The surface of maximum wind as an alternative to the isobaric surface for wind climatology

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[1] The surface of maximum wind (SMW) is introduced as an analysis frame for climate studies of fast upper tropospheric winds. As defined here, the vertical domain for the SMW includes the troposphere above 500 hPa and upper tropospheric jet streams that may protrude into the lower stratosphere. We use NCEP-NCAR Reanalysis data for summers 1958–2004 in the vicinity of the tropical easterly jet (TEJ) to show how the spatial and temporal variability of the SMW relate to jet stream cores and the tropopause. We then compare the SMW climatology of the TEJ to an isobaric climatology of the TEJ, demonstrating that the SMW climatology reveals descent and slowing of the TEJ over the period of record, whereas the isobaric climatology provides only an overestimate of the TEJ slowing trend. Citation: Strong, C., and R. E. Davis (2005), The surface of maximum wind as an alternative to the isobaric surface for wind climatology, Geophys. Res. Lett., 32, L04813, doi:10.1029/2004GL022039.

1. Introduction

[2] Upper tropospheric circulation has frequently been studied as an indicator of climate change and variability [e.g., Angell, 1998; Frauenfeld and Davis, 2003; Tanaka et al., 2004]. In cases where upper tropospheric wind speed is the variable of interest, isobaric surfaces are convenient because upper air data are commonly archived at constant pressures (e.g. 300 hPa), and because the horizontal momentum equation is simplified in isobaric coordinates by the elimination of density. However, some inaccuracy may result when data obtained from a single isobaric surface are used to evaluate climate trends in jet streams or fast upper tropospheric winds in general. At any one time, no isobaric surface can capture a jet stream whose altitude varies spatially in the hemisphere, and, at any one location, no isobaric surface can capture a jet stream whose vertical position varies temporally. The temporal variability of the tropopause jets motivates consideration of an analysis frame that moves vertically over time at each node to follow the position of the fastest winds. The U.S. Navy began operationally mapping the altitude and vertically averaged speed of the three-dimensional layer of maximum wind (LMW) beginning at the end of 1958 [Reiter, 1961] for aviation purposes, and researchers have considered properties of fast upper tropospheric flow in coordinate systems horizontally aligned with jet stream axes in both analytical [Cressman, 1950] and theoretical [Houghton, 1965] applications. The NCEP-NCAR Reanalysis [Kalnay et al., 1996] includes a maximum wind level pressure that tracks wind speed maxima in the upper troposphere and lower stratosphere.

[3] Here, we define an upper tropospheric surface of maximum wind (SMW) that excludes stratospheric wind maxima not associated with tropospheric jet streams, and we present the SMW as an alternative to isobaric surfaces for climate analysis of fast upper tropospheric winds. The SMW is objectively located by algorithm (Section 2.2) and enables study of simultaneous changes in the speed and altitude of jet streams. The SMW is vertically Lagrangian (changes its pressure temporally to coincide with the fastest observed wind speeds), thus avoiding the temporal speed signal on isobaric surfaces that is generated by vertical motion of jet cores. As an example demonstrating how the spatial and temporal variability of the SMW relate to jet streams and the tropopause, we use Reanalysis data for summers 1958–2004 in the vicinity of the tropical easterly jet (TEJ). We then compare the SMW climatology of the TEJ to an isobaric climatology of the TEJ.

2. Data and Methods

2.1. Data

[4] We use wind speed on seven isobaric surfaces from 500 to 100 hPa resolved every six hours on a 2.5° grid from the NCEP-NCAR Reanalysis, and Reanalysis tropopause pressure data at the same temporal and horizontal spatial resolution. Artificial trends or discontinuities may exist in the Reanalysis because of variations in the observing system including changes in the density of rawinsonde data and the use of satellite data after 1978 [Kistler et al., 2001]. Concerning the present analysis of the TEJ, Rao et al. [2004] report favorable comparison of 100 hPa rawinsonde and Reanalysis data over India near the TEJ for 1964–1998. More generally, Kanamitsu et al. [1997] found that satellite data impacted the Northern Hemisphere tropopause Reanalysis less than it impacted data sparse areas such as the stratosphere and eastern oceanic areas of the Southern Hemisphere.

2.2. The Surface of Maximum Wind

[5] The domain for the SMW includes the troposphere above 500 hPa and upper tropospheric jet streams that may protrude into the lower stratosphere. We use Reanalysis tropopause pressure and wind speed data from 00 UTC on 1 January 2004 along 140°E (Figure 1) to illustrate the algorithm for locating the SMW. The tick marks on the axes of Figure 1 show the latitude and pressure resolution of the Reanalysis data, and the SMW is defined by selecting the fastest wind speed from the column of candidates above each 2.5° node (abscissa tick, Figure 1). Each wind speed datum from 500 to 100 hPa is a SMW candidate unless the
3. Results

3.1. SMW Variability

[8] Figure 2 shows the 1958–2004 central tendency and variability of summer (June–August) $P$ in a cross section that includes the TEJ (near 10°N) and the subtropical jet (STJ, near 40°N). The SMW is close to or above the tropopause in the vicinity of the TEJ and STJ wind maxima at 15 and 40°N, but is up to 100 hPa below the tropopause in the intervening latitudes. In polar latitudes, the SMW resides below the tropopause while approximating its slope. The vertical variability of $P$ along the meridian in Figure 2 indicates potential inaccuracies associated with using isobaric surfaces to study fast upper tropospheric flow over large latitudinal domains. Even over a restricted latitudinal domain, the use of an isobaric surface for the study of fast upper tropospheric flow is problematic because of the temporal variability of $P$ about its local mean. The range between the 10th and 90th percentiles of seasonal mean $P$ at 10°N, for example, is larger than 30 hPa. Figure 3 illustrates the vertical motion of the TEJ core underlying the tropical $P$ variability in three dimensions. The mean position of the TEJ jet core initially clustered about 110 hPa and 12°N, but underwent descent and equatorward drift, clustering about 135 hPa and 7°N toward the end of the record.

3.2. SMW Versus Isobaric Climatology

[9] We directly compare the SMW climatology of the TEJ to an isobaric TEJ climatology, selecting the 100 hPa isobaric surface commonly used in TEJ climate studies [e.g., Rao et al., 2004]. The SMW and 100 hPa variables the seasonal mean sample size is 368 at each node. Thus, for summer (June–August) 2004, the 95% confidence intervals on the estimate of the seasonal mean are typically $P \pm 7.2$ hPa and rarely exceed $P \pm 19.0$ hPa. Although vertically coarse relative to rawinsonde data, the Reanalysis data provide seasonal mean variations in SMW pressure and speed that correlate logically with the seasonal mean speed on nearby isobaric surfaces (Section 3.2).

Figure 2. For 1958–2004 data along 72.5°E, mean isolochs in m/s (gray lines), the tropopause (bold solid line) with 10th to 90th percentile of seasonal means (gray region) and the surface of maximum wind (SMW, bold dashed line) with 10th to 90th percentiles of seasonal (June–August) means (dashed lines).
are averages from 5–15°N, 60–90°E (the bold-edged box at the base of Figure 3) which we will call the “TEJ zone,” noting that the mean position of the TEJ core is captured by this zone for each of the summers 1958–2004. The $P$ time series in Figure 4a reflects the general descent of the TEJ away from 100 hPa ($P$ increase) in the TEJ zone, including a relatively abrupt $P$ increase between 1977 and 1983 (bounded by dashed vertical lines on the figure). The abrupt change is coincident with the 1979 introduction of satellite data into the Reanalysis, and an iterative multiple change point test \[\text{Lanzante}, 1996\] does flag 1978–1979 as a change point significant at $\alpha = 0.05$. However, the change is also coincident with the mid-to-late 1970’s climate shift in numerous atmospheric and ecological data bases \[\text{Trenberth and Hurrell}, 1994\] and, as mentioned in Section 2.1, \text{Rao et al.} [2004] report favorable comparison of 100 hPa rawinsonde and Reanalysis data over India near the TEJ for 1964–1998. TEJ zone $V$ slowed by approximately 6 m/s over the record (Figure 4b). The mean 100 hPa wind speed in the TEJ zone also slowed ($\bar{V}_{100}$, Figure 4b), but always underrepresented the speed of the TEJ and the TEJ’s vertical motion relative to 100 hPa. The 12 m/s $\bar{V}_{100}$ slowing between 1977 and 1983, for example, reflects the combined effect of a 6 m/s wind speed decrease on the SMW ($\bar{V}$, Figure 4b) and a 15 hPa descent of the SMW away from 100 hPa ($\bar{P}$, Figure 4a).

The notion that isobaric wind speed variability is influenced by the speed and proximity of jet cores can be statistically explored by using speed on the SMW ($\bar{V}$) and the proximity of the SMW to 100 hPa ($|\bar{P} - 100|$) as candidate predictors of $\bar{V}_{100}$. The F ratio is higher for the SMW proximity regression ($F = 177.3$), so SMW proximity is introduced first, accounting for more than 90% of the $\bar{V}_{100}$ variability (Figure 5a). Regressing $\bar{V}_{100}$ against SMW proximity leaves the residuals $\bar{V}_{100}(|\bar{P} - 100|)$ (denoting the speed at 100 hPa with the effect of SMW proximity removed). Changes in SMW speed then accounted for 29% of the $\bar{V}_{100}(|\bar{P} - 100|)$ residual variability (Figure 5b).

Jet stream speed and jet stream vertical proximity may each exert a statistically discernible influence on the wind speed signal obtained from an isobaric surface, underscoring the challenges associated with the interpretation of such signals and their inadequacy for the proper representation of jet stream speeds. The SMW signal, although still confounded by some jet core vertical motion due to the relatively coarse vertical resolution (50–100 hPa between data levels), more accurately reflects the speed trends in the TEJ and allows simultaneous conclusions about trends in TEJ altitude and TEJ speed.

4. Summary and Discussion

The TEJ was used to demonstrate that the wind speed signal on an isobaric surface is influenced simultaneously by variability in the strength and proximity of nearby jet streams. In the TEJ example, the 25 hPa descent

![Figure 3](image-url) (TEJ) core for each summer 1958–2004 (large circles with grayscale shading indicate the year associated with each circle). The small circles show the projection of the large circles onto the “edges” of the figure. A map of India is shown and the figure is oriented so that the cross section in Figure 2 meridionally bisects this volume with north to the right in both figures. The bold black box over southern India outlines the area that captures all of the TEJ core positions between 1958–2004 (the “TEJ zone”).

![Figure 4](image-url) (a) The summer (June–August) mean pressure of the surface of maximum wind (SMW) within the tropical easterly jet zone (TEJ zone: 5–15°N, 60–90°E) as defined in Figure 3. Gray shading shows 95% confidence limits. (b) The summer mean SMW speed in the TEJ zone ($\bar{V}$, line with circles), and the mean speed at 100 hPa within the TEJ zone (bold gray line). The dashed vertical lines bound six years during which the SMW underwent a 15 hPa descent.
and 6 m/s slowing of the TEJ between 1958 and 2004 combined to produce a 12 m/s slowing signal at 100 hPa. Additional examples can be envisioned in which jet core vertical motion combines with a jet core speed acceleration or deceleration to produce an isobaric wind speed signal that overestimates or underestimates the actual core speed trend. The potential for difficulties in interpreting isobaric results motivates an analysis frame that is vertically Lagrangian with respect to jet stream altitude, and we recommend the repeatable and objectively defined SMW described here. In addition to producing more representative time series of tropopause jet stream speeds for climate studies of wind, the SMW also enables climatologists to study jet stream altitude variability.

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References

Figure 5. For summer (June-August) data from within the tropical easterly jet zone (TEJ zone, 5–15°N, 60–90°E): (a) least squares linear regression of the seasonal mean wind speed at 100 hPa ($V_{100}$) versus the vertical proximity of the surface of maximum wind (SMW) to 100 hPa ($|P - 100|$). (b) the residuals of (a) regressed against SMW wind speed ($V_{SMW}$). The regression statistics are the probability of a Type I error ($p$), the squared Pearson correlation coefficient ($r^2$), the effective degrees of freedom ($df$), and the mean squared error (MSE).

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