Variability in the Position and Strength of Winter Jet Stream Cores Related to Northern Hemisphere Teleconnections

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ABSTRACT

Numerous teleconnections have been identified based upon spatial variability in sea level pressure or lower-tropospheric geopotential height fields. These teleconnections, which are commonly strongest in winter when the mean meridional temperature gradient is large, typically are neither derived from nor linked to changes in the jet stream. Here, winter tropospheric jet stream cores over the Northern Hemisphere (NH) are recovered from 6-hourly gridded data and interannual variability in winter jet core position, speed, and pressure are investigated in the context of NH teleconnections. Common methods for researching jet stream speed and position variability may yield unrepresentative results because jet core pressure variability is ignored (only one isobaric surface is evaluated) or pressure variability effects are smoothed (values are vertically averaged across several isobaric surfaces). In this analysis, data are extracted at the surface of maximum wind, thus controlling for jet core pressure variability and allowing for a more representative tracking of three-dimensional jet core variations.

In the extratropics, the leading pattern of variability in jet core frequency is correlated with the Arctic Oscillation index (AOI) and appears as an oscillation about the spiral-shaped mean configuration of the winter jet stream. In contrast to previous research, the authors find no evidence of Pacific jet deceleration during positive AOI. The second leading mode of variability appears as a split (merged) winter-mean jet stream in the east Pacific together with a merged (split) winter-mean jet stream over North America, a pattern of change that correlates with the Pacific–North American pattern and is reflected in the amplitude of the long-wave ridge over western North America.

1. Introduction

Spatially organized patterns of correlated anomalies in Northern Hemisphere (NH) pressure, temperature, and geopotential height fields have been identified on weekly, decadal, and longer time scales. These modes of variability, commonly known as teleconnection patterns, indicate the variability associated with preferred planetary wave configurations, and relatively small changes in them are capable of significantly impacting regional weather patterns (e.g., van Loon and Williams 1976; Hurrell 1995; Hoerling et al. 1995). Teleconnection patterns may be initiated by recurrent, regional external forcing of the atmosphere [e.g., sea surface temperature anomalies (Bjerknes 1969)] or by disturbances that derive their energy from the zonally varying climatological basic state through barotropic instability (Simmons et al. 1983).

In the extratropics, there is a tendency for a meridionally oriented seesaw of surface pressure between the high latitudes and midlatitudes (Lorenz 1951) with weekly to decadal periodicity (Kutzbach 1970). This meridionally oriented variability pattern is most evident over the Atlantic and is known as the North Atlantic Oscillation (NAO) (e.g., Barnston and Liveze 1987). Links have been established between the NAO and variability in meteorological fields over the Atlan-

NH circulation variability has also been described in the context of a more hemispheric seesaw of atmospheric mass known as the Arctic Oscillation (AO) (Thompson and Wallace 1998), the leading empirical orthogonal function (EOF) of sea level pressure poleward of 20°N. The AO partially overlaps the NAO in the Pacific but features a zonally symmetric structure over the Pacific, covers more of the Arctic, and is discernible in geopotential height fields from the troposphere into the lower stratosphere (Thompson and Wallace 1998, 2000a,b). Stochastic modeling indicates the role that jet stream migration may play in the presence of such meridional dipoles (e.g., Wittman et al. 2005). Models of random motions constrained to conserve mass and momentum, moreover, show that meridional dipole structures robustly arise in geopotential height and zonal wind EOF analyses and may take on annular (AO like) or zonally localized (NAO like) configurations depending on the model’s zonal correlation structure (Gerber and Vallis 2005). Ambaum et al. (2001) found that symmetrical bands of anticorrelated upper-tropospheric zonal wind appear in the Pacific and Atlantic in conjunction with the AO, but the nature of the Pacific and Atlantic jet stream oscillations are dissimilar (the AO high phase is associated with separation and intensification of the Atlantic jets versus a weakening of the Pacific jet).

From the mid-Pacific to eastern North America, circulation variability relates to the Pacific–North American (PNA) pattern (Dickson and Namias, 1976; Wallace and Gutzler, 1981). The PNA is an alteration between ridge amplification over western North America (positive PNA) and more zonal midtropospheric flow (negative PNA). Variability in the PNA is closely tied to surface temperature and precipitation anomalies over the United States (Wallace and Gutzler 1981; Leathers et al. 1992) and the North Pacific (Gutzler et al. 1988; Wallace et al. 1993; Wallace et al. 1995, 1996). Tropical sea surface temperature anomalies have been extensively investigated as a possible source of variability in the PNA (e.g., Zhang et al. 1996).

The present documentation of the relationship between jet stream variations and teleconnections leverages two related methodological concepts. First, we analyze both jet core frequencies and jet core speeds, enabling us to separate jet core migration from jet core speed changes. If mean or zonally averaged wind speeds were analyzed instead, modeling (Fyfe and Lorencz 2005) and analytical (Monahan and Fyfe 2006) work demonstrate that the leading EOF would likely reflect primarily jet stream migration but also contain contributions from fluctuations in jet stream strength and width.

Second, we account for jet stream core pressure variations by working on the surface of maximum wind (SMW), an analysis frame that moves vertically with all tropospheric jet streams (Strong and Davis 2005; 2006a,b). The method in essence isolates jet cores in three-dimensional space (with use of the SMW accounting for vertical motion and focus on jet core statistics accounting for horizontal motion). Previously reported changes in the position or strength of jet streams in the context of teleconnection patterns may include unrepresentative results because jet core pressure variability was ignored (evaluating composite cross sections or data from a single isobaric surface), pressure variability effects were smoothed (speed values were vertically averaged across several isobaric surfaces), or jet stream horizontal motion was ignored (mean wind speeds were used rather than jet core statistics).

2. Data and methods

a. Data and circulation indices

We use 6-hourly National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis data (Kalnay et al. 1996) on 2.5° grids for winters (December–February) 1958–2006, including geopotential height, tropopause pressure, and wind speed components on seven isobaric surfaces from 500 to 100 hPa. Three circulation indices are obtained from the Climate Prediction Center (CPC): the Arctic Oscillation index (AOI), the North Atlantic Oscillation index (NAOI), and the Pacific–North American pattern index (PNAI). The CPC develops the AOI by projecting monthly-mean 1000-hPa geopotential height anomalies onto the loading pattern of the first EOF of year-round, deseasonalized monthly-mean 1000-hPa geopotential height poleward of 20°N for 1979–2000. The loading patterns for the NAOI and PNAI are based, respectively, on the first and second rotated EOFs (Richman 1986) of monthly-mean 500-hPa height anomalies for deseasonalized, year-round, NH data from 1950 to 2000. The CPC normalizes the AOI, NAOI, and PNAI by each index’s monthly standard deviation. Here, we average each index for December through February to represent winter.

b. Surface of maximum wind

The upper-tropospheric SMW is identified in the reanalysis by the algorithm introduced in Strong and Davis (2005). Summarizing briefly, the SMW at a given
observation time passes through the fastest wind above each grid point with a vertical search domain restricted to the upper troposphere (500 hPa to the measurement level nearest the tropopause) and any tropospheric jet streams extending into the lower stratosphere (tropopause to 100 hPa). Tropospheric jet streams that protrude into the lower stratosphere are defined by 25.7 m s⁻¹ (50 kt) isosurfaces that intersect the tropopause. For complete details on the SMW analysis method and examples, please see Strong and Davis (2005, 2006a,b).

### c. Jet core frequency, speed, and pressure

For this research, a jet core is defined as a local speed maximum on the SMW greater than or equal to 25.7 m s⁻¹ (50 kt). Local wind speed maxima were identified in the 6-hourly SMW wind speed data along each meridian using finite differencing. The along-meridian analysis is motivated by the predominately zonal orientation of jet stream axes and our focus on the meridional motions of zonally oriented features that characterize the teleconnections examined here. The frequency of jet cores at a grid point on a particular meridian during a winter (\( \tilde{C} \)) is the number of jet core occurrences at the grid point divided by the total number of 6-hourly observations for the winter. Using winter-mean frequencies, changes in the position of a jet stream are deduced where we observe a synoptic-scale, zonally oriented band of jet core frequency decrease flanked by a comparable band of jet core frequency increase. Likewise, regions where jet streams split or merge are deduced where we observe three adjacent bands of alternating frequency change with an increase (decrease) in the middle indicating merging (splitting). The mean winter jet core pressure (\( \tilde{P}_{\text{core}} \)) or speed (\( \tilde{V} \)) at a grid point is the mean value of SMW pressure or speed when jet cores are present over the grid point during winter.

### d. Statistical methods

Principal components analysis (PCA) is used to identify the leading patterns of simultaneous anomalies that account for variability in \( \tilde{C} \) and \( \tilde{V} \). For each variable, the PCA is performed for the NH poleward of 20°N (“extratropics”). In each PCA, the variables are winter-mean \( \tilde{C} \) and \( \tilde{V} \) at each 2.5° grid point for 1958–2006. The data are standardized and weighted by the square root of the cosine of latitude to ensure proper representation in the correlation matrix, and singular value decomposition (SVD) is used to determine the eigenvalues and unit length eigenvectors. The SVD is performed on the correlation matrix rather than the variance–covariance matrix because the latter may be undesirably influenced by the tendency for the vertical spacing of isobaric surfaces to increase equatorward. The time series produced by projecting the weighted, standardized data onto the \( m \)th eigenvector is “principal component \( m \)” (PC \( m \)), and the contoured map of the \( m \)th eigenvector’s elements is the PC \( m \) “loading pattern.” Maxima and minima in the loading pattern will be referred to, respectively, as positive and negative “centers of action” (Barnston and Livezey 1987). Each PC is standardized for display by subtracting its mean and dividing by its standard deviation. The variance accounted for by PC \( m \) is the PC \( m \) eigenvalue normalized by the sum of the eigenvalues for the correlation matrix.

The difference between the mean of two groups of data is tested for significance using a two-sample \( t \) test. Correlations between time series are measured with least squares linear regression and significance is determined based on the effective degrees of freedom (df) adjusted to account for Pearson lag-1 autocorrelation in the residuals. Where correlation coefficients are mapped for the hemisphere, results have been tested for field significance at \( \alpha = 0.05 \) using the Monte Carlo method described in Livezey and Chen (1983). Field significant at \( \alpha = 0.05 \) indicates less than or equal to a 5% probability that the fraction of area with locally significant results occurred by chance given the effective spatial degrees of freedom in the field.

### 3. Results

#### a. Primary jet core patterns

In the NH extratropics (20°–90°N), the loading pattern for the leading mode of variability in jet core frequency (PC 1) indicates expansion and contraction of a spiral-shaped corridor (Fig. 1a). In the positive phase of PC 1, jet cores become more frequent in the band that originates over Central America, gradually approach the pole while spiraling approximately 400° counterclockwise around the hemisphere, and extend into western Europe (Fig. 1a, solid contours accentuated by bold arrow). During the negative phase, jet cores are more frequent in the adjacent band originating over central Africa and spiraling approximately 540° counterclockwise around the hemisphere into the central North Pacific (Fig. 1a, series of dashed contours shaded for emphasis). The bands of positive and negative loadings for PC 1 flank the winter jet’s spiral-like mean configuration mapped in Koch et al. (2006), showing that the PC 1 pattern represents oscillation about a spiral-like mean configuration.
The leading mode of extratropical jet core speed variability (\(\tilde{V} PC 1\)) is most strongly loaded in the Western Hemisphere and indicates a tendency for higher jet core speeds to migrate between the mid- to high and subtropical latitudes (Fig. 1b). There is fairly strong alignment between the \(\tilde{V} PC 1\) and \(\tilde{C} PC 1\) centers of action in the Western Hemisphere (cf. Figs. 1a,b), and the two score series are correlated with each other (Fig. 1c; statistics in Table 1), indicating that jet core speed changes are linked to meridional migration of the jet stream (particularly in the Western Hemisphere). Moreover, the \(\tilde{V} PC 1\) and \(\tilde{C} PC 1\) score series are each significantly correlated with the AOI (Fig. 1c; statistics in Table 1). As the AOI is a frequently referenced climate index and accounts for a substantial portion of the \(\tilde{V} PC 1\) and \(\tilde{C} PC 1\) variance, we now examine jet core speed and frequency variability as the function of the AOI.

b. The Arctic Oscillation

The AOI accounts for jet core frequency and speed variability most strongly over the Atlantic sector, where bands of \(\tilde{C}\) versus AOI correlation (Fig. 2a) and \(\tilde{V}\) versus AOI correlation (Fig. 2b) resemble the spiral pattern in the loading bands of \(\tilde{C} PC 1\) (section 3a). The AOI is based on surface pressure changes that correspond closely to like-signed changes in overlying geopotential height (Thompson and Wallace 1998). Differences in jet core speed between the AO positive and negative centers of action (Fig. 2b, + and − symbols) are physically consistent with the indicated height changes. Specifically, geopotential height decreases at the negative center and increases at the positive center.

![Fig. 1](image1.png)

**TABLE 1.** Correlation statistics related to score series of jet core frequency principal components (\(\tilde{C} PC 1\)–2), jet core speed principal components (\(\tilde{V} PC 1\)–2), the AOI, and the PNAI. The variable \(r\) is the Pearson correlation coefficient and all correlations are significant at \(\alpha = 0.05\).

<table>
<thead>
<tr>
<th>Variables</th>
<th>(r)</th>
<th>df</th>
<th>MSE</th>
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<tr>
<td>(\tilde{C} PC 1) vs (\tilde{V} PC 1)</td>
<td>0.77</td>
<td>48</td>
<td>0.41</td>
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<tr>
<td>(\tilde{C} PC 1) vs AOI</td>
<td>0.60</td>
<td>39</td>
<td>0.64</td>
</tr>
<tr>
<td>(\tilde{V} PC 1) vs AOI</td>
<td>0.71</td>
<td>48</td>
<td>0.49</td>
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<tr>
<td>(\tilde{C} PC 2) vs (\tilde{V} PC 2)</td>
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<td>34</td>
<td>0.66</td>
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<tr>
<td>(\tilde{C} PC 2) vs PNAI</td>
<td>0.78</td>
<td>25</td>
<td>0.40</td>
</tr>
<tr>
<td>(\tilde{V} PC 2) vs PNAI</td>
<td>0.57</td>
<td>48</td>
<td>0.68</td>
</tr>
</tbody>
</table>

The leading mode of extratropical jet core frequency (\(\tilde{C}\) for winters (December–February) 1958–2006 with contour interval 0.02 and negative values dashed (bold arrow and shading in negative regions accentuates interwoven spiral patterns). (b) Same as in (a) but for jet core speed (\(\tilde{V}\)). (c) Score series for \(\tilde{C} PC 1\), \(\tilde{V} PC 1\), and the AOI.
act cooperatively to strengthen the poleward-oriented geopotential height gradient ($\frac{\partial Z}{\partial y}$). Conversely, speeds equatorward of the AO positive center of action decrease because the geopotential height increases at the positive center of action and weakens $\frac{\partial Z}{\partial y}$. Near 60°N, the zone of increasing $\tilde{V}$ is associated with decreasing $\tilde{P}_{core}$ (Figs. 2c,d), whereas, between 30° and 45°N, the zone of decreasing $\tilde{V}$ is associated with increasing $\tilde{P}$ (Figs. 2c,d).

The relationships between jet cores and the AOI resemble jet core versus NAOI results over the Atlantic sector (not shown), reflecting the correlation of the AOI and NAOI ($r = 0.75$, df = 32, and mean square error (MSE) = 0.23 for the 1958–2006 mean winter series used here). Outside the Euro–Atlantic sector, the AOI is distinguished from the NAOI by accounting for $\tilde{V}$ variability west of the Canadian Rockies toward the Gulf of Alaska and in the tropical east Pacific near Baja (Fig. 2b). The changes in $\frac{\partial Z}{\partial y}$ caused by height changes at the AO’s Pacific center of action are consistent with the poleward speed increases shown in Fig. 2b, although the correlations are not as strong as those observed over the Atlantic.

The spatial pattern of $\tilde{P}_{core}$ variability accounted for

![Fig. 2](image-url)
by the AOI (Fig. 2c) reflects the characterization of the AO as an annular or zonally symmetric variability pattern (Thompson and Wallace 1998). Increases in the AOI are associated with jet core pressure increases over most of the subtropics and jet core pressure decreases over the mid- to high latitudes in an elliptical pattern with a long axis oriented from northeast North America across the pole toward East Asia (Fig. 2c). AO-related changes in zonally averaged jet core pressure are as large as 30 hPa for some locations (Fig. 2d). The tendency for a latitudinal offset between the jet core pressure and speed change maxima (e.g., Fig. 2d near 60°N) is related to the steep slope of the SMW through the jet core (Strong and Davis 2006b). The pattern of jet core pressure changes associated with the AO is similar to the pattern of pressure changes for the SMW (of which $\tilde{P}_\text{core}$ is a subset), and a detailed examination of the changes in tropospheric temperature underlying the AO-related vertical motion of the SMW is provided in Strong and Davis (2006a).

c. Secondary jet core patterns

The loading pattern for the second principal component of extratropical jet core frequency variability ($\tilde{C}$ PC 2, Fig. 3a) captures the contrast between split and merged jet streams over the Pacific and North America. During the negative phase of $\tilde{C}$ PC 2, split winter-mean jet streams are indicated over the east Pacific, while a merged winter-mean jet stream is present over North America (Fig. 3a, dashed contours). During the positive phase of $\tilde{C}$ PC 2, the reverse is indicated—a merged winter-mean jet stream over the east Pacific and split winter-mean jet streams over North America (Fig. 3a, solid contours). The positive $\tilde{C}$ PC 2 pattern implies ridge amplification over western North America (core motion following the arrows in Fig. 3a). The interpretation of the results in Fig. 3 as reflecting split versus merged jet streams is enabled by separate analyses of core frequency and jet core speed and is reinforced by the relationship between the depicted patterns and circulation shifts associated with the PNA, a linkage we examine more closely in section 3d.

Changes in jet core speed implied by the $\tilde{V}$ PC 2 loading pattern (Fig. 3b) are more hemispheric than the jet core frequency changes indicated by $\tilde{C}$ PC 2, but a relationship between the two patterns is suggested by

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**Fig. 3.** Same as in Fig. 1 but for the second PCs of (a) $\tilde{C}$ and (b) $\tilde{V}$ in the extratropics (20°–90°N). The arrows in (a) show motion of the circumpolar vortex associated with increases in the PNAI (Davis and Benkovic 1994). (c) Score series for $\tilde{C}$ PC 2 and $\tilde{V}$ PC 2 shown with the PNAI (correlation statistics in Table 1).
their well-aligned centers of action over the Pacific and North America (cf. Figs. 3a,b). Indeed, $\tilde{V}$ PC 2 is correlated with $\tilde{C}$ PC 2, and both are correlated with the PNAI (Fig. 3c; statistics in Table 1). The PNAI is a frequently referenced climate index, and it accounts for a substantial portion of the $\tilde{V}$ PC 2 and $\tilde{C}$ PC 2 variance. We will now examine jet core speed and frequency variability as a function of the PNAI.

d. Pacific–North American pattern

Changes in jet core frequency, which parallel changes in the PNAI, are most evident over North America and the Pacific near the international date line (Fig. 4a). The configuration and sign of the $\tilde{C}$ versus PNAI correlations over North America and the Pacific can be interpreted in the context of changes in the position of the circumpolar vortex (Fig. 4a, arrows). Increases in the PNAI are associated with 1) an equatorward shift of the circumpolar vortex from the Gulf of Alaska toward $40^\circ$N and 2) a poleward shift in the vortex over western North America from $40^\circ$ to $60^\circ$N (Davis and Benkovic 1994). The positive and negative dipoles of $\tilde{C}$ versus PNAI correlation joined by arrows in Fig. 4a also reflect the deepened Aleutian low/amplified western North American ridge configuration associated with increases in the PNAI (e.g., Rodionov and Assel 2001).

Increases in the PNAI account for increases in jet core speeds in a zonal band extending from east India and China across the Pacific to the Gulf of Mexico (Fig. 4b) and decreases in jet core speed in a zonal band extending from Siberia across the Gulf of Alaska to-

![Figure 4](image-url)
ward eastern North America (Fig. 4b). The $V_{\text{core}}$ changes over the Pacific and North America pair logically with the four PNA centers of action identified by Wallace and Gutzler (1981) (shown as + and − symbols in Fig. 4b). Since the jet stream level climatological mean geopotential height gradient is oriented poleward, increases in the PNAI indicate increases in geopotential height gradient for two regions: between the Pacific centers of action and equatorward of the negative center of action near the Gulf of Mexico. Likewise, increases in the PNAI indicate decreases in geopotential height gradient for two regions: between the North American centers of action and poleward of the negative center of action near the Gulf of Alaska.

Increases in the PNAI are associated with decreasing $P_{\text{core}}$ equatorward of the region over the Pacific where $V$ increases (Fig. 4c). Increasing $P_{\text{core}}$ (Fig. 4c) accompanies increases in the PNAI over the high-latitude regions that overlap, or sit just poleward of, the region of jet core slowing in Fig. 4b. In contrast, Fig. 4d highlights a zone between 30° and 37°N where jet core speeds and frequency are markedly increased, but jet core pressure remains relatively constant. The increase in jet core speed without jet core ascent during above-average PNAI is indicative of strong changes in underlying baroclinicity—changes driven by the opposing height changes that flank the zone of increasing speed (Fig. 4d, PNA positive and negative centers of action). For a detailed example of this behavior, see Fig. 1 in Strong and Davis (2006a).

4. Summary and discussion

In the extratropics of the NH, the leading pattern of variability in jet core frequency ($\bar{C}$ PC 1) is correlated with the AOI and indicates an oscillation about the spiral-shaped mean configuration of the winter jet stream. Increases in the AOI, or the NAOI, are accompanied by increases in jet core frequency and speed, and decreases in jet core pressure, in zonally oriented bands near 20° and 60°N (with opposite changes over intervening latitudes). Over the extratropical Pacific, neither the NAOI nor the AOI account for more than 4% of the variability in jet core frequency at any grid point. AO-related jet core pressure variability, however, is spatially expansive and the difference between above-average and below-average AOI winters exceeds 30 hPa for some locations. This result evidences the importance of using an analysis framework that moves vertically with the jet cores (e.g., the SMW) when evaluating jet core variability.

Under increasing AOI, jet core pressures increase over the Pacific and jet core speed increases over the Gulf of Alaska, but no evidence for AO-related weakening of Pacific jet cores was found here using the SMW and the NCEP–NCAR reanalysis. This contrasts with the finding in Ambaum et al. (2001) that, based on correlation between the AOI and 250-hPa zonal wind ($u_{250}$) in the European Centre for Medium-Range Weather Forecasts reanalysis, the Pacific jet weakens in response to increases in the AOI. Some of the negative $u_{250}$ versus AOI correlation identified by Ambaum et al. (2001) may be generated by descent of the jet stream away from 250 hPa during increasing AOI (positive $P_{\text{core}}$ versus AOI correlation shown here).

The second leading pattern of variability in jet core frequency and speed captures the tendency for a split (merged) winter-mean jet stream in the east Pacific to accompany a merged (split) winter-mean jet stream over North America. These patterns are correlated with the PNAI, and their positive phase reflects a deepened Aleutian trough and an enhanced ridge over western North America.

For the first and second principal components, we found alignment between the jet core frequency centers of action and jet core speed centers of action. This statistical finding implies a physical relationship between jet core speed and jet core frequency as they vary with teleconnection-related temperature and pressure changes: when jet cores become more (less) common over a region, they tend to be faster (slower) when they occur over that region. This implies, from the thermal wind relationship, that vertically integrated temperature gradients in the frontal zone underlying jet cores over a particular location tend to be larger as jet-supporting frontal zones occur more frequently over that location (winter-mean jet core frequency and speed are in fact positively correlated over most of the NH).

This work focused on jet core frequency, speed, and pressure variability in the extratropics. Variability in jet core frequency and speed in the tropics (equatorward of 20°N) is dominated by a pattern highly correlated with El Niño–Southern Oscillation, producing significant shifts in patterns of subtropical and extratropical jet core frequency, pressure, and speed to be explored in future publications.

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REFERENCES


