TEMPERATURE-RELATED TRENDS IN THE VERTICAL POSITION OF THE SUMMER UPPER TROPOSPHERIC SURFACE OF MAXIMUM WIND OVER THE NORTHERN HEMISPHERE

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ABSTRACT

The surface of maximum wind (SMW) is used to examine spatial and temporal variability in the vertical position of jet streams and fast upper tropospheric wind maxima over the Northern Hemisphere (NH) in the NCEP/NCAR Reanalysis for summers 1958–2004. At a given observation time in a gridded data set, the SMW is defined as the surface passing through the fastest analyzed wind above each grid node, with a vertical search domain restricted to the upper troposphere and any tropospheric jet streams extending into the lower stratosphere. The 47-year mean summer SMW generally resides below the tropopause, undulates in the tropics, and descends poleward in middle and high latitudes. Trends in the pressure of the summer SMW are primarily positive, exceed 30 hPa/decade at some locations, and are found over most of the tropics and subtropics for the period 1958–2004. The thermal wind relationship is used to establish links between the SMW pressure trends and temperature gradient changes in the upper troposphere. The changing temperature gradients are consistent with nonuniform tropospheric warming and are correlated with sea surface temperature (SST) variability related to El Niño and the Pacific Decadal Oscillation. Copyright © 2006 Royal Meteorological Society.

KEY WORDS: jet stream; circulation; climatology; climate change

1. INTRODUCTION

1.1. Perspective

Following the discovery of concentrated bands of fast upper tropospheric winds (Ooishi, 1926; Lewis, 2003 and references therein), research was undertaken to document and understand their position in the atmosphere. The location and strength of upper tropospheric jet streams were theorized and confirmed to be linked to the position and strength of underlying frontal boundaries (Staff Members, Department of Meteorology, University of Chicago, 1947; Palmen, 1948; Palmen and Newton, 1948). On the seasonal timescale, the winter intensification and equatorward expansion of the polar front jet (PFJ) in conjunction with the intensification and equatorward expansion of the polar high (Namias and Clapp, 1949) demonstrates the linkage between the thermal structure of the atmosphere and fast upper tropospheric wind features.

On the interannual timescale, the position and strength of fast upper tropospheric winds vary with the Arctic Oscillation (AO) – the meridional seesaw of atmospheric mass between the polar region and midlatitudes (Lorenz, 1951; Kutzbach, 1970) apparent in the leading principal component of wintertime sea-level pressure (Thompson and Wallace, 1998). The positive phase of the AO is associated with slower (faster) zonal wind
speeds near 30°N (60°N) from the surface into the stratosphere for data zonally averaged over the Northern Hemisphere (NH) (Thompson and Wallace, 2000). The phase of the winter AO also influences atmospheric circulation and the distribution of air temperature during the subsequent summer (Ogi et al., 2003a, 2004a), particularly during maxima in the 11-year solar cycle (Ogi et al., 2003b). The lagged correlation stems in part from the persistence of winter AO-related anomalies of sea ice cover (Rigor et al., 2002), snow cover, and sea surface temperatures (SSTs) in circumpolar regions (Ogi et al., 2003a). The AO is also discernible in summertime geopotential height fields as an annular structure with smaller meridional extent than its winter analog (Ogi et al., 2004b).

Interannual variability in the horizontal position of fast upper tropospheric winds can also be traced to SST variability. The Pacific jet stream is intensified and displaced equatorward during the warm phase of the El Niño/Southern Oscillation (ENSO), accompanied by changes in wind speed (Arkin, 1982; Rasmusson and Mo, 1993; Yang et al., 2002), the circumpolar vortex (Angell and Korshover, 1985; Frauenfeld and Davis, 2000; Angell, 2001), middle and upper tropospheric geopotential height topography (Horel and Wallace, 1981; Hastenrath, 2003), and storm tracks (Straus and Shukla, 1997; Chen and Van den Dool, 1999). The jet stream and general tropospheric circulation can also be influenced by SST variability with periods longer than the ENSO cycle. Multidecadal oscillations have been identified in Atlantic SST fields north of the equator in records dating to 1850 (Schlesinger and Ramankutty, 1994) and in climate reconstructions dating to 1650 (Delworth and Mann, 2000). The ability of this ‘Atlantic Multidecadal Oscillation’ (AMO) (Kerr, 2000) to account for tropospheric geopotential height variability extends beyond the Atlantic (Mestas-Nuñez and Enfield, 1999; Delworth and Mann, 2000; Enfield et al., 2001). Sutton and Hodson (2005) recently documented the AMO’s influence on summer sea-level pressure, precipitation, and surface air temperature over North America and Europe.

Changing tropospheric carbon dioxide and stratospheric ozone concentrations (Tett et al., 1996) may potentially influence the strength and position of upper tropospheric jet streams. Tropospheric warming and lower stratospheric cooling are robustly observed (Angell, 1991; Parker et al., 1997; Hansen et al., 1997; Santer et al., 1999; Gaffen et al., 2000; Ramaswamy et al., 2001) and consistent with theory and modeling work concerning changing concentrations of tropospheric carbon dioxide and stratospheric ozone (Schlesinger and Mitchell, 1987; Tett et al., 1996; Ramaswamy et al., 2001). The significance, magnitude, and sign of upper tropospheric and lower stratospheric temperature trends vary in the literature (e.g. Ramaswamy et al., 2001) in part because of variability in spatial coverage among data sets, regional temperature variations, analysis methods, and equipment changes (Gaffen, 1994; Parker et al., 1997; Basist and Chelliah, 1997; Stendel et al., 1998; Hurrell and Trenberth, 1997, 1998; Santer et al., 1999; Chase et al., 2000).

Jet stream-level temperature fields frequently exhibit temporal variability associated with the 11-year solar cycle (e.g. Labitzke and van Loon, 1988; van Loon and Shea, 1999, 2000) and volcanic emissions (e.g. Robock and Mao, 1992; Kelly et al., 1996; Kirchner et al., 1999). Some air temperature records near the jet stream level contain steplike changes (Pawson et al., 1998; Gaffen et al., 2000) temporally coincident with the mid-to-late 1970s climate shift documented in atmospheric fields (Nitta and Yamada, 1989; and Trenberth, 1990; Graham, 1994; Miller et al., 1994) and biological data (Ebbesmeyer et al., 1991; Francis and Hare, 1994; McGowan et al., 1998; Hare and Mantua, 2000). There is evidence that, rather than being unique, the mid-to-late 1970s climate shift was one of a sequence of shifts associated with at least two interdecadal climate oscillations: one with 50–70 year periodicity and the other with 15–25 year periodicity (Minobe, 1999, 2000). The interdecadal variability is often referred to as the Pacific Decadal Oscillation (PDO) (Mantua et al., 1997; Zhang et al., 1997), and its underlying physical mechanism is a focus of ongoing research (Mantua and Hare, 2002; Seager et al., 2004). The presence of a distinctly decadal signal in the climate of the Pacific Ocean was recently established by examining the coupled behavior of SSTs and NH atmospheric circulation (Frauenfeld et al., 2005).

The vertical positions of upper tropospheric wind maxima also relate to the thermal structure of the atmosphere as indicated by the thermal wind equation (Section 1.2). Jet stream vertical position variability, although less commonly studied than speed or horizontal position variability, has received some research attention. Mintz (1954) produced cross sections of zonally averaged summer and winter wind speed data from all latitudes, revealing the approximate vertical position of persistent jet cores. A distribution of pressure at
the maximum wind was developed for 261 soundings in jet stream regions over the United States, revealing a peaked distribution centered on 215 hPa with a 225 hPa range (Endlich et al., 1955). 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 for aviation purposes (Reiter, 1958, 1961). The LMW was found to protrude into the lower stratosphere on the poleward flank of the subtropical jet (STJ), sometimes manifesting a distinct wind maximum above the PFJ (Newton and Persson, 1962).

The upper tropospheric surface of maximum wind (SMW) (Strong and Davis, 2005) has been used to study how the vertical positions of tropospheric jet cores and other upper tropospheric wind maxima vary with changes in the atmospheric thermal structure. At a given observation time in a gridded data set, the SMW is defined as the surface passing through the fastest analyzed wind above each grid node, with a vertical search domain restricted to the upper troposphere and any tropospheric jet streams extending into the lower stratosphere. The dominant patterns of joint space-time variability in the winter SMW are related to the AO and ENSO (Strong and Davis, 2006).

The focus of this research is the joint variability of the atmospheric thermal structure and the vertical position of the SMW over the NH during summer. Section 1.2 covers background theory that informs the format of the results sections. Data and the method for locating the SMW are described in Section 2, and the mean spatial structure of the summer SMW is provided in Section 3.1. The leading principal components of joint space-time variability in summer SMW pressure are identified in Section 3.2. Patterns of SMW pressure variability associated with the leading principal component are explored on the hemispheric and regional scale in Sections 3.3–3.8. Section 3.9 explores links between the SMW-related upper tropospheric temperature variability and SST variability. Summary and discussion are provided in Section 4.

1.2. Theory

In this section, we provide a brief overview of the relationship between SMW pressure (denoted $\tilde{P}$) and the thermal structure of the upper troposphere, developing a conceptual framework that will inform the format of the results sections. Figure 1(a) and (b) shows idealized geostrophic wind speed profiles for one node at time $t_1$ and time $t_2$, and their local maxima (150 hPa in 1a; 250 hPa in 1b) mark $\tilde{P}$. Expressing vertical derivatives using $-\partial/\partial \ln p$ so that the vertical coordinate increases in value upward, $\tilde{P}$ is the pressure where $-\partial V_g/\partial \ln p = 0$ and $-\partial^2 V_g/\partial \ln p^2 < 0$. The first derivative $-\partial V_g/\partial \ln p$ is commonly referred to as the thermal wind shear (e.g. Bluestein, 1992), and can be expressed as a function of the spatial distribution of temperature and geopotential height in the atmosphere as

$$
-\frac{\partial V_g}{\partial \ln p} = \frac{R_d}{f} \| \vec{\nabla} T_v \| \cos \beta
$$

Figure 1. For a node at two times ($t_1$ and $t_2$), idealized profiles of (a, b) geostrophic wind speed $V_g$, (c, d) the magnitude of the thermal wind shear $-\partial V_g/\partial \ln p$, and (e) the $t_2$ minus $t_1$ difference in the thermal wind shear $\Delta[-\partial V_g/\partial \ln p]$. Filled circles indicate the pressure of the surface of maximum wind ($\tilde{P}$) at $t_1$ and $t_2$. 

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where \( p \) is pressure, \( R_d \) is the dry air gas constant, \( f \) is the Coriolis parameter, \( \| \vec{\nabla} T_v \| \) is the magnitude of the virtual temperature gradient \( (\vec{\nabla} T_v) \), and \( \beta \) is the angle between \( -\vec{\nabla} T_v \) and the geopotential height gradient (derivation in the Appendix).

The thermal wind shear profiles in Figure 1(c) and (d) were developed by differentiating the idealized \( V_g \) profiles in Figure 1(a) and (b) in accordance with Equation (1). Because the thermal wind shear is positive below \( \tilde{P} \) and negative above \( \tilde{P} \) (Figure 1(c), (d)), changes in \( \tilde{P} \) should be accompanied by changes in the thermal wind shear on nearby isobaric surfaces. This is illustrated in Figure 1(e), which shows the decrease in the thermal wind shear after the SMW descends from 150 hPa at \( t_1 \) to 250 hPa at \( t_2 \). In the results section of this manuscript, we will show the relationship between \( \tilde{P} \) variability and the spatial distribution of atmospheric temperature by analyzing time series of seasonal mean \( \tilde{P} \) and seasonal mean thermal wind shear on an isobaric surface within the vertical range of the SMW motion.

2. DATA AND METHODS

2.1. Data

From 2.5° grids in the NCEP-NCAR Reanalysis (Kalnay et al., 1996), we use six-hourly wind speed, geopotential height, humidity, and air temperature on seven isobaric surfaces from 500 to 100 hPa, and the pressure of the lapse-rate tropopause. Artificial 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). Kanamitsu et al. (1997) found that satellite data impacted the NH Reanalysis less than they impacted the data-sparse stratosphere and eastern oceanic areas of the Southern Hemisphere. Conclusions presented here include references to shifts in upper tropospheric temperature that occurred in the mid-to-late 1970s. Although the shifts nearly coincide with the introduction of satellite data, they appear in the radiosonde record as well (Lindzen and Giannitsis, 2002) and coincide with the mid-to-late 1970s climate shift identified by Nitta and Yamada (1989) and Trenberth (1990).

The Climate Prediction Center (CPC) AO index (AOI) values used here are developed by projecting the monthly mean 1000 hPa geopotential height anomalies onto the first principal component of the monthly mean 1000 hPa geopotential height field north of 20°N. Southern Oscillation Index (SOI) values are from the Climate Research Unit and are the normalized pressure difference between Tahiti and Darwin (Ropelewski and Jones, 1987; Allan et al., 1991; Können et al., 1998). PDO index (PDOI) values are obtained from the Joint Institute for the Study of the Atmosphere and Ocean, and are the leading principal component of monthly ‘residual’ SST anomalies in the North Pacific poleward of 20°N, in which residual indicates that the monthly mean global average SST anomalies are removed prior to performing the principal components analysis (PCA) (Zhang et al., 1997; Mantua et al., 1997). AMO values obtained from NOAA-CIRES Climate Diagnostic Center (CDC) are the 10-year running mean Atlantic SST anomalies north of the equator (Enfield et al., 2001) based on the Met Office Hadley Center’s sea ice and SST data set (Rayner et al., 2003). Also from the CDC, we use a multivariate El Niño/Southern Oscillation Index (MEI) that captures variability in sea-level pressure, zonal and meridional components of the surface wind, SST, surface air temperature, and total cloudiness fraction of the sky over the tropical Pacific (Wolter and Timlin, 1993, 1998). Finally from the CPC, we use the monthly mean SST for the global equatorial belt 10°S–10°N (denoted SST_{10°S–10°N}) (Smith and Reynolds, 1998).

2.2. The surface of maximum wind

At a given observation time in a gridded data set, the SMW is defined as the surface passing through the fastest analyzed wind above each grid node, with a vertical search domain restricted to the upper troposphere and any tropospheric jet streams extending into the lower stratosphere. Figure 2 is used to illustrate how SMW pressure is identified in gridded data such as the Reanalysis (or in rawinsonde data that are analogous to one column of gridded data). An example from January is used because the utility of the domain restriction described above is most apparent in winter (when the polar night jet stream is present and the tropospheric...
jet streams extend poleward into the lower stratosphere through tropopause folds or breaks). In the column above each node (each tick mark on the abscissa in Figure 2), the wind speed data points are flagged as candidates or noncandidates for the SMW (with the distinguishing criteria to be described presently). After SMW candidates are flagged (open circles, Figure 2), the fastest speed from among the candidates above each node defines one point on the SMW (filled circles, Figure 2).

Any datum from 500 to 100 hPa is a SMW candidate unless the point is above the level closest to the lapse-rate tropopause and outside a 25.7 m/s (50-knot) contour that captures at least one tropospheric datum. In other words, SMW candidates are points in the upper troposphere or within upper tropospheric jet streams extending into the lower stratosphere. For example, the datum at 40° N and 150 hPa is above the measurement level closest to the tropopause, but remains a SMW candidate because the point is within a 25.7 m/s contour capturing at least one tropospheric datum. The datum at 75° N and 100 hPa is not a SMW candidate because it is above the measurement level closest to the tropopause and the 25.7 m/s contour encircling the point does not capture a tropospheric datum (i.e. is not associated with a tropospheric jet stream). Properties associated with the SMW will be denoted by a tilde (SMW pressure is $\tilde{P}$ and SMW wind speed is $\tilde{V}$).

The SMW is distinguished from the operationally defined LMW (Reiter, 1958) and the Reanalysis maximum wind layer (MWL) because the upper bound of the SMW search domain varies spatially and temporally with the position of the tropopause and the bounds of upper tropospheric jet streams. The MWL has a fixed search domain from 500 to 70 hPa and is shown for comparison in Figure 2. The sometimes large differences between the SMW and MWL poleward of 40° N result largely from the SMW’s intentional focus on upper tropospheric winds, whereas the differences equatorward of 40° N result from the use of cubic spline interpolation in the MWL.

2.3. Derived variables

The magnitude of a gradient is its Euclidian norm

$$\| \tilde{V}Z \| = (| \partial Z / \partial x |^2 + | \partial Z / \partial y |^2 )^{1/2}$$

where first derivatives are numerically approximated using second-order central differences for most nodes, second-order backward differences at the pole, and second-order forward differences at the equator. The cosine of the angle $\beta$ between two gradients is

$$\cos \beta = \tilde{V}Z \cdot \tilde{V}T (\| \tilde{V}Z \| \| \tilde{V}T \|)^{-1}. $$
If the gradient at a node rotates temporally, the relative importance of its components changes. To convey this graphically a ‘mean unit-length vector’ is calculated for a block of time and a series of mean unit-length vectors is presented on a timeline. The mean unit-length vector is calculated in three steps: six-hourly vectors are normalized to magnitude 1 (converted to unit-length vectors), the mean x-component and y-component of the unit-length vectors are calculated for the block of time (yielding the mean unit-length vector), and the mean unit-length vector is normalized to magnitude 1 for display. Using virtual temperature as an example, the mean unit-length vector normalized to magnitude 1 is denoted $-\vec{\nabla}T_v/\|\vec{\nabla}T_v\|$ (temporal averaging is implicit).

2.4. Statistical methods

PCA is used to identify patterns of simultaneous anomalies that account for variability in summer mean $\bar{P}$. The PCA variables are Reanalysis grid nodes and the PCA cases are the 47 summer (June–August) or winter (December–February) seasonal mean SMW pressures at each node for the years 1958–2004. Seasonal mean $\bar{P}$ values are weighted by the square root of the cosine of latitude to ensure proper representation in the correlation matrix. The unit-length eigenvectors and eigenvalues of the correlation matrix are obtained via singular value decomposition. The $m$th principal component time series (‘PC $m$’) is produced by projecting the standardized, weighted seasonal mean $\bar{P}$ data onto the $m$th eigenvector and then standardizing the time series by subtracting its mean and dividing by its standard deviation. When mapped, the $m$th unit-length eigenvector is referred to as the PC $m$ ‘loading pattern.’ The total variance explained by PC $m$ is the PC $m$ eigenvalue divided by the summation of the correlation matrix eigenvalues.

Correlations are measured with least squares linear regression and, unless otherwise specified, statistics for all reported correlations are significant at $\alpha = 0.05$ based on effective degrees of freedom ‘$df$’ (adjusted to account for Pearson lag-1 autocorrelation in the regression residuals). MSE is used to denote the mean squared error of the regression. When the results of statistical tests at multiple locations are mapped, the collection of results is tested for field significance at the 95% confidence level using Monte Carlo simulations (Livezey and Chen, 1983). Passing the field significance test indicates there is less than a 5% probability that the amount of area with significant local results occurred by chance given the field’s spatial degrees of freedom. Time series change points are detected using an iterative, multiple change point detection algorithm based on data ranking (Siegel and Castellan, 1988; Lanzante, 1996).

3. RESULTS

3.1. Summer SMW spatial variability

The summer mean SMW pressure is highest near the pole and lowest over the tropical eastern hemisphere, especially in the vicinity of the tropical easterly jet (TEJ) (dark shading centered over India in Figure 3(a)). Figure 3(b) shows a cross-section from Figure 3(a) at 45°E, illustrating how the SMW relates to the mean tropopause and isotachs. The SMW slopes down toward the equator within the TEJ (12°N 150 hPa, Figure 3(b)). The subtropical jet core at 40°N marks a local minimum of SMW pressure flanked by zones of fairly steep SMW pressure increase. In higher latitudes, the mean SMW generally follows the slope of the mean tropopause, residing approximately 60 hPa below it. The spatial relationship between the high-latitude SMW and tropopause is fairly robust from summer to summer. For all latitudes poleward of 45°N, zonally averaged summer SMW pressure and zonally averaged summer lapse-rate tropopause pressure ($p_{LRT}$) are correlated, accounting for up to 42% of the SMW pressure variance (Figure 3(c)). Equatorward of 45°N, the SMW offset from the tropopause is generally larger and more variable (Figure 3(b)), and the $\bar{P}$ versus $p_{LRT}$ correlation is not significant (Figure 3(c)).

3.2. Summer $\bar{P}$ principal components

Results of the PCA of summer $\bar{P}$ are shown in Figure 4. Summer $\bar{P}$ PC 1 accounts for more than three times the variance of the second PC (Figure 4(a) vs 4(c)). The loading pattern of summer $\bar{P}$ PC 1 is concentrated
Figure 3. (a) Mean surface of maximum wind pressure (\( \tilde{P} \)) for summers 1958–2004. The line marks the meridian used for the cross-section in the lower panel. (b) Cross-section showing \( \tilde{P} \) from upper panel (bold dashed line) with the mean pressure of the summer lapse-rate tropopause (bold solid line) and mean isotachs (light gray lines showing \( ||V|| \) at 10 m/s interval) along 45°E for summers 1958–2004. Dashed lines bound the 10th to 90th percentile of \( \tilde{P} \) and the gray area bounds the 10th to 90th percentile of seasonal mean tropopause pressure. (c) Pearson correlation coefficient \( r \) for regression of zonally averaged summer mean \( \tilde{P} \) versus zonally averaged summer mean lapse-rate tropopause pressure \( p_{LRT} \), with circles indicating correlations significant at \( \alpha = 0.05 \)

in the tropics and subtropics, with positive values over most of the eastern hemisphere and Atlantic, and negative values appearing over the equatorial East Pacific and South America (Figure 4(a)). Summer \( \tilde{P} \) PC 1 has a positive trend (Figure 4(b)) that indicates a decadal-scale SMW pressure increase (descent) over most of the eastern hemisphere tropics and subtropics, and a SMW pressure decrease (ascent) over the equatorial central and eastern Pacific. The SMW variability associated with summer \( \tilde{P} \) PC 1 is examined in more detail in Sections 3.3–3.9.

Summer \( \tilde{P} \) PC 2 is correlated with the summer AOI (Figure 4(d)). \( \tilde{P} \) PC 2 has a zone of positive loadings near 35°N from the East Pacific across the United States and from northern Africa across central Asia (Figure 4(c)). During the positive phase of the summer AO, SMW pressures are anomalously high in this zone (positive loadings) and zonal wind speeds are anomalously low (Ogi et al., 2004b). Consistent with the
thermal wind relationship, this zone of increased SMW pressure and decreased zonal winds is located just equatorward of a belt where temperatures are anomalously high during the positive phase of the summer AO (Ogi et al., 2004b).

Summer $\tilde{P}$ PC 3 is correlated with the AMO (Figure 4(f)). $\tilde{P}$ PC 3 has strong negative loadings in the North Atlantic and Northeast Pacific (where SSTs are anomalously high during the AMO warm phase). $\tilde{P}$ PC 3 also has strong loadings through portions of the tropics, particularly over the tropical Pacific. The AMO reflects variations in the thermohaline circulation (THC) of the North Atlantic (Delworth and Mann, 2000), and recent modeling work (Zhang and Delworth, 2005) links changes in the THC to shifts in the position of the intertropical convergence zone and changes in the strength of the Walker and Hadley circulations. The presence of strong, opposite loadings in the eastern and western tropical Pacific may also suggest a
relationship between $\tilde{P}$ PC 3 and the Southern Oscillation. $\tilde{P}$ PC 3 is significantly correlated with the SOI ($r = 0.38$, df = 24, MSE = 0.86), but its relationship to the AMO is stronger.

3.3. Summer $\tilde{P}$ trend

The trend in $\tilde{P}$ PC 1 is associated with significant SMW pressure trends (primarily positive) as large as 30 hPa/decade at nodes accounting for more than two thirds of the area from 0 to 30°N (Figure 5(a)). Since $\tilde{P}$ changes are related to temperature gradient changes (Section 1.2), we expect regions with large $\tilde{P}$ trends to be flanked by air temperature trends of opposing sign near the jet stream level. Indeed, the axis of large positive $\tilde{P}$ trends near 15°N from the Atlantic east toward India in Figure 5(a) is bounded by equatorward temperature increases and poleward temperature decreases (compare Figure 5(a), (b)). The relationship between jet-level air temperature and the $\tilde{P}$ trends to the west of 60°W in Figure 5(a) is explored in the regional analysis in Section 3.7.

Although the focus of this research is variability in the pressure of the SMW, speed trends on the SMW are shown in Figure 5(c) and briefly discussed in this paragraph to provide a context for the observed changes in $\tilde{P}$. Many of the areas of SMW descent in Figure 5(a) also exhibit decreasing wind speeds on the SMW (negative $\tilde{V}$ trends, Figure 5(c)). Wind speed deceleration often accompanies descent of the SMW because $V_g$ on the SMW is the integral of the thermal wind Equation (1) from the surface to $\tilde{P}$, and convergence of the limits of integration (SMW descent toward the surface) yields a smaller integral unless there is a compensating increase in the integrand $\| \nabla T_v \| \cos \beta$. In other words, a jet stream core that descends will have slower wind speeds unless the underlying temperature gradients strengthen. Variability in $\tilde{V}$ may thus depend on changes in the thermal structure of the entire underlying troposphere, whereas $\tilde{P}$ is sensitive to relatively shallow upper tropospheric temperature changes that relocate the point where $\| \nabla T_v \| \cos \beta$ decreases through zero.

![Figure 5](image-url)

Figure 5. 1958–2004 trend in three summer variables: (a) surface of maximum wind (SMW) pressure ($\tilde{P}$) (contour interval is 5 hPa/decade with negative values dashed), (b) air temperature for the 250 to 150 hPa layer (contour interval is 0.1 K/decade with zero contour bold and negative values dashed), and (c) SMW wind speed ($\tilde{V}$) (contour interval is 0.5 m/s per decade with negative values dashed). Panels (a–c) are each field significant at $\alpha = 0.05$. The small gray boxes in panel (a) mark locations where regional analyses are performed in Sections 3.4–3.7.
Figure 6. For each summer (June–August) 1958–2004 over the tropical Atlantic (30°W, 10–15°N): (a) mean surface of maximum wind pressure \( \tilde{P} \) and mean 250-hPa thermal wind shear \( -\partial V_g/\partial \ln p = K_d f^{-1} \| \nabla T_v \| \cos \beta \), (b) mean 250-hPa temperature gradient magnitude \( \| \nabla T_v \| \) and cosine of the angle \( \beta \) between \( \nabla T_v \) and \( \nabla Z \), and (c) 250-hPa mean unit-length vectors of geopotential height gradient \( -\nabla Z/\| \nabla Z \| \) and temperature gradient \( -\nabla T/\| \nabla T \| \) for 5-year blocks centered on 1960, 1970...2000. (d) 1958–1962 mean 250-hPa temperature with the 1960 \( -\nabla T/\| \nabla T \| \) shown, (e) trend of summer mean 250-hPa temperature contoured at 0.1 K/decade with negative values dashed, and (f) 1998–2002 mean 250-hPa temperature with the 2000 \( -\nabla T/\| \nabla T \| \) shown. The small boxes in panels (d–f) show the averaging domain for data in panels (a–c).

Regions around the locations marked with small gray boxes in Figure 5(a) (the tropical Atlantic, tropical East Africa, and the tropical West and East Pacific) are further examined in Sections 3.4–3.7 to more quantitatively demonstrate the relationship between SMW pressure trends and upper tropospheric thermal structure changes. Changes in the isobaric thermal pattern responsible for the observed temporal patterns of \( \tilde{P} \) and the thermal wind shear are then mapped and discussed.

3.4. \( \tilde{P} \) and temperature pattern trends over the tropical Atlantic

The SMW over the tropical Atlantic descended from 150 to 250 hPa between 1958 and 1990, and resided close to 250 hPa thereafter (Figure 6(a)). The descent of the SMW was accompanied by a decrease of the 250-hPa thermal wind shear toward zero (Figure 6(a)), consistent with the thermal wind equation as discussed in Section 1.2. The decrease of the thermal wind shear resulted from weakening of the 250-hPa temperature gradient \( \| \nabla T_v \| \to 0 \), Figure 6(b)) and reorienting of the temperature gradient toward orthogonality with the geopotential height gradient \( \cos \beta \to 0 \), Figure 6(b); orientation of \( -\nabla T_v/\| \nabla T_v \| \) and \( -\nabla Z/\| \nabla Z \| \), Figure 6(c)). For the five-summer period around 1990, the SMW periodically descended below 250 hPa (Figure 6(a)) and, consistent with thermal wind theory, the 1990 \( -\nabla T_v/\| \nabla T_v \| \) was at an angle larger than 90° from \( -\nabla Z/\| \nabla Z \| \) (Figure 6(c)).

Figure 6(d–f) illustrate the changes in 250 hPa temperature that caused the weakening and reorientation of \( -\nabla T_v \). Early in the record (1958–1962), the 250 hPa temperature gradient was approximately 2.2 \times 10^{-3} K/km and was directed toward the south-southwest (Figure 6(d)). A 1600 km-wide cooling trend southwest
of the Cape Verde Islands interacted with a warming trend spanning 0–7.5°N (Figure 6(e)) to essentially eliminate the meridional temperature gradient (\(\partial T/\partial y\) component of \(-\nabla T_v\)) by the year 2000 (Figure 6(f)). Late in the record, residual \(-\nabla T_v\) was approximately 30% weaker (Figure 6(b)) and nearly orthogonal to \(-\nabla Z\) (Figure 6(c), (f)), no longer supporting wind speeds increasing with height through 250 hPa. Summarizing for the tropical Atlantic, the upper tropospheric temperature pattern early in the record supported wind speeds increasing up to 150 hPa, but was later modified by a dipole of warming and cooling that weakened and rotated the temperature gradient toward orthogonality with the geopotential gradient such that the SMW descended to 250 hPa.

3.5. Tropical East Africa

The SMW over tropical East Africa descended gradually between 1960 and 1975, then abruptly descended toward 200 hPa coincident with the mid-to-late 1970s climate shift, and ended the period of record near 160 hPa (Figure 7(a)). Considering the entire period of record, SMW motion is well-correlated with the thermal wind she at 150 hPa (time series and statistics in Figure 7(a)). Comparison of the 150 hPa temperature structure associated with \(\tilde{P}\) before and after the 1970s climate shift provides further insight. Between 1960 and 1990, the thermal wind shear decrease resulted from the 150 hPa temperature gradient weakening (\(\|\nabla T_v\|\to 0\), Figure 7(b)) and rotating toward orthogonality with the geopotential height gradient (cos \(\beta\) → 0, Figure 7(b); orientation of \(-\nabla T_v/\|\nabla T_v\|\) and \(-\nabla Z/\|\nabla Z\|\), Figure 7(c)).

Figure 7(d–f) illustrate the changes in upper tropospheric temperature that caused the 1960–1990 weakening and reorientation of \(-\nabla T_v\). Early in the record (1958–1962), the 150-hPa temperature gradient was weakened and rotated toward orthogonality with the thermal wind shear.
directed almost due south (Figure 7(d)) with a magnitude of approximately $2.8 \times 10^{-3}$ K/km (Figure 7(b)). Widespread 150-hPa warming over Africa from the equator to 15°N and cooling from 15 to 30°N (Figure 7(e)) interacted to weaken the meridional temperature gradient (Figure 7(b)) and orient it nearly orthogonal to $-\nabla Z$ by 1990 (Figure 7(c), (f)). The near orthogonality of $-\nabla T_v$ and $-\nabla Z$ no longer reliably supported wind speed increasing with height through 150 hPa, and the SMW fluctuated around 170 hPa (Figure 7(a)). Summarizing for tropical East Africa, spatially extensive ($11 \times 10^6$ km$^2$) regions of cooling and warming in the upper troposphere caused a rather abrupt rotation and weakening of the temperature gradient coincident with the mid-to-late 1970s climate shift, and the SMW descended to around 170 hPa.

3.6. Tropical West Pacific

The seasonal mean SMW in the West Pacific near 15°N has been descending at approximately 16 hPa/decade in concert with the 250-hPa thermal wind shear (Figure 8(a)). The decreasing trend in the thermal wind shear is driven in large part by rotation of the temperature gradient toward orthogonality with the geopotential height gradient ($\cos \beta$, Figure 8(b); $-\nabla T_v/\|\nabla T_v\|$ and $-\nabla Z/\|\nabla Z\|$, Figure 8(c)), and is driven in small part by a modest weakening trend in $\|\nabla T\|$ (Figure 8(b)). The weakening and eastward rotation of the temperature gradient is also made apparent by comparing the early versus late 250 hPa temperature field (Figure 8(d) vs 8(f)). The temperature changes underlying the weakening and rotation of $-\nabla T_v$ consist of a cooling over East Asia and equatorial warming over the ocean east of Southeast Asia (Figure 8(e)). In summary for the tropical West Pacific, upper tropospheric cooling and warming trends caused 100 hPa of SMW descent by weakening the temperature gradient and aligning it almost orthogonal to the geopotential height field.

![Figure 8](image-url)
3.7. Tropical East Pacific

Recall from Figure 5(a) that the East Pacific has a zonally oriented band of SMW descent ('descent-region') poleward of a zonally oriented band of SMW ascent ('ascent region'). Both phenomena are associated with an upper tropospheric warming trend located over the ocean near 10°N (0.2 K/decade contour, Figure 9(e)). The tendency toward more frequent El Niño events during the second half of the record accounts for a portion of the warming. The 250-hPa temperature within the region bounded by the 0.2 K/decade contour in Figure 9(e) is positively correlated with the March MEI ($r = 0.75$, df = 31, MSE = 0.14). The relationship between SST variability and SMW-related upper tropospheric temperature changes is explored further in Section 3.9.

The 250 hPa temperature topography was relatively flat prior to the 0.2 K/decade warming trend, and the descent-region gradient had an eastward orientation (97.5°W, 15°N in Figure 9(d)). The 250-hPa temperature trend reoriented the descent-region temperature gradient away from the warming center (Figure 9(f)) and out of alignment with the geopotential gradient (Figure 9(c)). The resultant decrease in $\cos \beta$ at 250 hPa (Figure 9(b)) weakened the thermal wind shear and the SMW descended (Figure 9(a)).

In the ascent region, the 250-hPa temperature trend rotated the temperature gradient away from the warming center (Figure 9(f)) and into alignment with the geopotential gradient (Figure 10(c)). The resultant increase in 250-hPa $\cos \beta$ (Figure 10(b)) caused the thermal wind shear to increase and the SMW to ascend (Figure 10(a)). The magnitude of the temperature gradient lacks a significant trend in the descent and ascent regions (Figures 9(b), 10(b)), and so played a minimal role in the SMW pressure changes. Summarizing for the tropical East Pacific, a center of upper tropospheric warming reoriented the temperature gradient on its poleward side out of alignment with the geopotential field, causing SMW descent. The same warming reoriented the temperature gradient on its equatorward side into alignment with the geopotential field, causing SMW ascent.

![Figure 9](image-url)
3.8. Relative importance of the temperature gradient’s magnitude and orientation

Trends in the thermal wind shear and \( \vec{P} \) may result from a trend in only \( \parallel \nabla T_v \parallel \), a trend in only cos \( \beta \), or reinforcing (like-signed) trends in both variables. We now consider the relative importance of \( \parallel \nabla T_v \parallel \) and cos \( \beta \) trends on a summary level for the portion of the hemisphere where the SMW pressure and the thermal wind shear exhibit their strongest trends (0–25 °N). Two hundred hPa is a useful surface for this analysis because most of the SMW pressure time series in Sections 3.4–3.7 pass through or reach this pressure, and it is the closest data surface to the zonally averaged SMW pressure from 0 to 25 °N.

Approximately 66% of the significant SMW pressure trends (by area) are driven by reinforcing, significant trends in both \( \parallel \nabla T_v \parallel \) and cos \( \beta \) (e.g. the tropical Atlantic). Approximately 23% of SMW pressure trends are driven by a cos \( \beta \) trend without a reinforcing \( \parallel \nabla T_v \parallel \) trend (e.g. tropical East Pacific). SMW pressure trends driven by a \( \parallel \nabla T_v \parallel \) trend without a reinforcing cos \( \beta \) trend are less common (11% of cases) and tend to be weak relative to the cos \( \beta \)-driven changes (<5 hPa/decade) in part because seasonal mean \( \parallel \nabla T_v \parallel \) is typically of the order \( 10^{-6}/\text{Km} \), whereas cos \( \beta \) is of the order \( 10^0 \). Locations lacking significant SMW pressure trends most often (88%) lack significant cos \( \beta \) trends.

3.9. Link to sea surface temperature

We now present a cross-section view of zonally averaged temperature gradient trends in the vicinity of the SMW, and explore how these trends relate to SST variability. Figure 11(a) shows the trend in zonally averaged summer thermal wind shear. A large region with thermal wind shear decreases exceeding 1 m/s per decade is found in the upper troposphere between 5 and 20 °N. The summer SMW resided in this region of
Figure 11. (a) Trend in the zonally averaged thermal wind shear for summers (June–August) 1958–2004 contoured at 0.2 m/s per decade with negative values dashed. The bold gray line indicates the pressure of the 47-year mean tropopause and the gray boxes show domains used to produce the zonal mean air temperatures $T_{\text{Equ}}$ and $T_{\text{Sub}}$. (b) Trend in the zonally averaged surface of maximum wind pressure ($\tilde{P}$) with filled circles indicating results significant at $\alpha = 0.05$. (c) $\tilde{P}$ PC 1 (from Figure 4) shown with the difference $T_{\text{Equ}}$ minus $T_{\text{Sub}}$. (d) $T_{\text{Equ}}, T_{\text{Sub}}$, the July Pacific Decadal Oscillation Index (PDOI) and a third-order harmonic fit to the July PDOI. Decreasing thermal wind shear between 1958 and 2004 (right axis of Figure 11(a)), resulting in increasing trends in zonally averaged $\tilde{P}$ ranging from 3 to 10 hPa/decade (Figure 11(b)).

The weakening of the thermal wind shear is in part attributable to weakening of the meridional component of $|\nabla T|$ by convergence of the air temperature in the latitude belts to the north and south of the weakening. To demonstrate this quantitatively, two boxes are drawn flanking the region of decreasing thermal wind shear in Figure 11(a): an equatorial box (100–300 hPa, 0–5°N with zonally averaged summer mean temperature denoted $T_{\text{Equ}}$) and a subtropical box (100–300 hPa, 20–30°N with zonally averaged summer mean temperature denoted $T_{\text{Sub}}$). The difference $T_{\text{Equ}} - T_{\text{Sub}}$ approaches zero over the period 1958–2004 and accounts for more than 95% of the variability in the leading principal component of summer SMW pressure ($\tilde{P}$ PC 1, Figure 11(c)). The strong correlation in Figure 11(c) indicates that $\tilde{P}$ PC 1 represents a pattern of joint space-time variability driven largely by the difference in zonally averaged temperature between the equatorial and subtropical upper troposphere. $T_{\text{Equ}}$ and $T_{\text{Sub}}$ converge during the first 20 years of the record because $T_{\text{Sub}}$ decreases more rapidly than $T_{\text{Equ}}$ (Figure 11(d)). $T_{\text{Equ}}$ and $T_{\text{Sub}}$ undergo positive shifts significant at $\alpha = 0.05$ in 1978, and the faster increase in $T_{\text{Equ}}$ produces a convergence of the $T_{\text{Equ}}$ and $T_{\text{Sub}}$ time series (Figure 11(d)).

The rapid convergence of $T_{\text{Equ}}$ and $T_{\text{Sub}}$ in 1978 is coincident with the decadal climate shift documented in various atmospheric fields (Nitta and Yamada, 1989; and Trenberth, 1990; Graham, 1994; Miller et al., 1994) and biological data (Ebbesmeyer et al., 1991; Francis and Hare, 1994; McGowan et al., 1998; Hare and Mantua, 2000). The multidecadal variability associated with the climate shift is often quantified using the SST-based PDOI. The temporal patterns of $T_{\text{Equ}}$ and $T_{\text{Sub}}$ share some similarities with the July PDOI, including a significant change point in 1978 (Figure 11(d)).
The PDOI discussed in the preceding paragraph is based on SST anomalies poleward of 20°N in the Pacific. The notion that SST variability may influence the temporal patterns observed in $T_{\text{Equ}}$ and $T_{\text{Sub}}$ is further strengthened by considering SST-related time series geographically closer to $T_{\text{Equ}}$ and $T_{\text{Sub}}$. Strong links have been established between air temperature in the deep tropical troposphere and SST anomalies associated with El Niño (Horel and Wallace, 1981; Yulaeva and Wallace, 1994; Sobel et al., 2002). Here, the MEI accounts for variability in summer $T_{\text{Sub}}$ at the $\alpha = 0.05$ level with $r \geq 0.50$ at lags of $-10$ to $-3$ months, where negative numbers indicate that the MEI lags $T_{\text{Sub}}$ (Figure 12(a)). The $T_{\text{Sub}}$ versus MEI correlation peaks when the MEI is taken from March of the contemporaneous year (lag $-4$ months, Figure 12(a), (b)). For the equatorial latitudes, SST$_{10^\circS-10^\circN}$ accounts for variability in summer $T_{\text{Equ}}$ at the $\alpha = 0.05$ level with $r \geq 0.50$

Figure 12. (a) Bold line shows lagged correlation between $T_{\text{Equ}}$ and mean sea surface temperature around the hemisphere between $10^\circS$ and $10^\circN$ (SST$_{10^\circS-10^\circN}$) where negative lags indicate that SST$_{10^\circS-10^\circN}$ lags $T_{\text{Equ}}$ and symbols show correlations significant at $\alpha = 0.05$. The thin line shows lagged correlation between $T_{\text{Sub}}$ and the multivariate El Niño/Southern Oscillation Index (MEI) where negative lags indicate that the MEI lags $T_{\text{Sub}}$ and symbols show correlations significant at $\alpha = 0.05$, (b) $T_{\text{Sub}}$ and March MEI, and (c) $T_{\text{Equ}}$ and May SST$_{10^\circS-10^\circN}$

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at lags of 0 to −10, with peak correlation when SST$_{10^\circ S–10^\circ N}$ is taken from May of the contemporaneous year (lag −2 months, Figure 12(a), (c)).

Additional phenomena potentially influencing upper troposphere air temperature near the SMW include solar output variability, volcanic emissions, and changing greenhouse-gas concentrations. Considering the net effect of changes in sulfate aerosol, ozone, and carbon dioxide concentrations over the past several decades, modeling work indicates a maximum net warming in the upper troposphere over the equator, transitioning poleward to a maximum net cooling in the lower stratosphere over the polar latitudes (Santer et al., 1996). Poleward decreases in the net warming rate may contribute to the decadal SMW pressure changes reported here. Such a temperature change pattern is consistent with the observed convergence of $T_{\text{Equ}}$ and $T_{\text{Sub}}$, because $T_{\text{Sub}}$ is higher than $T_{\text{Equ}}$ during the summer (Figure 11(d)).

4. SUMMARY AND DISCUSSION

The summer SMW generally slopes up equatorward, following and residing below the tropopause in polar latitudes and exhibiting undulations associated with jet cores in middle and lower latitudes. Changes in the altitude of fast upper tropospheric wind features are climatologically important because of the role of the jet streams in global circulation and surface weather (Riehl et al., 1954; Skaggs, 1967; Nakamura, 1992; Christoph et al., 1997; Rao et al., 2004). The documentation of vertical displacement of upper tropospheric wind maxima is also important because such motion may complicate the use of isobaric surfaces for climate studies of the jet stream (Strong and Davis, 2005). In this research, we used the SMW to map the 47-year mean position of fast upper tropospheric winds during summer. Variability of summer SMW pressure was then documented and linked to variability in the thermal structure of the upper troposphere in accordance with the relationships expressed in the thermal wind equation. The temperature gradient changes influencing the SMW were then linked to SST variability on quasi-biennial to decadal timescales, ENSO variability, and the AO.

The dominant pattern of variability (PC 1) is related to spatially expansive trends as large as 30 hPa/decade over regions of the tropics and subtropics. Over the Atlantic, a dipole of cooling poleward of warming weakened and reoriented the upper tropospheric temperature gradient, causing the SMW to descend from 150 hPa to below 250 hPa. Similar temperature field changes over Africa caused the SMW to descend from 120 to 180 hPa, including a rather abrupt transition coincident with the mid-to-late 1970s climate shift. Over the Pacific, changes in the magnitude of the temperature gradient were less important than changes in its orientation. Rotation of the West Pacific temperature gradient caused approximately 100 hPa of SMW descent. In the tropical East Pacific, a warming trend changed the horizontal orientation of the temperature gradients such that zonally oriented bands of descent and ascent were generated.

The gradient of zonally averaged temperature in the upper troposphere between equatorial and subtropical latitudes accounts for more than 95% of the variability of the leading principal component of SMW pressure. Subtropical upper tropospheric air temperatures are significantly correlated with the MEI, and equatorial upper tropospheric air temperatures are significantly correlated with the mean SST between $10^\circ S$ and $10^\circ N$ (SST$_{10^\circ S–10^\circ N}$). SST$_{10^\circ S–10^\circ N}$ rapidly increased coincident with the decadal climate shift during the mid-to-late 1970s, and the correlated weakening of upper tropospheric temperature gradients accounts for much of the observed SMW pressure increases.

The results for the summer SMW presented here include $\hat{P}$ PC 1 pressure trends in the tropics and subtropics, and extratropical oscillations associated with the AO appearing as $\hat{P}$ PC 2. This is in contrast to the winter season in which the leading pattern of $\hat{P}$ variability from the hemispheric PCA is more oscillatory and AO-like, with ENSO-related tropical changes appearing as PC 2 (Strong and Davis, 2006). In addition to providing a framework for objectively documenting the speed and vertical position of fast upper tropospheric winds, the SMW serves as an informative climate diagnostic when temperature gradients in the stratosphere and mid-to-upper troposphere may be changing. Interesting directions for future work include querying a global climate model for SMW pressure trends and oscillations to reinforce understanding of the underlying mechanisms and determine the sensitivity of the SMW speed and pressure to global climate change. Research
is currently underway to study the joint variability of SMW speed and pressure, particularly in the context of greenhouse-gas related temperature variability at continental middle and high latitudes during winter.

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APPENDIX

Given that the magnitude of the geostrophic wind \( V_g = (u_g^2 + v_g^2)^{1/2} \) is nonzero and its vector components are differentiable functions of pressure,

\[
\frac{\partial V_g}{\partial p} = \frac{1}{2} (u_g^2 + v_g^2)^{-1/2} \left( 2u_g \frac{\partial u_g}{\partial p} + 2v \frac{\partial v_g}{\partial p} \right) \tag{A1}
\]

Substituting the magnitude of the geostrophic wind \( (f^{-1} \| \vec{\nabla} Z \|) \) for \( (u_g^2 + v_g^2)^{1/2} \), the geostrophic wind components for \( u_g \) and \( v_g \), and the thermal wind components in isobaric coordinates for \( \partial u_g/\partial p \) and \( \partial v_g/\partial p \), we obtain

\[
\frac{\partial V_g}{\partial p} = \frac{f}{\| \vec{\nabla} Z \|} \left[ \left( -\frac{1}{f} \frac{\partial Z}{\partial y} \right) \left( \frac{R_d}{f} \frac{\partial T_v}{\partial y} \right) + \left( \frac{1}{f} \frac{\partial Z}{\partial x} \right) \left( -\frac{R_d}{f} \frac{\partial T_v}{\partial x} \right) \right]
\]

\[
\begin{align*}
\frac{\partial V_g}{\partial p} &= \frac{R_d}{f\| \vec{\nabla} Z \|} \left[ \frac{\partial Z}{\partial y} \frac{\partial T_v}{\partial y} + \frac{\partial Z}{\partial x} \frac{\partial T_v}{\partial x} \right] \\
\frac{\partial V_g}{\partial \ln p} &= -\frac{R_d}{f\| \vec{\nabla} Z \|} (\vec{\nabla} Z \cdot \vec{\nabla} T_v) = -R_d f^{-1} \| \vec{\nabla} T_v \| \cos \beta \tag{A2}
\end{align*}
\]

where \( \beta \) is the angle between \( \vec{\nabla} T_v \) and \( \vec{\nabla} Z \). The quantity \( -\| \vec{\nabla} T_v \| \cos \beta \) is the virtual temperature gradient projected onto the geopotential height gradient, so Equation (A2) may be converted to natural coordinates

\[
\frac{\partial V_g}{\partial \ln p} = \frac{R_d}{f} \frac{\partial T_v}{\partial n} \tag{A3}
\]

where \( n \) is oriented normal to \( \vec{\nabla} T_v \) and defined positive to the left of the flow direction.

REFERENCES


SUMMER SURFACE OF MAXIMUM WIND VERTICAL POSITION TRENDS


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