) associated with a \(-1\) standard deviation of Koror \( T_{\text{trop}} \) (-2.2 degC) from radiosondes for December-February (DJF) 1979/80 to 1997/98. This season was chosen because it is the time of maximum amplitude in the MJO (Madden and Julian 1994). Fluctuations in global \( T_{\text{trop}} \) associated with Koror tropopause variability are coherent and statistically significant over large regions within 45° of latitude of the equator. Over the base region of the western Pacific, anomalously cold tropopause values extend westwards in a horseshoe-shaped pattern into the subtropics of the Indian sector. This connects to a zonally oriented low \( T_{\text{trop}} \) along the poleward side of the jet in the North Pacific. A warm tropopause signal extends along the equator to the west of Indonesia and in the subtropics of the Pacific sector. The overall pattern is reminiscent of that in Fig. 6(b), except with opposite sign, even over extratropical regions such as eastern Europe and North America. As was the case for the interannual fluctuations, global \( T_{\text{trop}} \) appears to be controlled by large-scale dynamics and not local forcing, at least on the time-scale represented in Fig. 8.
The tropical convective and planetary-scale circulation and fields associated with the dominant mode of DJF intraseasonal tropopause variability over the warm pool were isolated by regressing the OLR and 100 hPa wind against Koror $T_{trop}$ in Fig. 9. Recall that, as in Fig. 7(b), these signals lie at a level close to the tropopause in the Tropics, but are well into the stratosphere beyond the latitudes of the subtropical jets. By design, the Koror $T_{trop}$ is coldest at zero lag and is associated with the convective and flow-perturbation pattern of Fig. 9(b). Notably, the deepest convection at this time (the heavy-shaded negative OLR anomalies in Fig. 9(b)) is not centred over Koror, but is located to the west over Indonesia. At the same time, convection is suppressed just south of the equator near the date line. The lag relationships in Figs. 9(a) and (c) show that the negative OLR signal appears to move over Indonesia from the Indian Ocean, and eventually propagates into the SPCZ at 10°S.

A Hovmuller diagram of the near-equatorial OLR signal (not shown) confirms that the convective portion of the disturbance propagates eastwards at approximately 5–10 m s$^{-1}$, moving most slowly over the warm pool, consistent with signal of the MJO (Madden and Julian 1972, 1994). In the original studies of Madden and Julian, tropopause variability associated with the oscillation was inferred using 100 hPa temperature data from radiosondes. They also found that the tropopause appeared to be higher and colder within and especially to the east of MJO convection.

The circulations of Fig. 9 are very nearly equivalent barotropic through a large range of levels in the upper troposphere (not shown), in agreement with many studies of large-scale dynamical fields of the MJO (e.g. Knutson and Weickmann 1987; Rui and Wang 1990; Salby and Hendon 1994; Matthews and Kiladis 1999). The convective heating field of the MJO is associated with subtropical anticyclonic upper-tropospheric perturbations at the longitude of the convection and to its west, most pronounced in the northern (winter) hemisphere, as seen by the anticyclonic stream-function perturbations propagating eastwards from the Indian sector to the western Pacific in Fig. 9. To the east of the convection, twin cyclones also move across the Pacific in this sequence. These signals are broadly consistent with the generation of anticyclonic Rossby-wave sources in the subtropics forced by the divergent circulation associated with the convective
diabatic heating, and a cyclonic Rossby response to the east (Sardeshmukh and Hoskins 1988). In reality, the picture is more complicated than the barotropic dynamics would indicate owing to the fact that there is upward propagation of wave energy associated with the heating as well.

Some insight into the vertical structure associated with the propagating MJO heating field in Fig. 9 can be obtained by calculating the vertical temperature perturbation accompanying Koror tropopause perturbations, as shown in Fig. 10. Here the radiosonde temperatures at all levels are regressed against 120-day high-pass filtered $T_{\text{ trop}}$ from Day $-12$ to Day $+12$. The evolution of the OLR perturbation at the grid point nearest
Figure 10. Vertical cross-section of lagged regressed values of the temperature perturbation at 29 pressure levels (hPa) from Koror radiosonde data associated with a $-1$ standard deviation anomaly in less than 120-day filtered $T_{trop} (-2.2 \text{ degC})$. Contour interval is 0.2 degC (negative contours dashed). Data used in the calculation are from December–February from 1979/80 to 1997/98. Lags are in days from the simultaneous relationship (lag 0). Also shown (top) are lagged outgoing long-wave radiation values at the nearest grid point to Koror (W m$^{-2}$).

to Koror is shown by the time series on top. Cold Koror $T_{trop}$ appears as a negative temperature perturbation centred at the 100 hPa level, with an anomalously warm signal throughout the entire troposphere below. Just above the cold tropopause signature, a vertically thin warm signal is seen, with a negative perturbation above that extending all the way up to 10 hPa. These upper features show a downward phase propagation over time. As expected from Fig. 9, the OLR is at a minimum value from two to four days following the minimum in $T_{trop}$.

To understand the thermal response of the atmosphere to convective heating on these time-scales, it is useful to consider the theoretical signal of a vertically propagating gravity wave generated by a moving diabatic heat source, as might be associated with MJO convection. The phase lines of such an eastward propagating wave should tilt upwards and eastwards with height, which is also the direction of the group velocity (e.g. Andrews et al. 1987, Fig. 4.19). Thus above the level of the energy source for such waves (the convective heating), the maximum wave amplitude should be to the east, and downward phase propagation should be observed locally. This is as observed in Fig. 10, with the lowest $T_{trop}$ occurring before the OLR minimum passes by the station. The troposphere is dominated by positive temperature perturbations, also peaking a day or so prior to the passage of the convective maximum. The greater persistence of the stratospheric signals compared to those in the troposphere implies a larger spatial extent of the former features, which is confirmed by maps of the MJO temperature perturbations produced from reanalyses (not shown).

Similar vertical structure is observed for a variety of higher-frequency convectively coupled waves (Wheeler et al. 2000). In general, these waves all have a positive temperature maximum at about the level of maximum latent heating (250 hPa or
so, as in Fig. 10), low temperatures at the tropopause, and large-amplitude, wave-like temperature perturbations propagating vertically into the stratosphere. Thus the horizontal and vertical tropospheric and stratospheric structure of these waves, and consequently the behaviour of the tropopause, agree quite well with linear gravity-wave theory.

5. DISCUSSION

We have examined a portion of the variability in the global tropopause and lower stratosphere that can be directly tied to ENSO and the MJO. On the largest spatial scales, these signals can be explained as a dynamical response to the shifting of tropical convection. Previous work has also related interannual variability in the tropopause and stratosphere to the QBO (e.g. Angell and Korshover 1964; Reid and Gage 1985; van Loon and Labitzke 1987; Randel et al. 2000), solar fluctuations (e.g. Stranz 1959; Cole 1975; Gage and Reid 1981; van Loon and Labitzke 1987, 1998) and volcanic eruptions (e.g. Labitzke and van Loon 1989; Angell 1993; Reid 1994). Taken together, these forcings can explain a large portion of the interannual variability of the tropopause on planetary scales (Reid and Gage 1981, 1993; Reid 1994; Randel et al. 2000).

It has been known for a long time that the extratropical tropopause fluctuates significantly with the passage of synoptic-scale disturbances (e.g. Palenm 1951). This relationship is such that an upper-level cyclonic perturbation is associated with a lower and warmer tropopause at higher pressure, with anomalously high stratospheric temperatures above and low tropospheric temperatures below (and vice versa for anticyclones). Accompanying perturbations in stability and vertical motion follow directly from this arrangement based on PV considerations (Hoskins et al. 1985). The large extratropical tropopause signals related to ENSO can be understood to be a result of the propagation of quasi-stationary Rossby waves (e.g. Trenberth et al. 1998), and the accompanying shift in the storm tracks (Shapiro et al. 2000), which modulate synoptic-scale activity and result in the time-mean tropopause responses seen in Fig. 6. Similar effects appear to operate on the MJO time-scale, and signals there appear to be related to the horizontal and vertical propagation of low-frequency Rossby and gravity waves forced by the moving heat source related to the disturbance.

These results contrast with those invoked to explain the climatological distribution and seasonal cycle of the tropopause and lower stratosphere, where the effects of radiation, convective and baroclinic adjustment, and wave driving within the middle atmosphere appear to be of more than secondary importance (e.g. Reid and Gage 1981, 1996; Held 1982; Thuburn and Craig 1997, 2000; Highwood and Hoskins 1998; Simmons et al. 1999).

Much of the intraseasonal variability at the tropical tropopause and in the lower stratosphere at higher frequencies than the MJO (i.e. with less than 30-day periods) is related to the passage of equatorially trapped disturbances such as Kelvin waves (Wallace and Kousky 1968; Tsuda et al. 1994; Shimizu and Tsuda 1997), some of which are coupled to deep tropical convection (Wheeler and Kiladis 1999; Wheeler et al. 2000). These disturbances have been shown to possess high-amplitude temperature and wind perturbations at the levels of the tropopause and lower stratosphere. We plan further work emphasizing the role of these higher-frequency disturbances and local mesoscale convection in global and local tropopause behaviour.

It is well known that substantial interannual signals in ozone and water- vapour distribution in the upper troposphere and lower stratosphere are related to ENSO (e.g. Randel and Cobb 1994; Chandra et al. 1998; Langford et al. 1998). Intraseasonal
fluctuations in upper-tropospheric ozone and water vapour have also been tied to the passage of Kelvin waves and the MJO (Fujiwara et al. 1998; Mote et al. 2000). Understanding the dynamics of these signals will be crucial to the development of better conceptual models of stratosphere–troposphere exchange processes.

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