A quasi-biennial signal in western US hydroclimate and its global teleconnections

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Abstract On interannual time scales, quasi-cyclic activity in the troposphere is generally thought to arise from chaotic feedback processes, and therefore to be unpredictable in nature. Here, evidence is presented for periodic ‘clockwork’ climate behavior, seen as a stable quasi-biennial (QB) oscillation in Western US winter precipitation, estimated at a period of $2.16 \pm \sim 0.05$ years. In the Western US, the QB precipitation mode has an influence comparable to that of ENSO, with strongest effects observed at central latitudes of the region. The oscillation displays systematic phase-shifting with respect to the annual cycle, signifying a stable, nonseasonal and strongly periodic QB mechanism. The cycle in precipitation results proximally from a meridional standing pressure wave over the North American Pacific coast. Analysis of monthly pressure data confirms the nonseasonal, periodic character of the oscillation, which appears to have maintained phase from the mid-20th century to recent years. The regional QB pressure cycle is traced to a quasi-synchronous pulsation of meridional pressure waves in the extratropics of both hemispheres, termed a Biennial Annular Mode Oscillation (BAMO). The BAMO oscillates in quadrature with a weaker ~ 26 -month

signal in the tropical Southern Oscillation. Due to a shorter period, both oscillations operate independently of the stratospheric Quasi-biennial Oscillation (QBO).

Keywords Quasi-biennial oscillations · Western US winter precipitation · Annular mode · ENSO · TBO · California

1 Introduction

The Western United States (Fig. 1a, b) is a largely semi-arid region with ecological and economic vulnerability to seasonal hydroclimate anomalies. The region experiences considerable interannual variability in water supplies, largely due to changes in the behavior of North Pacific winter cyclones, which provide the bulk of the annual precipitation. Winter precipitation patterns are known to be sensitive to the state of the El Niño–Southern Oscillation (ENSO) (Redmond and Koch 1991), which continually reorganizes the ocean–atmosphere circulation of the tropical Pacific over irregular periods of ~ 2 –7 years. ENSO effects on Western US hydroclimate are thought to be driven by tropical convective anomalies that alter sea-level pressure (SLP) patterns (Fig. 1c) and the westerly flow in the midlatitude troposphere, including winter storm tracks over the NE Pacific. ENSO tends to produce north–south shifts in the Western US winter precipitation distribution, as seen through correlations between local precipitation and the winter Southern Oscillation Index (SOI) (Fig. 1d). ENSO effects are greatest at far southern latitudes, where El Niño conditions typically correspond to positive precipitation anomalies, and La Niña conditions to meteorological drought. A weaker opposite relationship prevails in the far north. This simple quasi-linear ENSO teleconnection

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Fig. 1 **a** Western US study region **b** 373 precipitation stations used in the study. *Filled markers* indicate long-term stations with at least 90% complete coverage from 1871 to 1900. 2° grid boxes for station averaging are bounded with *bold lines*. **c** Winter SLP correlations with the SOI. Contours indicate 95% significance. **d** Gridded winter precipitation correlations with the SOI. *Shading* indicates 95% significance. Negative contours *dashed*

forms the primary basis for understanding and forecasting Western US hydroclimate anomalies.

Less clear, however, are large-scale controls on precipitation at central latitudes of the Western US and over the region as a whole. Hydroclimate variability at intermediate latitudes displays only weak or irregular connections to ENSO (Schonher and Nicholson 1989; Haston and Michaelsen 1994; Mitchell and Blier 1997; Andrews et al. 2004); and regionwide precipitation, unlike the north–south dipole, fails to display strong connections to Pacific-wide circulation patterns (Dettinger et al. 1998). The systematic variability of regionwide precipitation remains unclear, and important unanswered questions remain over its behavior and predictability.

An element of regularity is suggested, however, by the presence of a significant ‘quasi-biennial’ (QB) oscillation in the range of 2.1–2.2 years, a cycle noted but unexplained in studies of regional and local precipitation variability (Granger 1977; McGuirk 1982; Dettinger et al. 1998). Significant periodicities in this range also appear in precipitation-sensitive tree-ring records from the region dating back 500–1,000 years (Bryson and Dutton 1961; Diaz and Pulwarty 1994; Gray et al. 2004), suggesting a long-lived QB phenomenon. This oscillation, identified here in a regionwide precipitation index dating to 1871, forms the basis of the present investigation.

In 1884, a specific QB mode of variability was first noted in US weather records by meteorologist H.H. Clayton, who documented a coherent ~ 25 -month cycle in

pressure, temperature, and precipitation among widely separated stations (Clayton 1884). In more recent decades, periodicities in the range of 25–26 months (2.1–2.2 years) have also been identified in a NE–SW contrast in US temperature anomalies (Walsh and Mostek 1980; Rasmusson et al. 1981; Dettinger et al. 1995). The narrow range of reported QB frequencies across different time intervals and climate variables suggests the presence of a persistent, frequency-specific atmospheric circulation over the United States. Observed QB periods are generally shorter than the prominent 2.3–2.4 year (28–29 month) Quasi-biennial Oscillation (QBO) of equatorial stratospheric winds, known since the early 1960s (Baldwin et al. 2001). The QBO dominates the circulation of the tropical lower stratosphere, but is typically found to operate independently of QB cycles identified in surface climate, suggesting that the stratosphere and troposphere may possess essentially different QB modes (Angell et al. 1969; Angell and Korshover 1974; Trenberth 1975, 1980; Barnett 1991; Xu 1992; Mann and Park 1996; Baldwin et al. 2001).

QB modes have also been observed in several components of tropical surface climate, including ENSO (Rasmusson et al. 1990; Barnett 1991; Ropelewski et al. 1992; Jiang et al. 1995; Zhang et al. 1998), Indian monsoon rainfall (Kane 1995; Fasullo 2004), a zonal dipole of Indian Ocean sea surface temperature (SST) and SLP (Saji et al. 1999), and the Indo-Pacific trade winds (Barnett 1983). These regional QB phenomena are connected by eastward-propagating SST and SLP waves involving the entire

global tropical belt (White and Allan 2001). Monsoon-related QB variations are often discussed as the Tropospheric Biennial Oscillation (TBO) (Meehl 1997; Meehl and Arblaster 2002), characterized by year-to-year anomaly reversals, the defining signature of an annually-sampled biennial cycle. Model representations of the TBO have invoked seasonally-dependent air–sea interactions in the equatorial Indian Ocean and the far western Pacific to account for interannual flip-flops in regional SST, zonal wind, SLP, and precipitation anomalies associated with monsoon intensity (Brier 1978; Nicholls 1978; Clarke et al. 1998; Meehl et al. 2003; Li et al. 2006). The eastward propagation of QB climate anomalies from the Indian Ocean to the western Pacific links the TBO with QB variations in ENSO (Yasunari 1985; Rasmusson et al. 1990; Barnett 1991; Ropelewski et al. 1992). The QB activity in the tropics is considerably masked by stronger climate variations over longer interannual time scales,

leaving remaining uncertainties over its basic nature and origins. Because QB variability is well-established in both the tropics and the Western US, connections between these regions are a topic of emphasis here.

Results from a number of studies suggest that QB variations in the troposphere are global in scale. In the early 1960s more than twenty separate observations of similar QB cycles around the world were presented as evidence for a 2.2-year global ‘biennial pulse’ in the lower atmosphere (Landsberg 1962; Landsberg et al. 1963). Since then, at least thirty additional studies have noted climate oscillations in the range of 2.1–2.2 years in wide-ranging areas of the globe (Tables 1, 2). Significant periodicities in this narrow span also appear in hemispheric- to global-scale patterns of temperature and pressure (Mann and Park 1996), temperature (Mann and Park 1994) and precipitation (Lau and Sheu 1988), however the essential mechanisms remain unresolved.

Table 1 Survey of tropospheric quasi-biennial oscillations over the range of 2.1–2.2 years/25–26 months since Landsberg et al. (1963)

Region	Variable	Period	References
California	Precipitation	25–26 months	Granger (1977)
N. American West Coast	Precipitation	2.1 years	McGuirk (1982)
Western US	Precipitation	2.2 years	Dettinger et al. (1998)
United States	Temperature	2.2 years	Dettinger et al. (1995)
United States	Temperature, precipitation	2.1 years	Walsh and Mostek (1980)
United States	Temperature	25.6 months	Rasmusson et al. (1981)
Atlantic Ocean	SST, SLP	2.2 years	Tourre et al. (1999)
N. Atlantic	SST, zonal wind, SLP	2.2 years	Deser and Blackmon (1993)
N. Atlantic	SST	25 months	Moron et al. (1998)
N. Atlantic	SLP	2.2 years	Kelly (1977)
N. Atlantic, N. Pacific	SLP	25–30 months	Angell et al. (1969); Angell and Korshover (1974)
NH midlatitudes	Surface pressure	24–30 months	Shapiro (1964)
ENSO-Arctic-N. Atlantic	SLP, sea ice	2.2 years	Jevrejeva et al. (2004)
Arctic	SLP, sea ice	2.1 years	Venegas and Mysak (2000)
East Africa	Precipitation	2.1–2.2 years	Rodhe and Virji (1976)
India (subregions)	Monsoon rainfall	2.0–2.3 years	Kane (1995)
Pacific	SST	26 months	Zhang et al. (1998)
NW Pacific	Cyclone, typhoon freq.	25.6 months	Chan (1985)
E. Equatorial Pacific	SST	25 months	Moron et al. (1998)
Indian Ocean	SST, SLP	2.2 years	Tourre and White (2003)
Tropical Indian -Pacific	Winds	26 months	Barnett (1983)
Australia/New Zealand	SLP, SST	25.8 months	Trenberth (1975, 1980)
Patagonia	Precipitation, SLP	2.0–2.2 years	Villalba et al. (1998)
S. America	Stream flow	2.2 years	Robertson and Mechoso (1998)
Northern Hemisphere	Temperature, SLP	2.1–2.2 years	Mann and Park (1996)
Global tropics	Temperature, SLP	2.2 years	White and Tourre (2003)
Global tropics	Temperature, SLP	2.0–2.4 years	White and Allan (2001)
Globe	Temperature	2.2 years	Mann and Park (1994)
Globe	Precipitation	26 months	Lau and Sheu (1988)

Table 2 Quasi-biennial oscillations over the range of 2.1–2.2 years/25–26 months in paleoclimate proxies since Landsberg et al. (1963)

Region	Variable, time interval	Period (years)	References
California	Tree ring widths (Precip), 800–1988	2.2	Diaz and Pulwarty (1994)
Wyoming	Tree ring widths (Precip), 1260–1998	2.1	Gray et al. (2004)
Nile River	Flood intensity, 629–1520	2.2	Diaz and Pulwarty (1994)
Nile River	High, low water years 622–1922	2.2	Kondrashov et al. (2005)
Queelccaya, Peru	Ice sheet $\delta^{18}\text{O}$, 744–1984	2.2	Diaz and Pulwarty (1994)
Patagonia	Tree ring widths (Precip, SLP), 1746–1984	2.2	Villalba et al. (1998)
NE USA	Lake varve thickness, 13–15 kya	2.1	Rittenour et al. (2000)

The purpose of this study is to examine the nature and origins of a significant QB cycle identified in a regional index of Western US winter precipitation. The oscillation is first identified by power spectrum analysis of a 137-year winter precipitation series and confirmed in century-long river flow records. The temporal signature of the QB oscillation is shown in detail over the past ~ 60 years, including a persistent interaction with the annual cycle that suggests the presence of a steady quasi-sinusoidal climate signal. The spatial signatures of this signal are next illustrated through analysis of its correlations with winter tropospheric pressure and SST over the globe. The analysis is then extended to the monthly time scale in order to inspect the nonseasonal structure of the QB cycle at lags during and prior to its appearance in the Western US. These findings lead to a discussion of the possible origins of the QB oscillation in the Western US, the Pacific, and worldwide.

2 Data and methods

The study is based primarily on an index of Western US winter precipitation computed from station data for November–April winters ending in years from 1871 to 2007. This index, denoted P , was compiled from 373 Western US monthly station records included in the US Historical Climate Network (USHCN). The index includes all USHCN stations from the Rocky Mountain states to the Pacific coast, in addition to El Paso, TX, included for regional completeness. Data through 2006 were obtained directly from USHCN update NDP-019 (Williams et al. 2007), available from the Carbon Dioxide Information and Analysis Center (http://cdiac.ornl.gov/epubs/ndp/ushcn/usa_monthly.html), and augmented through 2007 with updated records from the same stations in the National Weather Service Cooperative Network, available from the National Climatic Data Center. (<http://www.ncdc.noaa.gov/oa/ncdc.html>).

To construct the index, monthly station precipitation values were first converted to percent anomalies relative to 1971–2000 monthly means. Station anomalies were then

averaged over $2^\circ \times 2^\circ$ grid boxes covering Western US land areas from 125°W to 105°W and 49°N to 31°N (Fig. 1b). Grid box anomalies were computed by averaging station anomalies within a latitude–longitude circle circumscribing each grid box, contingent on the availability of five stations. Grid box circles not meeting this criterion were incrementally expanded in radius by 1° until five stations were encompassed. Monthly grid box anomalies were then re-standardized over the 1971–2000 period and weighted with the high-resolution PRISM (Parameter-elevation Regressions on Independent Slopes Model) climatology (Daly et al. 1994) <http://www.prism.oregonstate.edu/>. PRISM-based weightings include both mean precipitation depth and land area (to properly weight grid boxes that extend beyond Western US boundaries). These procedures yielded volumetric monthly precipitation estimates at the grid box level. Western US monthly precipitation depths were then computed in units of cm, and winter sums were calculated for each November–April season. A file containing grid box and regional winter precipitation totals is included in the Electronic Supplementary Material as Datafile 1. Individual winters are identified herein by the year of termination (e.g. the 2006–2007 winter is discussed as 2007). For analysis of global teleconnections, gridded monthly atmospheric data were obtained from the NCEP–NCAR Reanalysis (Kalnay et al. 1996) for the 715-month period from January, 1949 to July, 2008. Monthly SST data were obtained from the NOAA Extended Reconstructed SST data set, version 3b (Smith et al. 2008). The stratospheric QBO was analyzed using wind data from Barbara Naujokat at Freie Universität Berlin <http://www.geo.fu-berlin.de/met/ag/strat/produkte/qbo/qbo.dat>. Century-long records of inflows to the Sacramento and San Joaquin Rivers of northern and central California were obtained from the California Department of Water Resources <http://cdec.water.ca.gov/cgi-progs/iodir/wsihist>. Monthly values of the SOI for the 1951–2008 period were obtained from the Climate Prediction Center <http://www.cpc.noaa.gov/data/indices/soi>.

The QB cycle in precipitation was first identified by multi-taper method (MTM) power spectrum analysis and

signal reconstruction. The atmospheric circulation signatures of the QB cycle were investigated with global gridded Z500 and SLP data, which reflect pressure variability at mid-tropospheric ($\sim 5\text{--}6\text{ km}$) and near-surface elevations. Spatial teleconnection patterns were determined by mapping pressure and SST correlations with the QB precipitation cycle. Statistical significance was determined with Monte Carlo techniques, discussed as the results are presented.

3 Western US winter precipitation

3.1 Overall variability

The mean and standard deviation of the 1871–2007 winter precipitation series (Fig. 2a) were estimated at 32.8 and 5.5 cm, respectively. The frequency distribution is indistinguishable from normal, based on a Kolmogorov–Smirnov test, and the time series displays insignificant lag-1 autocorrelation ($r = -0.13$, $p > 0.10$) and a negligible trend. The greatest positive anomaly (+47%) occurred in 1890, while the record minimum occurred in 1977 (−58%). Local precipitation correlations with the regional index are strongest ($r > 0.80$) over California, Oregon and Nevada, where ENSO correlations are generally weak. Accordingly, an insignificant ($p > 0.10$) degree of correlation ($r = -0.14$) was obtained with the winter SOI.

Regional winter precipitation is strongly negatively correlated with atmospheric pressure over the NW corner of the Western US. The Z500 correlation (Fig. 2b) is strongest at 45°N , 130°W ($r = -0.74$), while the SLP correlation (Fig. 2c) reaches $r = -0.82$ at 45°N , 122.5°W . Interannual variability is thus directly linked to changes in the tropospheric circulation over the NE Pacific. Anomalous troughing in this area leads to enhanced regional precipitation totals, as frontal storms rotate cyclonically around the SE sector of low-pressure centers. Ridging in the same area tends to deflect storms to the north of the Western US.

Significant positive SLP correlations are also seen over the equatorial Indian Ocean and western Pacific, indicating a degree of connection between tropical circulation and Western US winter precipitation. Weak correlations in the eastern tropical Pacific account for the poor relationship to the SOI. Scattered areas of significant SLP correlations are also seen over the North Atlantic, Asia, and the Southern Hemisphere (SH). Significant SST correlations (Fig. 2d) occur primarily over the North Pacific and North Atlantic, and evidence of a clear connection to large-scale tropical SST variability is lacking.

Fig. 2 a Winter precipitation time series (P), 1871–2007. Correlations (1950–2007): **b** Winter $P \times Z500$. **c** Winter $P \times \text{SLP}$. **d** Winter $P \times \text{SST}$. Thin and bold contours mark two-tailed 90 and 95% significance levels

3.2 Quasi-biennial cycle

In order to investigate the cyclical properties of the precipitation series, a power spectrum analysis was performed using the multi-taper method (MTM), a technique commonly used for short geophysical time series (Thomson 1982; Percival and Walden 1993; Ghil et al. 2002). The MTM multiplies the original time series by a small number of orthogonal windows (discrete prolate spheroidal

sequences) prior to Fourier transform computations. This approach yields multiple independent spectral estimates that are averaged to reduce the variance of the overall spectrum. A small number (K) of tapers are constructed to minimize spectral leakage outside a prescribed half-bandwidth $p/(N\Delta t)$. For the present analysis MTM parameters were set at $p = 2$ and $K = 3$. The method can also be used to reconstruct prominent signals as time series (Park 1992; Park and Maasch 1993), an approach used here to examine the evolution of the QB cycle.

MTM spectra are illustrated in Fig. 3a for the full 1871–2007 precipitation record, and for the recent 1950–2007 period. The full-record spectrum includes a narrow and significant ($p < 0.01$) peak at 2.16 years, in addition to a secondary peak at 12.8 years ($p \sim 0.01$). The QB peak is also significant ($p < 0.05$) over the recent interval. Similar QB and quasi-decadal (QD) spectral peaks are also evident in century-long records of water-year (October–September) inflows to the Sacramento and San Joaquin Rivers of California (Fig. 3b, c). The winter SOI spectrum, illustrated for comparison (Fig. 3d), shows no significant power at either QB or QD frequencies.

The QB cycle in precipitation, as represented by the MTM signal reconstruction, is shown in Fig. 4. The correlation with the untreated series ($r = 0.41$) equates to a 17% fraction of variance explained (r^2) by the QB mode. The reconstructed QB signal (Fig. 4) displays a persistent quasi-decadal amplitude modulation, a predicted feature of an annually-sampled, narrowly-defined QB process with a period slightly offset from an exact 2-year length. A purely sinusoidal oscillation of 26 months (~ 2.17 years), for example, will slowly advance through the seasons so that peaks and troughs occur 2 months later with each cycle and complete a revolution through the calendar in exactly 13 years. Modulation is generated at this 13-year period by an annual sampling regime: When QB peaks and troughs drift into alignment with the sampling season, the amplitude of annual or seasonal means becomes inflated, and when extremes occur out of season, the amplitude is muted. This mathematical sampling effect also predicts an alternation of QB peaks between odd and even years in successive high-amplitude intervals.

The validity of these apparent sampling effects was investigated in the post-1949 precipitation record by comparison to a sinusoidal representation of the QB oscillation generated at monthly resolution (Fig. 5a, b) and sampled and averaged over each November–April winter (Fig. 5c). The sinusoidal QB signal, termed P_{SIN} , was identified by systematically generating sine functions at 0.1 month period increments and 1-month lag increments, and testing correlations with the untreated precipitation series (Fig. 5a). The best least-squares fit ($r = 0.48$, $r^2 = 0.23$) at periods longer than 2 years was found at

Fig. 3 Multi-taper power spectra: **a** Western US winter precipitation. Horizontal lines denote significance levels. **b** Sacramento River (CA). **c** San Joaquin River (CA). **d** Winter SOI. Vertical lines mark precipitation spectral peaks

25.9 months (~ 2.16 years), and is represented by the equation:

$$P_{\text{SIN}} = \sin[2\pi/25.9t + 2\pi(22/25.9)] \quad (1)$$

where t_1 represents January, 1949.

Fig. 4 Quasi-biennial (QB) oscillation in winter precipitation. MTM reconstruction is shown in *bold* over the untreated precipitation record

Fig. 5 25.9-month sinusoidal winter precipitation signal P_{SIN} , 1949–2008. **a** Least-squares sinusoidal fits to P , by frequency. **b** P_{SIN} at monthly resolution. **c** P_{SIN} winter (November–April) averages. **d** Winter means of P_{SIN} compared to MTM reconstruction ($r = 0.93$)

The noted sampling effects could also result from an oscillation slightly shorter than 2 years. An equally strong fit to the precipitation record can be obtained at a period of

22.3 months (~ 1.86 years) (Fig. 5a), a cycle that slowly retrogrades with respect to the seasons. Analysis of monthly pressure data from the vicinity of the NW US indicates, however, that the longer QB period is prevalent. Over the full precipitation record, the best fit was obtained at 26.4 months (2.20 years). The recent interval is emphasized here due to the prominence of the QB signal over this span, and its coincidence with the availability of global atmospheric circulation data. This ~ 60 year span includes 27 complete QB cycles for analysis. The winter-mean series of P_{SIN} is nearly identical to the MTM reconstruction ($r = 0.93$) (Fig. 5d) during the period of overlap.

Monte Carlo simulations were performed to assess the likelihood of the post-1950 sinusoidal fit. Based upon randomly generated Gaussian time series ($N = 58$), the probability of explaining this level of variance with a 25.9-month signal is approximately $p = 0.0005$ (1 in 2000). The probability of obtaining such a fit at any interannual frequency (2–10 years) was estimated at $p = 0.05$.

The persistent influence of the QB signal is revealed by a comparison of precipitation values selected according the upper and lower quartiles of P_{SIN} (1952–2007), capturing the effects of the oscillation when it is aligned with the winter season. Twelve of the 14 winters in the upper P_{SIN} quartile received precipitation in excess of 35 cm; 12 of the 14 lower-quartile winters received less than 35 cm. The precipitation difference between upper and lower P_{SIN} quartiles amounts to 22% of the mean (Student's t test: $p \sim 0.0003$, one tailed).

An inspection of overall precipitation variability from the mid-1950s to late 1960s reveals strong year-to-year reversals (Fig. 6a), in phase with the sinusoidal signal, including an even-to-odd phase shift during the early 1960s. Over more recent decades, the QB signal is partially obscured by lower-frequency variations, however the persistence of the P_{SIN} phase-shifting pattern is evident in year-to-year precipitation differences (ΔP), which highlight the QB time scale of interest (Fig. 6b). Interannual precipitation spikes (> 8.5 cm wetter than an adjacent winter) are seen only in even winters during the late-1950s, mid-1970s to mid-1980s, and mid-2000s (1956, 1958, 1974, 1976, 1978, 1980, 1982, 1986, 2006), and only in odd winters during the late 1960s and mid-1990s (1965, 1969, 1993, 1995), matching the pattern of odd–even reversals in P_{SIN} . Moreover, multi-year periods of quiescence in ΔP coincide with intervals when monthly P_{SIN} extremes have occurred in summer (~ 1961 , 1973, 1989, and 2001) (Fig. 6c). These decadal patterns of interannual variability suggest a stable, nonseasonal QB signal that periodically disappears from winter sampling for several years before returning in the opposite phase, as predicted by the sinusoidal model.

Fig. 6 Winter P_{SIN} , 1950–2007. **a** Comparison to the raw precipitation record. **b** Comparison to interannual precipitation differences (ΔP). **c** Winter P_{SIN} amplitude, compared to 7-year means of absolute interannual differences ($|\Delta P|$)

The time scale of amplitude modulation is a sensitive diagnostic of the underlying QB period. Modulation periods in the range of 11–21 years correspond to QB cycles in the narrow span of 2.1–2.2 years, according to the relation:

$$T_{\text{MOD}} = (1 - 2T_{\text{QB}}^{-1})^{-1} \quad (2)$$

where T_{QB} and T_{MOD} are the quasi-biennial period and its modulation period (in years) due to annual sampling (Fig. 7a). The decadal modulation of the QB precipitation cycle also distinguishes it from the longer QBO of the stratosphere, which rotates through the calendar just once every 6–7 years. The difference in QB periods is apparent in the autocorrelation functions (ACFs) of P (1950–2007) and the winter 30 hPa QBO (1955–2008) (Fig. 7b–e). Persistent QB variability produces prominent sawtooth patterns in the ACFs of both series (Fig. 7b, d), however the timing of amplitude peaks and odd–even phase-shifts in

Fig. 7 Amplitude modulation of an annually-sampled QB cycle. **a** Modulation period as a function of QB period. Autocorrelation functions (ACFs) of **b** winter P **c** winter P_{SIN} **d** winter 30 hPa equatorial winds (stratospheric QBO) **e** 28.2-month sinusoid fitted to the QBO. Brackets mark the intervals between consecutive AC values of the same sign, which indicate the period of QB phase-shifting with respect to the annual cycle

the ACFs are clearly different. QB phase-shifting in the ACF of P displays an expected 13-year period, while shorter 5–7 year periods are seen for the QBO. To illustrate the idealized patterns, ACFs are also plotted for P_{SIN} , and for a 2.35-year (28.2-month) sinusoid fitted to the QBO (Fig. 7c, e).

In combination with the spectral results, these decadal sampling effects point to the existence of narrowly-defined QB oscillation, distinct from the QBO, which maintains its frequency and phase over multiple decades. This unexpected behavior contradicts widely-held assumptions about the nature of QB activity at the earth's surface, and tropospheric climate dynamics in general. As postulated for the TBO, quasi-cyclic patterns in the troposphere are generally thought to arise from feedback mechanisms, with

ENSO (Bjerknes 1969; Zebiak and Cane 1987) as the leading example. By contrast, Western US precipitation variability appears to be regulated in part by a strongly periodic QB process that maintains its identity and phase in spite of strong interference by the seasonal cycle. The well-defined structure of the QB oscillation suggests that it may not simply be a random occurrence or a QB tendency for interannual reversals (as proposed for the tropical TBO), but rather a particular periodic phenomenon with a specific frequency.

Several station precipitation records with strong correlations to P_{SIN} are illustrated in Fig. 8. These sites are found primarily at central and slightly northern latitudes of the Western US, with greatest correlations seen from central California to southern Oregon and Idaho, a zone which includes the agriculturally important Sacramento, San Joaquin, and Snake River Basins. In this area, correlations between grid point precipitation and P_{SIN} generally lie in the range of 0.4–0.5 (16–25% of variance explained),

exceeding the influence of ENSO over most of the Western US (see Fig. 1d). If the QB cycle maintains its identity into the future, it may offer a complementary tool for probabilistic hydroclimate forecasting, particularly where ENSO correlations are weak. The P_{SIN} correlation with ΔP ($r = 0.63$) is comparable to the strongest ENSO-precipitation correlations in the SW US, suggesting the interannual precipitation difference as a good object for prediction.

4 Winter QB teleconnections

The remainder of the analysis uses the periodic signal P_{SIN} as an index to examine correlations with gridded climate series over the globe. Its use is justified by its close match to the objective MTM signal and by the real amplitude modulation and phase-shifting patterns detailed in Sect. 3.2. This approach is designed to examine large-scale teleconnections with the apparently periodic QB oscillation, using the same correlation method as for overall precipitation. A time domain approach is pursued in order to examine and illustrate the oscillation as it is realized in the climate system, including the characteristic interaction with the annual cycle. The use of a monthly sinusoid also facilitates nonseasonal analysis of the QB cycle and readily permits future projection for predictive purposes. The use of a sinusoidal signal as a climate teleconnection index requires care in the assessment of statistical significance. Standard parametric methods for computing correlation significance levels demand that both series be serially uncorrelated, Gaussian stationary processes. Monte Carlo techniques were therefore used to estimate significance levels.

The circulation signatures of the QB precipitation cycle were first investigated by mapping correlations between the winter series of P_{SIN} and global gridded Z500, SLP, and SST. Correlation significance was estimated by comparing the actual P_{SIN} correlation value at each reanalysis grid point to a synthetic null distribution obtained by its correlation against 10,000 random white noise series. Because QB variability itself has strong implications for interannual serial correlation, no attempt was made to pre-whiten the gridded winter series to bring them toward a closer match with the synthetic white noise null distribution. To the degree that the interannual gridded series display lag-1 autocorrelation, a white noise null hypothesis provides a conservative estimate of correlation significance. Interannual persistence in the climate system tends to obscure QB activity, which is synonymous with lag-1 anti-persistence. Two-tailed 95% significance levels, obtained from the 2.5 and 97.5 percentiles of the synthetic correlation distribution, were identified at $r = \pm 0.26$. This significance

Fig. 8 Local QB precipitation variability. **a** Mapped correlations between winter means of P_{SIN} and Western US grid box precipitation, 1950–2007. Shading indicates 95% correlation significance. Negative contours are dashed. Capital letters mark sites plotted below. **b** Four precipitation station records (black) displaying strong correlation to P_{SIN} (gray)

threshold is comparable to that expected of an annually-sampled white noise process ($N = 58$ years).

The strongest (negative) Z500 correlations with the winter series of P_{SIN} (Fig. 9a) were found over the NW US, as for the untreated precipitation index P . Nearby positive correlation centers occur over the Aleutian Islands and Baja California. This three-cell correlation pattern connects QB precipitation enhancement to eastward shifts in the Aleutian Low (the region of North Pacific cyclone maturation) and to intensification of the meridional Z500 gradient over the west coast of North America. The winter Z500 difference between gridpoints near Baja California (25°N , 112.5°W) and Seattle (47.5°N , 125°W) correlates with P_{SIN} at $r = 0.55$ ($r^2 = 0.30$) (Fig. 10a). The QB signal is thus somewhat more prominent in this Z500 dipole than in precipitation. Consistent with these results, a

Fig. 9 Winter climate correlations with P_{SIN} . **a** Z500. **b** SLP. **c** SST. Thin and bold contours mark 90 and 95% significance levels

Fig. 10 Winter circulation comparisons to P_{SIN} . **a** Quasi-meridional Z500 difference over western North America between grid points near Baja California (25°N , 112.5°W) and Seattle (47.5°N , 125°W) (black), compared to winter P_{SIN} (gray). **b** Interannual differences in the winter Z500 gradient versus P_{SIN}

significant positive P_{SIN} correlation was also obtained with the 500 hPa zonal wind speed over central California at 37.5°N , 120°W ($r = 0.51$, $r^2 = 0.26$). Onshore winds in the mid-to-low troposphere contribute to moisture advection and orographic precipitation enhancement in the mountains of California and the broader Western US. Linear regression of California zonal wind on P_{SIN} indicates a peak QB cycle amplitude of 14%, equating to a peak-to-trough difference of 3.1 m s^{-1} about a mean value of 11.1 m s^{-1} .

The QB signal in winter precipitation can thus be directly linked to a similar oscillation of mid-tropospheric pressure and wind over the NE Pacific. The magnitudes of P_{SIN} correlations with first winter differences of California zonal wind ($r = 0.68$, $r^2 = 0.46$) and the western North America Z500 dipole ($r = 0.72$, $r^2 = 0.52$) (Fig. 10b) indicate that roughly half of the variance in these winter-to-winter circulation changes can be explained by a purely sinusoidal QB signal over the past ~ 60 years.

The zones of positive P_{SIN} -Z500 correlations over the Aleutians and Baja California form part of a midlatitude arc that stretches from the North Pacific to Europe, traversing North America and the Atlantic (Fig. 9a). A similar meridional pattern is seen in the SH, suggesting

the involvement of a global-scale zonally-symmetric atmospheric mechanism. In the SH, significant positive Z500 correlations are seen near the Antarctic coast at the Date Line, an apparent analogue of the Aleutian cell in the Northern Hemisphere (NH).

Significant winter P_{SIN} correlations with SLP (Fig. 9b) include an east–west contrast over the N. Pacific, matching the pattern in Z500. A pattern resembling the Southern Oscillation (SO) is also seen over the tropical Pacific, such that La Niña conditions are associated with anomalous wetness over the Western US at the QB frequency. The significant SO-related correlation centers display a quasi-meridional alignment at subtropical latitudes. In the SH, significant P_{SIN} -SLP correlations are found primarily in the midlatitudes. Negative SST correlations with P_{SIN} are significant in the Gulf of Alaska ($p < 0.05$) (Fig. 9c), and marginally significant ($p < 0.10$) values are seen in the eastern equatorial Pacific, further supporting a simultaneous relationship between La Niña and precipitation increases at the QB frequency. However, the North Pacific P_{SIN} -SLP correlation pattern in Fig. 9b differs substantially from the overall SOI teleconnection pattern shown in Fig. 1c. Most notably, SLP in the NE Pacific oscillates in phase with the western tropics on the QB time scale, whereas pressure anomalies in these regions are anti-phased in the canonical ENSO pattern.

Over the entire Pacific sector, significant Z500 and SLP correlations form a meridional standing wave pattern that extends from the Arctic to the Southern Ocean (Fig. 9a, b). Significant correlation cells are located in regions where precipitation and/or pressure oscillations have been previously documented over the range of 2.1–2.2 years. These areas include the Arctic (Trenberth and Shin 1984; Venegas and Mysak 2000), the Aleutians (Angell et al. 1969; Angell and Korshover 1974), the subtropical NW Pacific (Tomita and Yasunari 1996; Tomita et al. 2004), Tasmania-New Zealand (Trenberth 1975, 1980), and the Pacific coast of Patagonia (Villalba et al. 1998). These ~ 26 -month regional cycles appear to vary together with Western US winter precipitation as constituents of a single global-scale oscillation. The pressure signatures of P_{SIN} suggest that the QB precipitation signal is a regional effect of a strongly periodic, Pacific-wide standing pressure wave that oscillates at a unit nearly from pole to pole.

5 Monthly QB teleconnections

As detailed in Sect. 3.2, the observed temporal patterns of QB cycle modulation and phase-shifting suggest the existence of a nonseasonal atmospheric oscillation which is hidden from winter sampling when cycle extremes fall in summer and neutral conditions occur in winter. This

hypothesized nonseasonal signal was investigated by computing global correlations with monthly values of P_{SIN} . Gridded monthly climate series were first deseasonalized by subtracting monthly means and dividing by monthly standard deviations. Because climatic persistence is generally much stronger on monthly than interannual time scales, anomaly series were pre-whitened before calculating correlations with P_{SIN} . Pre-whitening was accomplished for each gridpoint series with significant lag-1 correlation by obtaining residuals from a first-order autoregressive [AR(1)] model, and testing for remaining autocorrelation with the Durbin-Watson statistic. In cases of residual autocorrelation, AR model order was iteratively increased until serial correlation was effectively removed. All anomaly series were successfully pre-whitened with AR(3) and lower-order models.

This procedure was performed in order to facilitate the use of white-noise simulations for Monte Carlo significance testing. Significance levels were obtained by correlating synthetic white noise series with the actual P_{SIN} index, effectively incorporating the properties of the sinusoid in the simulation. The 90, 95 and 99% two-tailed significance levels were estimated at $r = \pm 0.062, 0.073, 0.096$, respectively, closely matching levels expected for two white noise stationary processes ($N \sim 715$).

Monthly P_{SIN} correlations with pre-whitened gridded climate series are illustrated in Fig. 11. Z500 correlations (Fig. 11a) are globally strongest over the North Pacific and generally similar in structure to the winter patterns. Significant correlations encompass the North Pacific in the form of a meridional standing wave stretching from the Gulf of Alaska to the tropics. Included in this pattern is the meridional dipole over the North American west coast found in the winter analysis. This regional Z500 oscillation thus appears as a year-round signature of the 25.9-month signal that contributes to variability in winter precipitation. A similar meridional correlation structure is apparent in the South Pacific, though areas of significance are limited to isolated locations in the mid-latitudes as part of a wavenumber-3 pattern spanning the Southern Ocean [see also Trenberth (1975)]. Globally, monthly Z500 correlations also exhibit symmetry about the equator. Monthly SLP correlations with P_{SIN} (Fig. 11b) include the SO in the tropical Pacific, though 95% significance is reached only at isolated points in the east. Negative SST correlations are also significant ($p < 0.05$) in the eastern Pacific El Niño region (Fig. 11c).

6 Discussion

The 25.9-month signal observed in Western US winter precipitation does show evidence of a nonseasonal identity, based on its significant climate correlations at monthly

Fig. 11 Monthly pre-whitened pressure anomaly correlations with P_{SIN} . **a** Z500. **b** SLP. **c** SST. Thin and bold contours mark 90 and 95% significance levels

resolution. The QB oscillation thus appears to be a continuous, global-scale, quasi-sinusoidal phenomenon that systematically advances through the seasons, as suggested by the modulation of the winter QB signal. In the vicinity of western North America, the nonseasonal QB oscillation in Z500 is strongest over SW Canada ($r = -0.126$ at 50°N , 127.5°W) and Baja California ($r = 0.102$ at 27.5°N , 122.5°W) (difference: $r = 0.165$), where it precedes P_{SIN} by one month. These correlation levels, while strongly significant ($p < 0.01$), are considerably smaller in magnitude than those found in the winter analysis. This difference reflects the fact that a QB oscillation in a noisy background is inherently most influential over seasonal to annual time integrations of ~ 6 to ~ 12 months, and less effective on the monthly time scale. In order to better

illustrate the nonseasonal oscillation, the North Pacific meridional Z500 difference is shown (Fig. 12a, c) after smoothing monthly anomalies with a 9-month running mean and removing frequencies longer than 120 months with a low-pass filter, leaving intact interannual-scale, nonseasonal variations.

A visual inspection of this dipole index reveals its year-round and long-term adherence ($r = 0.42$) to the periodic 25.9-month signal that leads P_{SIN} by 1 month (Fig. 12a). Running 121-month correlations remain positive throughout the record (Fig. 12b), indicating the signal's persistence. A particularly strong fit ($r > 0.50$) is apparent from the mid-1960s to the late 1980s. The QB signal was obscured during the mid-1990s, however dominant QB activity resumed after 2000 *in phase with the earlier signal*. This behavior suggests the existence of an ever-present 'clockwork' QB process in the NE Pacific that may be occasionally be concealed by other interannual-scale anomalies, and confirms the inferences drawn from the modulated winter signal. Since 1990 (Fig. 12c), the meridional Z500 wave can also be seen to maintain its ~ 26 -month period in a gradual drift through the seasons up to the end of the analysis in July, 2008.

In order to locate the clearest evidence for a 25.9-month signal in the ENSO region, SLP and SST correlations were computed with systematically phase-shifted copies of P_{SIN} . The strongest 25.9-month variability in ENSO comes with a signal that precedes P_{SIN} by 4 months. Strongest SST correlations occur in the eastern tropical Pacific at 8°S , 120°W ($r = -0.126$). In SLP, the same QB signal is most prominent in an SO dipole defined by the anomaly difference between grid points at 5°S , 105°W in the eastern tropical Pacific, and 17.5°S , 117.5°W near the NW coast of Australia (difference: $r = 0.159$) (Fig. 12d, f). Consistently positive running correlations with the 25.9-month signal are evident since the mid-1950s (Fig. 12e), though at generally lower levels than seen in the western North America Z500 dipole.

The 3–4 month phase delay between ENSO and NE Pacific QB signals is consistent with that expected from an ENSO-driven teleconnection. However, the weak presence of the QB signal in the tropics contradicts this attribution. For example, a 26-month signal in El Niño region SST has been previously identified (Zhang et al. 1998), however this QB mode was found to be roughly one-third as strong as a ~ 4 -year mode, supporting prior studies noting the relative obscurity of the QB mode relative to lower-frequency variations (Rasmusson et al. 1990; Barnett 1991; Ropelewski et al. 1992; Jiang et al. 1995). These studies have observed that the QB mode in ENSO, though statistically discernible, exerts only a modest influence on the overall state of tropical climate, which drives extratropical responses in real time. The limited role of the ENSO QB

Fig. 12 Monthly QB signals in the North and tropical Pacific.

a 1949–2008 comparison of monthly P_{SIN} (back-shifted 1 month) and the monthly Z500 anomaly difference between 27.5°N, 122.5°W (Baja California) and 50°N, 127.5°W (SW Canada), filtered to highlight interannual variability. **b** 121-month running correlation **c** 1990–2008 comparison. **d** Southern Oscillation dipole between sites of strongest QB activity in the ENSO region, compared to P_{SIN} back-shifted 4 months. **e** 121-Month running correlation **f** 1990–2008 comparison. Vertical lines correspond to January of each year

mode is visually apparent in the SO index depicted in the right column of Fig. 12.

In order to search for evidence of the P_{SIN} signal in the global climate more generally, lagged copies of P_{SIN} were correlated against all gridded monthly time series to determine patterns of QB signal strength, regardless of the local phase. Signal strength was measured in terms of fractional variance explained (r^2) by each locally-fitted 25.9-month oscillation, using the pre-whitened monthly climate series. Squared correlations were calculated at all lags by sliding the monthly P_{SIN} index in 1-month increments from zero to 25 months, approximately a 14° phase resolution. This approach is comparable to previous methods for measuring QB climate activity by computing the variance explained by bandpass-filtered QB signals (Barnett 1991; Ropelewski et al. 1992). A primary difference here is the use of a sinusoidal signal, designed to identify the origins of the specific QB phenomenon observed in Western US winter precipitation. Additionally, the pre-whitening serves to standardize the autocorrelation structures of monthly gridded climate series, which vary considerably in space. Maximum r^2 values for Z500, SLP and SST are plotted by latitude in Fig. 13. Signal strength was measured separately for 1949–2008 (Fig. 13a) and 1990–2008 (Fig. 13b) periods. The latter analysis captures recent adherence to the long-term QB signal at each grid point.

Maximum 25.9-month signal strength (Fig. 13a) is greater overall in Z500 and SLP than in SST. The greatest prominence of the QB signal over the full record is found in monthly Z500 at subpolar Northern latitudes, with a maximum local r^2 value of 0.020 at 75°N , 85°W over NE Canada. Signal strength in SLP is greatest in midlatitudes of the SH, whereas SST r^2 values are most prominent in the tropics with a peak ($r^2 = 0.016$) in the El Niño region. Over the recent interval (Fig. 13b), continued QB signal strength in all three variables is most prominent in the extratropics of both hemispheres, exceeding values found at tropical latitudes. Figure 14 illustrates recent pronounced QB variations in Z500 in the extratropics of each hemisphere over the southern Indian Ocean and SW Canada. Note that each example reflects recent conformity to long-term QB signals, and therefore represents a continuation of prior behavior. This strong, continued QB activity is difficult to attribute to a tropical source in the absence of comparable QB signals at low latitudes, particularly since 1990 (Fig. 13b).

The persistent QB signal in Western US winter precipitation and extratropical circulation would therefore seem to require another explanation. The extratropical prominence of climate variations associated with P_{SIN} suggests high-latitude origins involving both hemispheres. Figure 15 shows the evolution of Z500, SLP, and SST variations at the 25.9 month frequency, depicted as lagged

Fig. 13 Global patterns of 25.9-month sinusoidal signal strength (r^2), by latitude. **a** 1949–2008. **b** 1990–2008

Fig. 14 Recent QB activity in Z500 in each hemisphere **a** S. Indian Ocean (42.5°S, 92.5°E). **b** SW Canada (50°N, 127.5°W). The S. Indian signal leads by 8 months. Gray curves illustrate locally-fitted 25.9-month signals. The SW Canada curve, when reversed in sign, precedes P_{SIN} by 1 month

correlations with P_{SIN} . The sequence shows the development of the oscillation in the months prior to its appearance in ENSO and the NE Pacific and Western US. The

animation begins with mapped climate correlations against a version of P_{SIN} that has been back-shifted 8 months, and progresses in 4-month steps to lag-zero.

The cycle progression commences with a pair of significant Z500 meridional standing waves over North America and the Australian sector of the Southern Ocean at lag -8 (Fig. 15a, b). The Canadian Arctic center of the Northern Z500 wave is the site of the most prominent 25.9-month signal globally. The Southern wave bears a strong resemblance to the Southern Annular Mode (SAM), the leading mode of SH pressure variability, characterized by an anomaly contrast between polar and middle latitudes (Thompson and Wallace 2000). The correlation pattern in SLP is considerably weaker, but more clearly reflects zonally-symmetric, in-phase variations over both polar caps. The SST correlations include a nascent La Niña pattern over the tropical Pacific. Over subsequent months, the Northern meridional wave appears to propagate energy toward the tropical Pacific, traversing the NE Pacific en route. The midlatitude Indian Ocean cell in the SAM pattern migrates northward toward the tropical Indian Ocean and the western tropical Pacific where it contributes to patterns resembling the TBO and ENSO at lag -4 (Fig. 15c, d). This equatorward propagation suggests extratropical origins for these tropical QB phenomena. Extratropical origins for tropical QB activity are suggested by the findings of Meehl (1987), which identified significant circumpolar pressure anomalies around Antarctica in the months prior to QB variations in Indian monsoon rainfall. Mann and Park (1996) also suggested that more prominent high-latitude QB activity might contribute to weaker, synchronous ENSO-like QB variability in the tropics. By lag-zero (Fig. 15e, f) the patterns related to Western US winter precipitation have fully evolved in the North Pacific. In the framework presented here, the precipitation cycle in the Western US appears to originate with the meridional wave that develops over North America, rather than the QB ENSO pattern. Z500 and SLP animations are also included in movie form in the Electronic Supplementary Material (Animation 1 and Animation 2).

Given the high-latitude prevalence of the ~ 26 -month signal in global pressure, it is concluded that the essential QB mechanism involves the quasi-synchronous oscillation of hemispheric annular mode patterns, generically termed here a Biennial Annular Mode Oscillation (BAMO). Oscillations of ~ 26 -months related to the TBO (Barnett 1983) and QB ENSO (Zhang et al. 1998) appear to be after-effects of converging meridional pressure waves originating at high latitudes. A strongly periodic extratropical forcing with a time-varying influence might reasonably account for the nature of the ~ 26 -month mode in the SO, which appears as an obscure but persistently identifiable signal. A subtly different interpretation is that BAMO and SO components, which oscillate approximately in

Fig. 15 Lagged monthly climate correlations with P_{SIN} at lags of -8 months (*top row*), -4 months (*middle row*), and 0 months (*bottom row*). The sequence illustrates the development of the QB oscillation in the months prior to its effects in the North Pacific and Western US Method and color scale match those for Fig. 11

quadrature, represent different aspects of a single globally-encompassing ~ 26 -month oscillation.

7 Summary and outlook

This study has sought to explore the nature and origins of a significant QB oscillation in Western US winter precipitation with a period of $2.16 \pm \sim 0.05$ years, evident both from 1871 to 2007 and 1950–2007. Over the latter interval this oscillation is found to display a quasi-decadal modulation pattern suggesting essentially periodic ‘clockwork’ behavior and systematic interaction with the annual cycle. This modulation pattern, evident in the timing and direction of year-to-year differences in the untreated record, was used as a justification for representing the oscillation as a sinusoidal signal. This sinusoidal QB mode is primarily evident in winter precipitation at central to northern latitudes of the Western US, where the magnitude of its role on interannual precipitation differences is comparable to the largest effects of ENSO in the extreme southwest. The QB mode in precipitation, due to its shorter period, is distinct from the QBO of stratospheric winds.

The QB cycle in winter precipitation results proximally from a nonseasonal, meridional standing pressure wave over the North American west coast, prominent at 500 hPa. Strongly periodic behavior is also apparent in the winter and monthly evolution of this dipole pattern over the past

60 years. This regional oscillation is a component of a meridional wave spanning the Northern Hemisphere, somewhat resembling the NAM. The annular structure, with a polar center over N. Canada, develops over North America in the 4–8 months prior to its strongest influence in the Western US. The QB wave in the NH develops in conjunction with a similar annular mode wave in the SH, most evident in the Indian and Australian sectors. The QB mechanism is therefore judged to be interhemispherically synchronous and global in scale, and is termed here a Biennial Annular Mode Oscillation (BAMO).

The QB cycle in precipitation displays significant connections to the Southern Oscillation, however, the ~ 26 -month signal in the SO and tropical SST is only a minor component of overall variability, arguing against a leading role for ENSO in the generation of more prominent extratropical QB variability. It is instead concluded that the ~ 26 -month tropical zonal wave likely results as an after-effect of BAMO forcing from high latitudes, where the QB signal is most prominent, or as a component of a global-scale atmospheric wave.

It remains to be explained why Western US winter precipitation and the related circulation exhibit such strongly periodic QB behavior over the past 60 years (and perhaps longer), and why significant meridional QB waves oscillate in both hemispheres in a quasi-synchronous fashion. The global scale of the QB cycle and its relative prominence at mid-tropospheric levels suggest that its

origins more likely reside in the realm of resonant planetary waves than ocean–atmosphere interactions. Paradoxically, the high-latitude zones of the strongest QB signals are also those most influenced by the seasonal cycle. These facts are suggestive of some periodic external forcing, and it is notable that ~ 26 month oscillations have also been observed in solar activity (Shapiro and Ward 1962; Apostolov 1985; Soukharev and Hood 2001). QB variations in sunspots and irradiance however are miniscule and irregular, and any steady QB solar forcing of the atmosphere may perhaps more plausibly involve magnetic-electrical forcing of atmospheric waves via the solar wind and interplanetary magnetic field, as proposed for the interannual variability of the North Atlantic Oscillation (NAO) (Boberg and Lundstedt 2002, 2003) and the NAM (Lu et al. 2008). Moreover, QB oscillations slightly longer than 2 years are independently established features of the solar magnetic field (Stenflo and Vogel 1986; Stenflo and Güdel 1988; Obridko and Shelting 2001; Knaack and Stenflo 2005). In this regard, it is interesting to note that the initial zones of significant QB activity depicted in Fig. 15, including the most prominent signal globally, lie in proximity to the magnetic poles, marked in the initial Z500 lag-correlation map (IGRF-11 model 1975 data, <http://wdc.kugi.kyoto-u.ac.jp/poles/polesexp.html>). Relationships between QB oscillations of the solar magnetic field and tropospheric circulation are being investigated further (Johnstone 2008). Regardless of origins, the short QB time scale is conducive to further monitoring and evaluation of the oscillatory patterns identified here at regional and global scales.

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