Earth Rotation as a Proxy for Interannual Variability in Atmospheric Circulation, 1860–Present

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ABSTRACT

Modern atmospheric and geodetic datasets have demonstrated that changes in the axial component of the atmosphere's angular momentum and in the rotation rate of the solid earth are closely coupled on time scales of up to several years. We therefore examine the feasibility of using a historical record of the earth's rotation as a proxy for year-to-year changes in the zonal wind field over the globe. The bulk of the earth rotation series acquired for this purpose is based on telescopic observations of the occultation of stars by the moon; semiannual values of changes in the length of day derived from these observations have acceptably small errors from about 1860 onwards. We filter these values to remove decade-scale fluctuations, which are driven primarily by non-atmospheric processes, and we examine the resulting proxy series to see if it contains a signal associated with one of the major modes of interannual variability in the atmosphere, namely that due to the El Niño/Southern Oscillation (ENSO). According to tests of statistical significance, such a signal is present in the historical earth rotation series, in that the day is typically longer during the year following an ENSO oceanic warm event than otherwise. We therefore proceed to consider other signals of interannual variability in the proxy series. In particular, we infer that noteworthy trends in atmospheric interannual variability have occurred over the last century; for example, the decade of the 1920s was marked by much larger year-to-year changes in the zonal circulation over the globe than that of the 1940s. Based on modern atmospheric data, we tentatively suggest that most of these circulation changes have resulted from anomalies in the region between 30°N and 30°S.

1. Introduction

Interest in the nature and causes of interannual variability and longer-term climate fluctuations has prompted meteorologists to turn to a variety of data sources to examine the behavior of the atmosphere over the last 100–150 years. Prior to the advent of routine upper-air soundings in the 1950s, these data were necessarily limited to observations of surface parameters, such as surface air temperature, for which useful records over land date back to the mid-19th century (Jones et al., 1986). A strong signal of interannual variability in the atmosphere involves the so-called Southern Oscillation, for which surface-based indices also date back to the 1800s (Wright, 1975). Measures of behavior in the ocean related to the Southern Oscillation, pertaining to the occurrence and strength of El Niño events, exist over this time period as well (Quinn et al., 1978). Despite their dependence solely on surface observations, such data records, when interpreted in light of our growing understanding of the physical basis for climate variability, can provide important insights into the past behavior of the upper-air circulation as well. Nevertheless, a more direct index of variations in upper-air winds prior to the rawinsonde era would also be desirable. The purpose of this paper is to describe one such index, namely the variable rotation rate of the earth, and to illustrate its potential as a proxy for global wind fluctuations since the middle of the last century.

The basis for considering earth rotation as a measure of the atmosphere's circulation derives from recent studies (e.g., Rosen and Salstein, 1983; Hide, 1984) that have demonstrated that changes in the relative angular momentum (M)[1] of the atmosphere about the earth's rotational axis are highly correlated with small, but measurable, changes in the length of day (Δlod) on time scales of up to a few years. This correspondence between the two quantities is illustrated in Fig. 1, in which daily values of M during 1980–83 are plotted against three-day mean values of Δlod derived by Morgan et al. (1985) for the same period from a combination of modern astrometric techniques. (The mean of each series and the effects of known solid body tides on Δlod have been removed to create these curves.) The M values are based on combining the global wind fields produced by the National Meteorological Center (NMC) for levels between 1000 and 100 mb with geostrophic wind estimates for levels between 100 and 1 mb computed by Hirota et al. (1983) from satellite-based stratospheric height analyses (Rosen and Salstein, 1985). The two scales along the ordinate are linearly

[1] $M = 2\pi a^3 g^{-1} \iint [u] \cos^2\phi \, d\phi \, dp$, where a is the radius of the earth, g the acceleration due to gravity, ϕ latitude, p pressure, and $[u]$ the zonal mean wind, positive in the eastward direction.

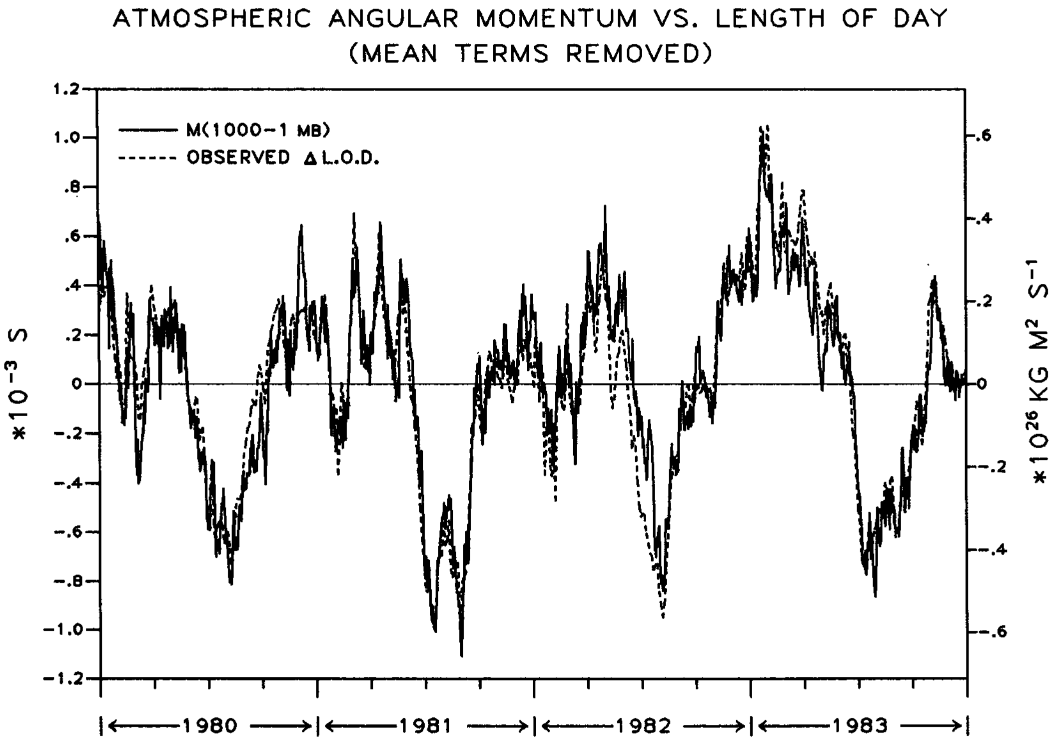

FIG. 1. Time series of daily values of the angular momentum of the entire atmosphere between 1000 and 1 mb (solid line; scale on right) and three-day means of observed changes in the length of day (dashed line; scale on left) for 1980–83. The mean value of each series during 1980–83 has been removed, as have solid body tidal terms from Δlod.

related through the equation $\Delta\text{lod (s)} = 1.68 \times 10^{-29} \Delta M$ (kg m^2 s^{-1}), which can be derived on the assumption that changes in M for the entire atmosphere are accompanied by equal, but opposite, changes in the angular momentum of the earth's crust and mantle, i.e., shell (Rosen and Salstein, 1983).

As Fig. 1 demonstrates, this assumption seems to be well justified on the time scales depicted, for although some discrepancies between the two series are present, by and large the agreement is very good. This is true not only for the seasonal cycle and shorter period fluctuations, such as those at 40–50 days (Langley et al., 1981), but also for interannual fluctuations, thereby suggesting the feasibility of utilizing historical records of Δlod as proxies for year-to-year changes in M. In section 2, we describe one such historical earth rotation series and its preparation for use as an atmospheric proxy.

2. A historical earth rotation series

To determine changes in earth rotation rate during the historical period before the development of atomic clocks in the 1950s, it is necessary to rely on observations of the position of the moon, either relative to that of the sun, as during eclipses, or relative to those of distant stars, as during their occultations at the lunar limb. The recorded locations and approximate times of such observations can be used to derive the departure of a cumulative measure of time based on the earth's rotation from a uniform time scale, determined by using an accurate, modern lunar ephemeris. The first derivative of this quantity yields the departure of the length of day from a standard day of 86 400 SI seconds.

Records of eclipses have been used to infer changes in the rotation rate of the earth back to Babylonian times (Stephenson and Morrison, 1984). Although useful for studying long-term changes in earth rotation, these ancient records are not precise enough for our present purposes, for which we must depend on observations of lunar occultations made since the invention of the telescope in the early 1600s. Even here, though, it was not until around 1860 that the number of observed occultations became sufficiently large that the error associated with the determination of Δlod is generally smaller than the interannual variation in length of day itself (McCarthy and Babcock, 1986).

In Fig. 2, we present a time series of semiannual values of the length of day since 1860 taken from the work by McCarthy and Babcock (1986). For the period 1860–1955, these values are based on the analyses of occultations by Morrison (1979a,b)[2] and, because of

[2] Values for 1940–55 were derived using an improved version of the lunation numbers contained in Fig. 5 of Morrison (1979a), which were supplied separately by him, rather than using the empirically fit numbers for this period listed in Table 1 of Morrison (1979b).

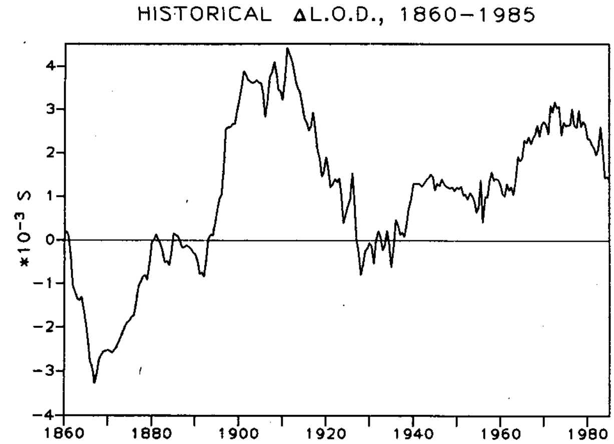

FIG. 2. Time series of semiannual values of the length of day during 1860–1985, taken from the work by McCarthy and Babcock (1986). A mean annual signal has been subtracted, as has a constant value of 86 400 SI seconds.

the smoothing and finite differencing schemes used to derive them, represent mean conditions over each half-year of the period,[3] which for convenience we will refer to as (Northern Hemisphere) winter and summer semesters. From 1956 onward, the McCarthy and Babcock values are based on more modern techniques used in conjunction with atomic clocks and have a finer temporal resolution than the pre-1956 values, but for consistency we include only semiannual values after 1956 in this study. In addition to the standard day, a mean annual signal has been subtracted to create the series in Fig. 2. Stephenson and Morrison (1984) present a series of once annual Δlod values for the same period shown in Fig. 2, but it has been more strongly smoothed than the McCarthy and Babcock series and so is less suited to our particular use.

As the figure reveals, fluctuations in the length of day since 1860 have occurred not only on interannual time scales but also on longer, decadal time scales. Indeed, it was the large change in the length of day between 1895 and 1910 that provided scientists in the 1920s with the first conclusive evidence that the rate of the earth's rotation does vary (Stephenson and Morrison, 1984). The amplitude of the so-called decade variations is in general so large that, for the most part, their origin cannot be meteorological; for example, the mean value of M during 1980–83 was $\sim 1.5 \times 10^{26}$ kg m^2 s^{-1}, which, if transferred entirely to the earth's shell (thereby bringing the atmosphere to a state of relative rest), would change lod by only 2.5 milliseconds, comparable to the decade variability shown in Fig. 2. The source for these variations has thus been generally ascribed to core–mantle coupling (Lambeck, 1980), although the nature of this coupling, be it viscous, topographic or electromagnetic, is a topical problem. To use the historical earth rotation series as a proxy for atmospheric behavior, therefore, it is necessary to filter out the decade variations, although this will perforce preclude the detection of any atmospheric fluctuations that may truly exist on this time scale. In passing, however, it is interesting to note some results (Lambeck and Cazenave, 1976) that indicate that the decade variations in lod are correlated with changes in various climate indices on this time scale, leading to the suggestion that both may have a common origin, such as volcanic activity. The hypothesis has also been offered (Courtillot et al., 1982) that the decade changes in lod originating in the earth's core are responsible for at least some of the changes in climate, by affecting the angular momentum of the atmosphere and hence its circulation. Such topics are beyond our current focus, however, which will be the higher frequency, interannual fluctuations in Fig. 2 and their implications for year-to-year differences in the atmosphere.

Because results such as those in Fig. 1 indicate that Δlod and ΔM are well correlated out to periods of four years or so, we wished to retain these shorter period fluctuations in Fig. 2 while damping the decade variations. Although a number of sophisticated filters could be designed for this purpose, we found the following five-point digital high-pass filter sufficient for our present analysis:

$$\hat{f}(x_0) = f(x_0) - \frac{1}{12}[f(x_{-2}) + 3f(x_{-1}) + 4f(x_0) + 3f(x_1) + f(x_2)],$$

where $f(x_i)$ represents the original value of Δlod in year x_i and $\hat{f}$ the filtered value. Winter and summer values of Δlod were filtered separately.

The response function for our filter is shown in Fig. 3, and the filtered series for both seasons are displayed in the curves of Fig. 4. It is clear from Fig. 4 that our filter has succeeded in damping the longer-period fluctuations in the original Δlod series while retaining the interannual variations of interest to us here. In the remainder of this paper, analyses using the historical Δlod series will be based entirely on the filtered values in Fig. 4.

3. Relationship between ENSO events and Δlod

A significant mode of interannual variability in the atmosphere's circulation is related to the ENSO phenomenon. Recent studies have demonstrated that the anomalous winds in the tropics and subtropics during an ENSO event can produce a marked signal in earth rotation time series. Thus, Rosen et al. (1984) found

[3] It is important to note, however, that prior to 1940 only data centered on northern winter are available from the works of Morrison; the midyear values of McCarthy and Babcock are, in essence, interpolations and do not represent independent information prior to 1940.

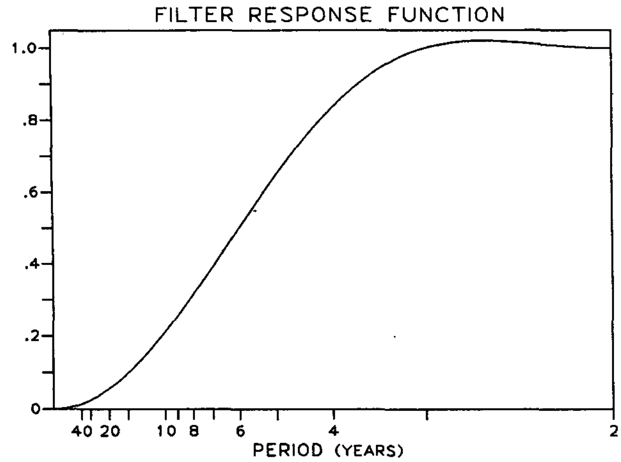

FIG. 3. Response of the high-pass filter used in our analysis, in terms of the ratio of the amplitude of fluctuations after filtering to that before, as a function of their period for all periods greater than two years. The abscissa is linear with respect to frequency.

that the unusually strong El Niño of 1982–83 was associated with anomalously high values of M and Δlod in January 1983. In addition, Stefanick (1982) and Chao (1984) documented a correlation between fluctuations in indices of the Southern Oscillation and interannual changes in earth rotation since the 1950s. It therefore seems reasonable to expect that a relationship between ENSO events and Δlod existed overall in earlier times as well. Indeed, discerning such a relationship in our historical earth rotation series would enhance our confidence in it as an atmospheric proxy.

Some uncertainty exists regarding the time lag involved between an ENSO oceanic warming event and the putative response in global atmospheric momentum. Teleconnections between the tropics and Northern Hemisphere extratropics are strongest in the northern winter following a warm event, during the so-called mature phase of the ENSO (Rasmusson and Carpenter, 1982). In addition, though, it appears on the basis of 200 mb wind anomalies between 48°N and 48°S composited by Arkin (1982; personal communication, 1985) for ENSO episodes since 1968 that positive anomalies in M can be expected not only during the northern winter but also as early as the summer following a warm event. Although, as noted earlier, much of the summer Δlod series is not independent of the winter Δlod values, we have examined both seasons for a relationship between ENSO events and anomalous values of Δlod in order to gain some further insight into the time scales linking the two.

Because we believe that any response in atmospheric momentum to ENSO would be nonlinear in character (i.e., dependent on ENSO achieving a certain threshold of intensity), our search for a relationship between the two involves compositing the Δlod values according to the existence or nonexistence of a well-defined ENSO warm event. A listing of 28 such warm events back to 1864 is given by van Loon and Shea (1985), based on criteria involving sea surface temperature in the equatorial Pacific, sea level pressure at a number of key stations, and rainfall at equatorial Pacific stations (van Loon, 1984). The Δlod values for the northern winters at the end of each of these 28 years have been marked by dots in the upper curve of Fig. 4; the Δlod values for the northern summers of the warm event years are similarly identified in the lower curve. (It is interesting to note in the winter series that the two largest values of Δlod are associated with the very strong ENSO events of 1925 and 1982.) The mean and standard deviation of these 28 winter and 28 summer Δlod values are given in Table 1, along with the same quantities computed for the winters and summers corresponding to the other 93, nonwarm event years during 1864–1984.

According to Fig. 4 and Table 1, the length of day is typically greater during the northern summer and winter following an ENSO oceanic warm event than otherwise. Values of Student's t computed for each season, included in the table, indicate that the difference in the mean Δlod between ENSO and non-ENSO years is statistically significant at the 95% level of confidence (Panofsky and Brier, 1958). The Student's t values become even larger (2.91 for northern winter, 2.38 for northern summer) if the years of 1923, 1932 and 1963 (indicated by open dots in Fig. 4) are eliminated from the warm event list of van Loon and Shea on the grounds that these three contained, at best, very weak manifestations of the ENSO phenomenon (van Loon, 1984).

Hence, our results demonstrate that a statistically significant ENSO signal exists in the historical record of the length of day dating back to the 1860s. Days are markedly longer, on average, during the summer and winter following a warm event than during these seasons of other years. The presence of this signal of interannual atmospheric variability in the historical Δlod data supports the thesis that these data provide a useful proxy index for global wind fluctuations. We exploit this result in section 4 by using the Δlod data to study trends in the variability of the global circulation during the last 125 years, which, unlike the case for ENSO events, cannot otherwise be determined well from other proxy information.

4. Trends in interannual atmospheric variability

As Boer and Higuchi (1980) have discussed, studies of climate change need to be concerned with trends not only in first-order quantities like temperature but also in higher-order quantities such as measures of temporal variance. In this latter regard, the filtered record of historical Δlod values can be used to infer long-term changes in the interannual variability of global M. One attempt at doing so is displayed in Fig. 5, which shows the standard deviation of the filtered Δlod series

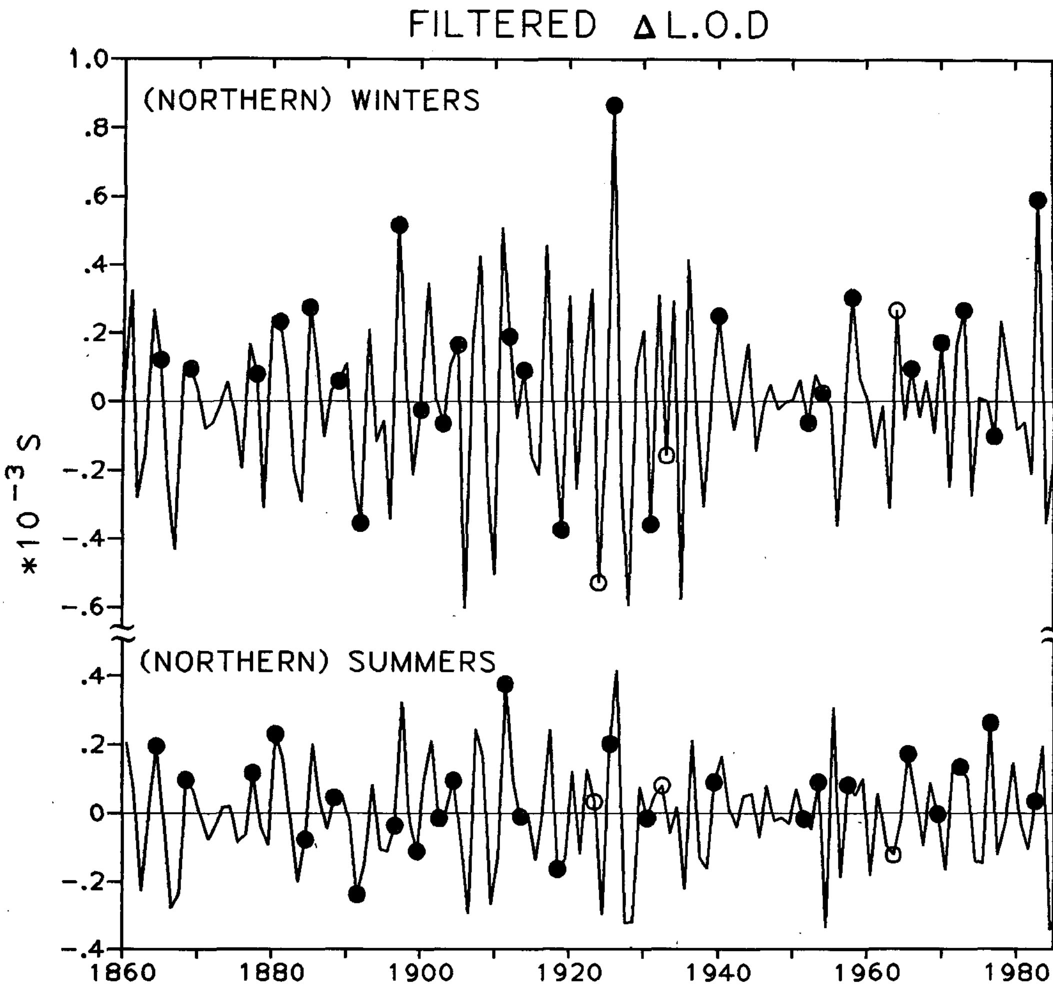

FIG. 4. Time series of high-pass filtered values of the length of day for northern winters during 1860–1985 (upper curve), and for northern summers during 1860–1984 (lower curve). Solid dots mark those winters that fall at the end of, or those summers that fall during, a year containing a well-defined ENSO warm event, according to the list of van Loon and Shea (1985); open dots indicate seasons associated with the three very weak ENSO events according to van Loon (1984).

for northern winter during the ten-year period centered about each year from 1865–1980. The equivalent scale for deviations in atmospheric angular momentum, based on the relationship between Δlod and ΔM given in section 1, is plotted along the right-hand ordinate. In essence, therefore, the figure provides a picture of decade-scale variations in the interannual variability of northern winter global M.

Although no secular linear trend in the running ten-year standard deviation of M is apparent, there are large fluctuations in this statistic during the study period. Most notable is the contrast between the high variability during the 1920s and the low variability of the 1940s to early 1950s. The relative quiescence of northern winters during this latter period was already evident in Fig. 4, of course. The difference in interannual variability between the 1920s and the 1940s cannot be attributed entirely to the difference in the frequency of ENSO events during the two decades, for although the 1940s were generally lacking in these events, four of the five ENSO episodes between 1918

TABLE 1. Mean and standard deviation of Δlod (in 10^{-3} s) associated with years having ENSO events and with all other years during 1864–1984. These statistics, along with the Student's t value found in comparing the difference between the means of the two groups, are presented separately for northern winter and summer seasons. A value of $t \geq 1.98$ implies significance at the 95% level.

	Northern winter		Northern summer.	
	ENSO	non-ENSO	ENSO	non-ENSO
Mean	0.095	−0.033	0.055	−0.019
Standard deviation	0.30	0.23	0.13	0.15
N	28	93	28	93
	$t = 2.40$		$t = 2.28$	

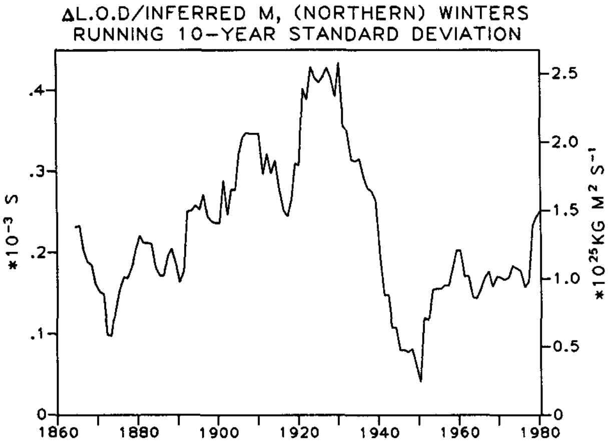

FIG. 5. Standard deviation of the filtered values of length of day for northern winters during the ten-year period centered about each year from 1865–1980. Equivalent scale for the running ten-year standard deviation in atmospheric angular momentum is plotted along the right-hand ordinate.

and 1932 were associated not with positive winter anomalies in Δlod, as might be expected from our results in section 3, but rather with negative anomalies (see Fig. 4). Indeed, the 1918–32 period in Fig. 4 is notable for its overall negative contribution to the relationship between ENSO and the mean northern winter Δlod tested in Table 1, which, however, is due in part to the inclusion of the weak events of 1923 and 1932. In any case, it is evident from Fig. 4 that the especially high variability depicted in Fig. 5 for northern winters during this period must be associated with some behavior beyond that due to ENSO events.

A possible explanation for the contrast in the variability of atmospheric momentum between the 1920s and 1940s may lie in the very different nature of the temperature regimes that were prevalent during these decades. According to the data provided by Jones et al. (1986) for the Northern Hemisphere, the period around the 1920s was marked by a rapid rise in surface temperature that then peaked and leveled off during the 1940s before undergoing a general downward trend. The connection, if any, between such temporal trends in hemispheric surface temperature and atmospheric circulation can only be a matter of speculation for the present, however. In addition, these temperature indices are dominated by records from land-based stations and so are not necessarily representative of truly hemispheric or global trends (Barnett, 1984). Because of these difficulties, we will not pursue this subject further here.

Further insight into the mechanisms responsible for the trends we have deduced in global M variability would be gained by understanding the regional sources of these trends, but to do so it is necessary to turn once again to contemporary datasets. Here, we use the daily NMC zonal wind fields mentioned in the Introduction to compute the angular momentum of the atmosphere within each of 46 equal-area latitude belts (see Rosen and Salstein, 1983, for the definition of these belts) for the northern winter season, 1 December–28 February, for the ten winters from 1976/77 through 1985/86. To obtain measures of interannual variability during this recent period, these belt and global M values were averaged within each individual winter season, and the variance among the ten seasonal values for each belt was computed. These values are plotted in Fig. 6, along with the 46 covariances between the belt and global M seasonal averages, normalized by the (interannual) variance in global M. This latter curve therefore depicts the fractional contribution made by each zonal belt to the total variance in global M on interannual time scales for the northern winter seasons from December 1976 through February 1986.

It would appear from the curves in Fig. 6 that although interannual variability in atmospheric momentum occurs over most latitudes to a comparable extent, the fluctuations between 30°N and 30°S are mainly responsible for the interannual changes in global M. *If we make the assumption that this relationship was as true in earlier decades as it is for the 1976–86 decade,* then we can estimate past tropical

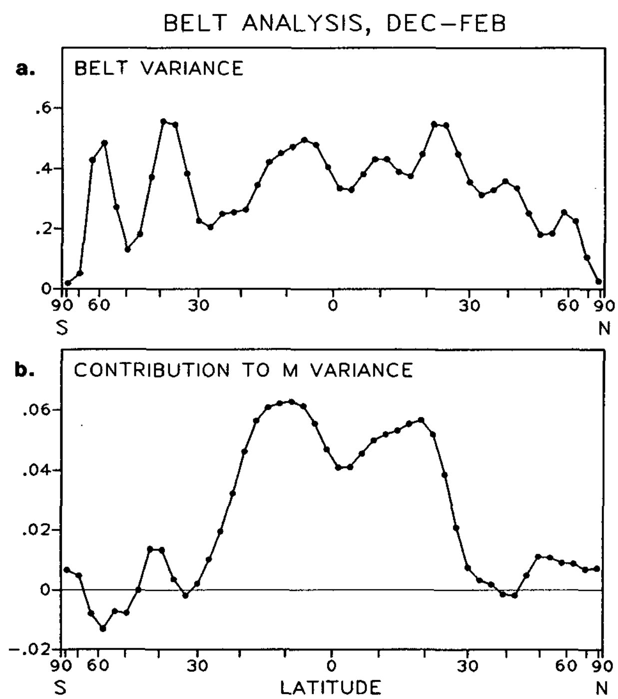

FIG. 6. (a) The variance among ten northern winter values of the angular momentum of the atmosphere in each of 46 equal-area latitude belts defined by Rosen and Salstein (1983), based on NMC analyses for December 1976–February 1986. Units are 10^{48} (kg m^2 s^{-1})^2. (b) Fractional amount of the interannual variance in global atmospheric angular momentum during the ten northern winters explained by fluctuations in each belt.

mean zonal wind variability from the variability in global M inferred from the proxy Δlod index (Fig. 5). To do this, we consider the profile in Fig. 6b to be uniform across the tropical band from 30°N to 30°S and zero elsewhere. Then we use the relationship in footnote 1 and the proportionality between ΔM and Δlod to obtain $\Delta u_{\rm tropics}$ (m s^{-1}) = 3.8×10^3 Δlod (s), where $\Delta u_{\rm tropics}$ and Δlod denote interannual standard deviations. Thus, for example, the difference in global Δlod variability between the 1920s and 1940s (0.3×10^{-3} s) translates to a difference between these two decades in the interannual standard deviation of the zonal wind $[u]$ of about 1.1 m s^{-1} throughout the tropics; i.e., the standard deviation among annual northern winter values of $[u]$ in this region was 1.1 m s^{-1} larger during the 1920s than during the 1940s. To place this value in perspective, it is worth noting that the standard deviation of $[u]$ averaged over 30°N–30°S and 1000–100 mb among the 10 December–February values for the recent 1976–86 period based on the NMC data is 0.74 m s^{-1}.

The inferences drawn above about the regional distribution of the interannual variability in M and $[u]$ must be regarded as quite tentative, of course, in light of the restrictive assumptions required. One critical assumption concerns the representativeness of the covariance curve in Fig. 6, derived from data of recent years, for behavior in the past. Updating this calculation as newer data become available would increase our confidence in applying it to earlier decades, but in any case it should be emphasized that only that portion of the interannual variability in latitude belts that is actually correlated with fluctuations in global M can be inferred through this approach. Nevertheless, we appear to have identified a sizable tropical signal in the trend of interannual variability that warrants further consideration and, if possible, corroboration by other datasets.

5. Concluding remarks

Because of the excellent agreement found between modern time series of atmospheric angular momentum and the length of day, we have explored the possibility of using historical records of Δlod, suitably filtered to remove obvious nonatmospheric signals, to infer past trends in atmospheric interannual variability. Unfortunately, the need to filter the effects of core–mantle coupling from the historical earth rotation series has precluded us from studying any variations among decadal means of atmospheric angular momentum that may really exist. If independent estimates of the magnitude of core-induced decade variations in Δlod were ever to become sufficiently precise, however, then it might be possible to separate out this atmospheric signal in our proxy series. For now, though, the proxy data can be used to infer only higher frequency changes in the atmosphere; nevertheless, as we have shown, this application can be quite fruitful in itself.

The success of a study such as this, of course, is highly dependent on the quality of the historical Δlod series, including its homogeneity. Morrison (1979a) describes the various sources of error in this series associated with the nonuniform distribution of occultation observations over time, uncertainties in the methods of timing occultations, and errors in the particular lunar ephemeris (j = 2) used to reduce the occultation data. Errors in this ephemeris may give rise to periodic errors in Δlod on time scales of three to four years (Morrison, 1979b), which is of particular worry to us here, although the amplitude of this periodic term is uncertain. It is partly because of the possible importance of this error source that the Δlod series of Stephenson and Morrison (1984), referred to in section 2, was so strongly smoothed. The less-smoothed McCarthy and Babcock values that we have used here, however, seem to contain real information on interannual time scales, as evidenced by their statistically significant correlation with ENSO event data.

Some of the uncertainty in the historical Δlod series could be reduced by applying a more accurate lunar ephemeris (such as "DE200") to the occultation data, and plans exist to do so (Morrison, personal communication, 1985). Although such a reanalysis may yield a more accurate resolution of the historical series into independent quarterly, or even monthly, values and extend it more reliably back to the early 1800s, ultimately the quality of the series is limited by that of the basic occultation data. Some assessment of the quality of these occultation data could be made by comparing Δlod values derived from them for the post-1955 period with those derived from more modern techniques. In any case, reanalyzing the raw occultation observations would be a major task.

The results presented here ought to offer encouragement to those who would seek to improve the historical Δlod series, for even in its current state the series seems to serve as a useful proxy for atmospheric behavior. In combination with other proxy data available for past centuries, such as those from tree rings (Lough and Fritts, 1985) or tropical ice cores (Thompson et al., 1985), historical earth rotation data can contribute to a more complete picture of the full range of variability inherent in the climate system.

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