A Footprint of Solar Variability on Climate

Alexander Ruzmaikin

Jet Propulsion laboratory, California Institute of Technology, Pasadena, California

Short title: A FOOTPRINT OF SOLAR ACTIVITY ON CLIMATE

Abstract. An 11-year variation is shown here to exist in an index of the atmospheric variability defined as the difference between the North Atlantic Oscillation and Arctic Oscillation indices. It is argued that this index is related to the second leading mode of the variability. The appearance of the solar activity signal in this mode is in accord with modeling of the solar UV forcing by *Shindell et al.*, [2000]. However, the solar activity input to the atmosphere is weak and its effect has to be amplified. Here the amplification by stochastic resonance due to assistance of other drivers of the variability (volcanic aerosols, Quasi-Biennial Oscillation and El Nińo) is discussed. It is shown that these drivers have considerable power for the mode excitation at a half of the solar cycle period (5.5 years)–the necessary condition for stochastic resonance.

Introduction

In spite of numerous indications of possible influences of solar variability on the Earth's climate, specific mechanisms for the influence are not clearly identified [Hoyt and Schatten, 1997]. New possibilities have been opened by recent atmospheric studies which show that disturbances produced in the stratosphere propagate down to the troposphere [Baldwin and Dunkerton, 1999]. The leading mode of atmospheric variability in the Northern Hemisphere, called the Arctic Oscillation [Thompson and Wallace, 1998], has an almost longitudinally symmetric structure extending from the Earth's surface to the stratosphere. The mode describes wintertime climate anomalies in the sea-level pressure coupled to surface air temperature anomalies.

Because this natural mode is sensitive to rather modest external forcing, recent studies have begun to investigate whether solar variability contributes to the mode variability [Balachandran et al., 1999; Shindell et al., 2000]. The solar UV flux at 200 nm in the stratosphere varies noticeably (by 6-7% from solar maximum to solar minimum) during the solar cycle which leads to changes in the ozone concentration and hence the temperature gradient in the stratosphere. These changes affect the refraction index for the planetary waves propagating from the troposphere and interacting with the zonal wind [Kodera, 1995]. However the calculated perturbations in zonal wind and temperature initiated through this mechanism are about a factor of four smaller than the observed changes (4% in ozone and 5°K in temperature near the low-latitude stratopause) [Hood et al, 1993]. The question arises whether there is a sea-level footprint of solar forcing of the AO. And, if the answer is yes, how the solar forcing is amplified?

It is shown here that an 11-year solar signal is present in the sea-level variations; but not in the index of the leading AO mode. Instead, the variation is visible in the difference between the North Atlantic Oscillation (NAO) and the Arctic Oscillation. We associate (but not identify) this difference with the second, longitudinally non-symmetric

Empirical Orthogonal Function (EOF). This supports the result of the GISS modeling which indicated a response to the solar forcing only in the second EOF [Shindell et al., 2000]. We also propose a mechanism of excitation of this mode that can amplify the solar forcing through the assistance of other drivers in the stratosphere and troposphere. The mechanism is based on "stochastic resonance" (for a recent review see *Gamatoni et al.* [1998]; suggestions that this mechanism amplifies the 11-year solar forcing of climate has been made earlier in the papers by *Lawrence and Ruzmaikin* [1998] and *Ruzmaikin* [1999]).

Search for a Solar Activity Signal

We use the Arctic Oscillation index and the North Atlantic Oscillation index for the last century. We look at time-dependent spectra of these indices because the standard Fourier spectrum is an integral over the entire time interval and the 11-year signal could be intermittent. The best technique for this purpose is wavelet analysis that provides spectral information as a function of time. Figure 1 shows the wavelet spectra of the AO and NAO indices. There may be some indications of a decadal mode in the NAO index consistent with the earlier results [Hurrell, 1995], but the spectrum is obscured by the presence of other modes. There is even less indication of a decadal signal in the AO mode. Note, however, that both spectra show considerable power near the 5-6 year scale.

We now consider a combined index, D, defined as the difference between NAO and AO indices. Bearing in mind that the probable mechanism, based on the interaction of planetary waves with the stratospheric zonal flow, is seasonally dependent, the spectra for winters and summers are calculated separately. For comparison, the sunspot number wavelet spectrum is calculated as well. The wavelet spectrum of the D index (Figure 2) shows a clear 11-year signal. Although the signal is present for both seasons it is much stronger in the winter data in accord with the fact that the planetary activity in the

Northern Hemisphere is most pronounced during winter.

What is the physical meaning of this index? The NAO index is defined as the sea surface pressure difference between Portugal (24°) and Iceland (65°). The AO index is defined as the time-dependent amplitude of the first EOF of the Arctic Oscillation [Thompson and Wallace, 1998]. Therefore the difference between these indices carries information about the amplitude of the second EOF (and higher EOFs). This finding is in accord with the results of the GISS modeling which showed that the solar forcing excites only the second EOF [Shindell et al., 2000]. Thus, there is a footprint of the solar forcing on the ground of the atmospheric variability.

There is an additional support of this result. One of the key element of the mechanism, by which the solar UV variability can affect the climate, is planetary waves generated by the land-ocean temperature contrasts and topography. There is a well known atmospheric response to the land-ocean contrast called the COWL (Cool Ocean Warm Land) pattern. We notice that this pattern corresponds to the second EOF [Corti et al., 1999].

Amplification of Solar Forcing by QBO, Volcanic Aerosols, and ENSO

Labitzke and van Loon, [1988] found that the stratospheric response to the solar forcing depends on the phase of Quasi-Biennial Oscillation (QBO). Similarities between the stratospheric response to volcanic aerosols and to the solar forcing were demonstrated by *Kodera*, [1995]. Here we discuss a possible non-linear mechanism of cooperative forcing of atmospheric variability by the stratospheric aerosols, QBO and solar changes induced by the UV flux.

We suggest here that volcanic aerosols, QBO in the stratosphere, and the ENSO in the troposphere are the primary drivers of the second leading mode of the atmospheric variability. (They also excite the first mode.) Note that on short time scale the mode is more effectively excited by eddy activity, specifically by synoptic ($<$ 10 day periods) and stationary waves [Limpasuvan and Hartmann, 1999]. We consider here low-frequency excitation. The solar forcing due to the 11-year cyclic changes in the UV flux is expected to be weak.

To illustrate the idea of amplification we consider two climate states: a state with the high D index (called $+C$); and a state with the low D index (called $-C$). These states are assumed to be separated from each other by a threshold. Note that the states are not supposed to be eigen functions of the system, they could be attractors found from non-linear dynamics. In the context of the atmospheric variability under consideration one state corresponds to " warm high latitudes - cold lower latitudes" situation and the other to "cold high latitudes - warm lower latitudes" situation. Let $p_{\pm}(t)$ be the probability that at time t the climate is in $+C$ or C state so that $p_{+} + p_{-} = 1$. Let $W_{\pm}(t)$ be the transitional probabilities from $+C$ to $-C$ state (upper subscript) and from $+-C$ to $+C$ state (lower subscript)). Then [McNamara and Wiesenfeld, 1989]

$$\frac{dp_{\pm}}{dt} = -W_{\mp}p_{\pm} + W_{\pm}p_{\mp}. \tag{1}$$

These linear equations can easily be solved when the transitional probabilities are given. The solutions used below are discussed in detail in the review by *Gamatoni et al.,* [1998].

Let W_0 be the transitional probability caused by an aperiodic (noisy) forcing (provided by a combine action of stratospheric aerosols, QBO and ENSO) with a mean-squared amplitude a. The adddition of a weak periodic solar signal $scos(\Omega t)$ leads to a periodic modulation of the transition rates similar to the Arhenius-Kramers law in chemical kinetics:

$$W_{\pm}(t) = W_0 \exp(\mp\epsilon cos(\Omega t)), \ \ \epsilon = \frac{Cs}{a}. \tag{2}$$

The parameter ϵ is considered to be small since the aperiodic forcing provides effective transitions between the climate states (i.e. C/a is of the order of unity). In this case the solution of Eqs (1) can easily be obtained. To the first order in ϵ:

$$2p_+(t) = (2\delta_{0+} - 1 - \epsilon cos\varphi)e^{-2W_0 t} + 1 + \epsilon cos(\Omega t - \varphi), \tag{3}$$

where $\delta_{0+} = 1$ when the system is initially in the $+C$ state and $\varphi = \arctan(\Omega/2W_0)$. Obviously, $p_- = 1 - p_+$. This solution allows us to calculate characteristics of the system response to the forcing. Particularly interesting to us is the residence- time distribution [Gamatoni et al., 1998]

$$R(t) \propto [1 - \frac{\epsilon^2}{2} cos(\Omega t)]e^{-W_0 t}. \tag{4}$$

The residence time is defined as the time between a transition from one state to another. The exponentially decreasing function describes the distribution in the absence of periodic forcing, with the mean residence time proportional to $1/W_0$. With the periodic forcing the distribution, Eq.(4), exhibits peaks of decaying amplitude at $t_n = 2\pi(n - 1/2)/\Omega, n = 1, 2, 3, ...$ The largest peak is at a half of the forcing period: $t_1 = \pi/\Omega$. The area under the peaks is a measure of synchronization between the periodic forcing and a switching the states due to aperiodic forcing [Gamatoni et al., 1998]. If the period of solar forcing, i.e. the position of the first peak, is much smaller than the mean residence time than the solar forcing is not effective because the system is not likely to jump from one state to another. Likewise, if the period of solar forcing is much larger than the mean residence time, i.e. positioned at place where the distribution is very small, the periodic impact has practically zero probability to affect the system. However, the synchronization of transitions driven by periodic and aperiodic forcing becomes possible when the mean residence time is close to a half of the solar forcing period. Thus in order to assist the solar periodic forcing the aperiodic forcing should drive the system with sufficient power at $11/2 = 5.5$ years.

The wavelet spectra of the NAO and (to lesser extent) the AO index show significant power near 5.5 years (Figure 1). This supports the idea that a combined driving of all non-solar forcings can indeed assist the solar forcing.

Let us see how the different forcing contribute. The spectrum of forcing by the stratospheric aerosols shown in Figure 3 has a significant power near the critical $(5.5\text{year})^{-1}$ frequency. The ENSO spectrum has also power in the same frequency band [Ruzmaikin, 1999]. The QBO is a quasi-periodic process with characteristic time about 28 months [Baldwin et al., 2000]. However, the effective excitation of the AO mode has been associated only with the west phase QBO [Holton and Tan, 1997; Labitzke and van Loon, 1988]. Thus the characteristic time of the QBO forcing related to the solar activity is 56 months (4.7 years), i.e. also close to the critical frequency. Therefore we conclude that the combined aperiodic forcing has a significant power near the critical frequency and hence can assist the solar signal through stochastic resonance.

Discussion

The stratospheric drivers of the atmospheric variability involve the mechanism first suggested by *Hines* [1974] : the changes in the refraction index induced by the ozone and temperature affect the interaction of the planetary waves with the zonal flow in the stratosphere. The differences among different drivers are in the strength and spatial pattern of forcing. For example, the volcanic influence is localized in the lower stratosphere, the solar influence is near the stratopause and the QBO influence is spread over the stratosphere. Because the interaction is effective in winters, phases of different forcing are close (by modulo of year, of course). For this reason, different drivers can interact constructively.

To simplify the picture we did not consider the altitude distribution of these drivers. In a more detailed model one has to take into account that he QBO propagates high in the stratosphere, where the solar-induced effects starts, volcanic aerosols act mostly in the lower stratosphere.

In the mechanism considered the solar periodic forcing produces an observable response because of the assistance of the aperiodic drivers. However situations in when the solar activity can cause a strong effect by itself are not excluded. These could happen in the periods of strongly reduced (or enhanced) solar UV flux, such as the Maunder Minimum or the Medieval Maximum periods in the solar activity.

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A. Ruzmaikin, Jet Propulsion Laboratory, California Institute of Technology, 4800 Oak Grove Dr., Pasadena, CA 91109, USA.

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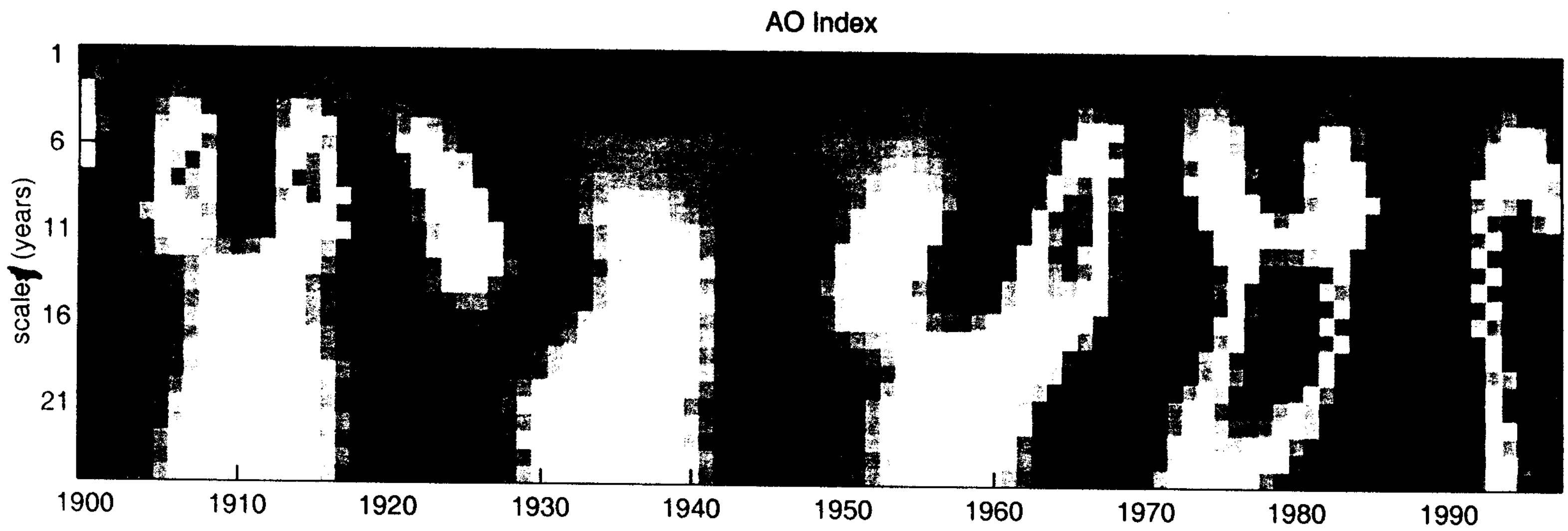
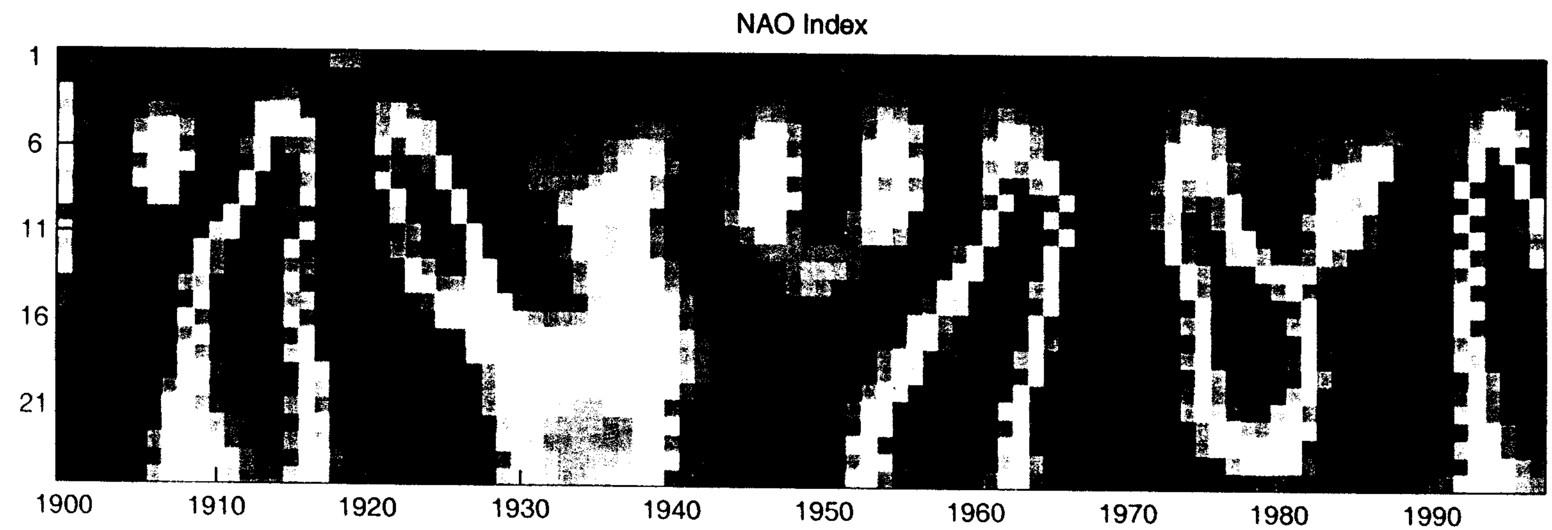
Figure Captions

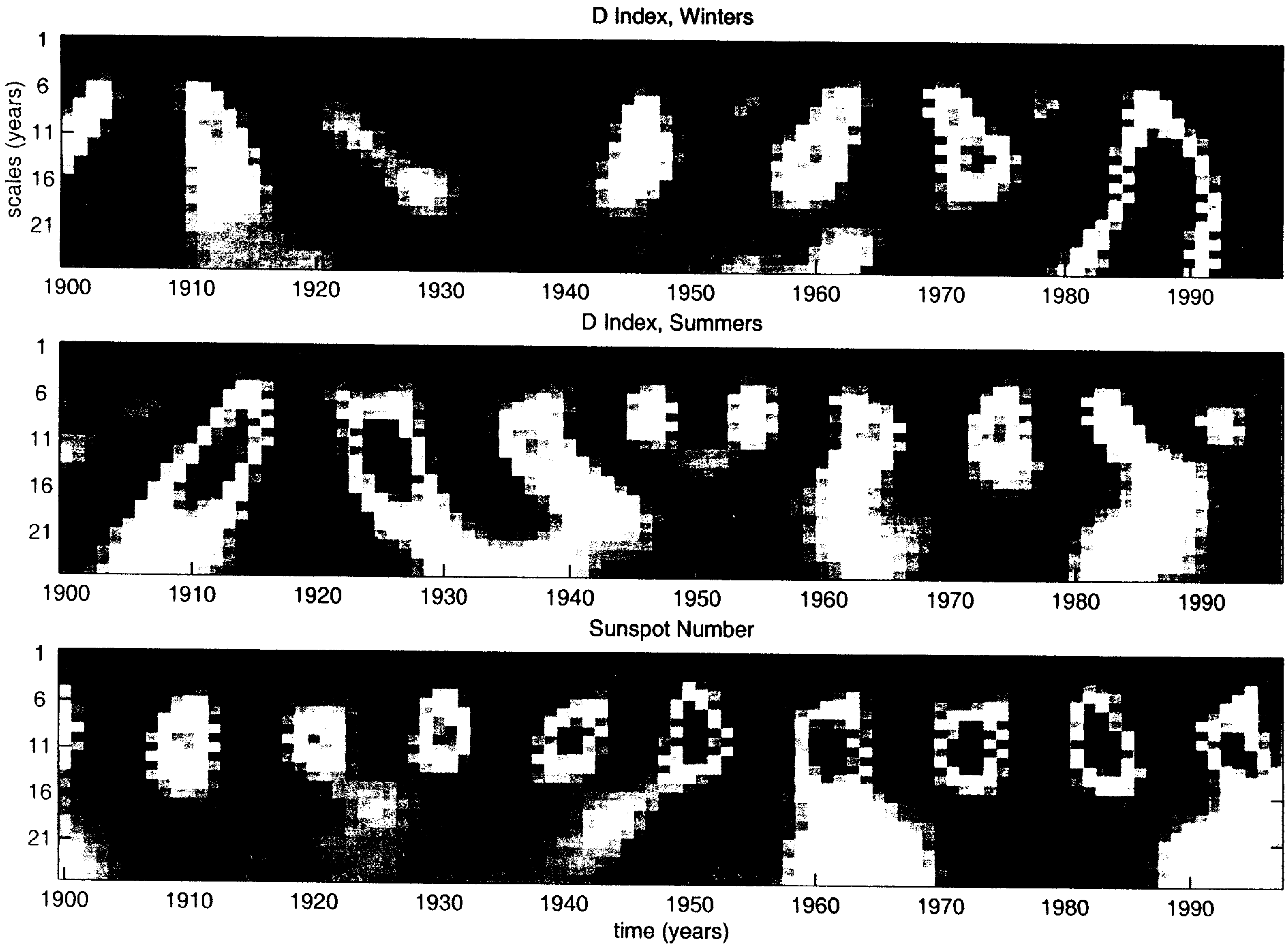
Figure 1. Wavelet spectra of the AO and NAO indices in the time interval between 1900 and 1998 (upper and middle panels). A bi-orthogonal wavelet w3.7 is used to calculate these spectra.

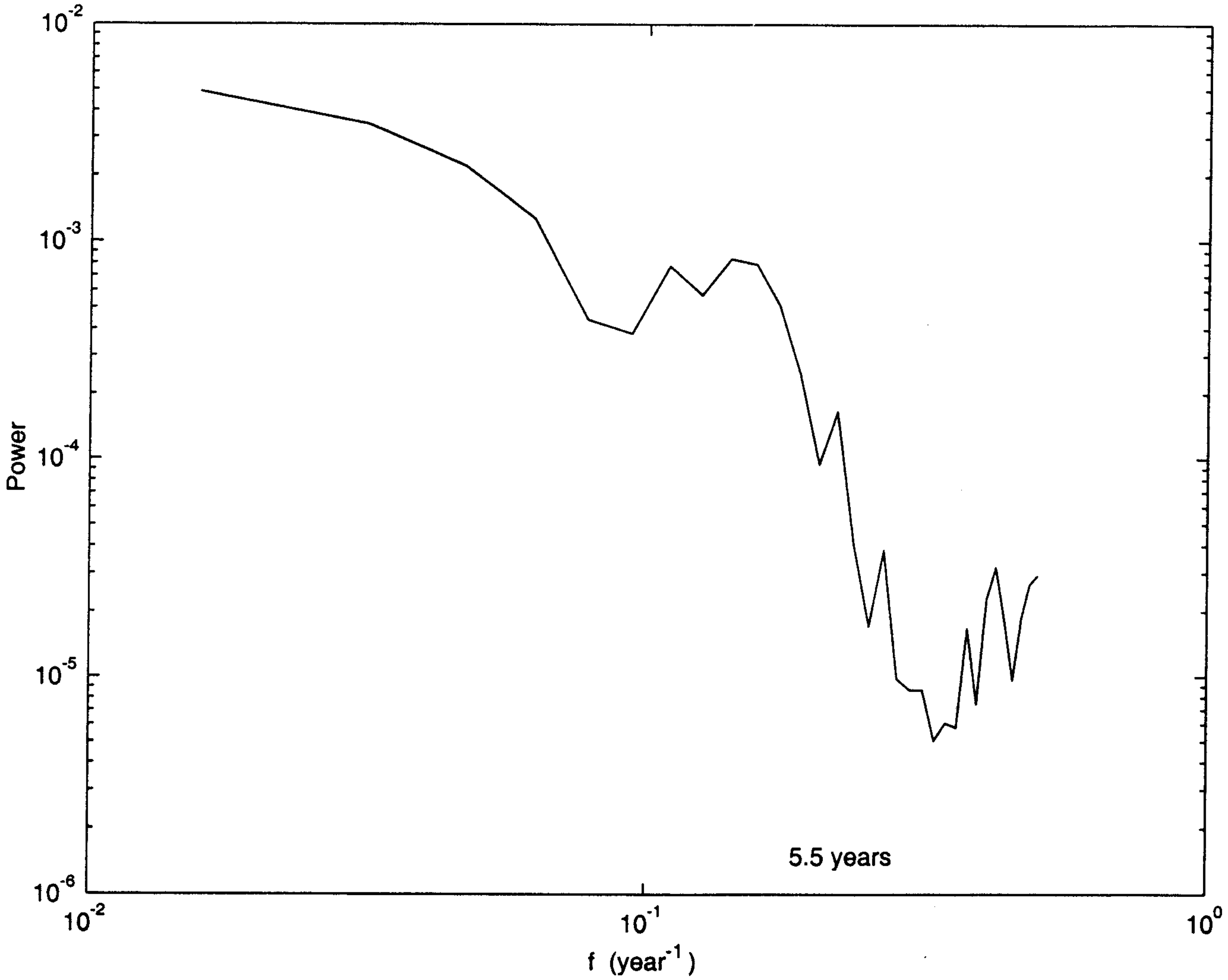
Figure 2. The wavelet spectrum of the difference index (NAO - AO) for Winter and Summer. The 11-year signal is apparent in winters. The data were pre-filtered for the range shown on the scale axis. The wavelet spectrum of the sunspots is shown for comparison (lower panel).

Figure 3. Spectrum of the optical depth in the stratosphere due to volcanic activity. The spectrum is calculated with NCEP data for the period 1850-1999.


AO Index
scale (years)
1
6
11
16
21
1900
1910
1920
1930
1940
1950
1960
1970
1980
1990


NAO Index
1
6
11
16
21
1900
1910
1920
1930
1940
1950
1960
1970
1980
1990


D Index, Winters
scales (years)
1
6
11
16
21
1900
1910
1920
1930
1940
1950
1960
1970
1980
1990
D Index, Summers
Sunspot Number
time (years)



10^{-2}
10^{-3}
10^{-4}
10^{-5}
10^{-6}
Power
5.5 years
10^{-2}
10^{-1}
10^{0}
f (year^{-1})