



ELSEVIER

Earth and Planetary Science Letters 158 (1998) 157–164

EPSL

The power spectral density of atmospheric temperature from time scales of 10^{-2} to 10^6 yr

Jon D. Pelletier*

Department of Geological Sciences, Snee Hall, Cornell University, Ithaca, New York, USA

Received 15 July 1997; accepted 18 February 1998

Abstract

After removing annual variability, power spectral analyses of local atmospheric temperature from hundreds of stations and ice core records have been carried out from time scales of 10^{-2} to 10^6 yr. A clear sequence of power-law behaviors is found as follows: (1) from 40 ka to 1 Ma a flat spectrum is observed; (2) from 2 ka to 40 ka the spectrum is proportional to f^{-2} where f is the frequency; and (3) below time scales of 2 ka the power spectrum is proportional to $f^{-1/2}$. At time scales less than 1 month we observe that the power spectra of continental stations become proportional to $f^{-3/2}$ while maritime stations continue to have power spectra proportional to $f^{-1/2}$ down to time scales of 1 day. To explain these observations, we model the vertical transport of heat in the atmosphere as a stochastic diffusion process. The power spectrum of temperature fluctuations at the earth's surface expected from this model equation in a two-layer geometry with thermal and eddy diffusion properties appropriate to the atmosphere and the ocean and a radiation condition at the top of the atmosphere agrees with the observed spectrum. The difference in power spectra between continental and marine stations can be understood with this approach as a consequence of the air mass above a maritime station exchanging heat with both the atmosphere above and the ocean below while a continental station exchanges heat mostly with the atmosphere above. © 1998 Elsevier Science B.V. All rights reserved.

Keywords: spectral analysis; atmosphere; temperature; heat transfer

1. Introduction

Understanding the natural variability of climate is one of the most important tasks facing climatologists. The Intergovernmental Panel on Climate Change [6] concluded that the “balance of evidence suggests a discernible human impact on the climate system”. This conclusion is based, however, on comparison with the natural variability of the climate sys-

tem expected from general circulation model (GCM) control runs. GCM control runs often exhibit significantly lower variability, by a factor of 2 to 5, and a different frequency dependence than paleoclimatic data [6]. Other model results exhibit natural variability comparable in magnitude to that observed in the last 100 years [1]. Understanding more about the natural variability of climate is essential for an accurate assessment of the human influence on climate. For example, an accurate model of natural variability would enable climatologists to make quantitative estimates of the likelihood that the observed warming trend is anthropogenically induced. One such

* Present address: Caltech, Mail Code 150-21, Pasadena, CA 91125, USA. Fax: +1 (626) 585-1917; E-mail: jon@gps.caltech.edu

analysis has been performed by Galbraith and Green [3], who rejected the possibility of a stochastic trend giving rise to the observed warming over the past century. However, their stochastic model for the natural variability of climate was an autoregressive model which had an exponential autocorrelation dependence on time lag. We present evidence for a power-law autocorrelation function, implying larger low-frequency fluctuations than those produced by an autoregressive stochastic model. This evidence suggests that the statistical likelihood of the observed warming trend being larger than that expected from natural variations of the climate system must be reexamined.

In this paper we consider the power spectral density or power spectrum of temporal variations in atmospheric temperature on time scales of 10^{-2} to 10^6 yr. We will show that at frequencies smaller than $f \approx 1/(40 \text{ ka})$ the power spectrum is flat (white noise). At frequencies between $f \approx 1/(40 \text{ ka})$ and $f \approx 1/(2 \text{ ka})$ the power spectrum is proportional to f^{-2} (a Brownian walk). At frequencies greater than $f \approx 1/(2 \text{ ka})$ the power spectrum is proportional to $f^{-1/2}$. At very high frequencies (above $f \approx 1/(1 \text{ month})$) the spectrum varies as $f^{-3/2}$ for continental stations and remains proportional to $f^{-1/2}$ for maritime stations.

We will further show that the observed power spectrum of atmospheric temperature is identical to the power spectrum of variations due to the stochastic diffusion of heat in a metallic film that is in thermal equilibrium with a substrate [17]. Temperature variations in the film and substrate occur as a result of fluctuations in the heat transport by electrons undergoing Brownian motion. The top of the film absorbs and emits blackbody radiation. In our analogy we associate the atmosphere with the metallic film and the oceans with the substrate. Turbulent eddies in the atmosphere and oceans are analogous to the electrons undergoing Brownian motion in a metallic film in contact with a substrate.

2. Power-spectral analysis of atmospheric temperature variations

We first consider the spectral behavior of the deuterium concentrations in the Vostok (East Antarctica)

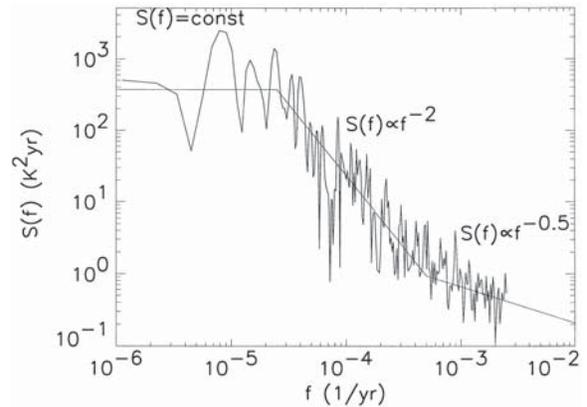


Fig. 1. Lomb periodogram of the temperature inferred from the Vostok ice core. The power spectrum, S , is given as a function of frequency for time scales of 500 yr to 1 Ma.

ice core. A 220 ka record of temperature fluctuations is obtained using the conversion $5.6 \delta D(\%) = 1 \text{ K}$ [8]. Jouzel and Merlivat [9] have concluded that the Vostok deuterium record is a proxy for local atmospheric temperature. Because the data are unevenly sampled we utilized the Lomb periodogram [15] to estimate the power spectrum. The results are given in Fig. 1. We associate the power spectrum with three regions of different spectral behavior. The first region, at frequencies less than $f \approx 1/(40 \text{ ka})$, is a white noise ($S(f) = \text{constant}$). The second region, between $f \approx 1/(40 \text{ ka})$ and $f \approx 1/(2 \text{ ka})$ is a Brownian walk ($S(f) \propto f^{-2}$). In the third region, with frequencies greater than $f \approx 1/(2 \text{ ka})$, there is a change to a flatter spectrum. This change is associated with rapid variations in the Vostok core. This flattening of the spectrum above a frequency of $f \approx 1/(2 \text{ ka})$ is also observed in ice cores from Greenland [20]. Several authors have reported the low-frequency behavior (a flat spectrum crossing over to f^{-2} behavior) with other paleoclimatological data [5,10]. The straight-line segments included in the graph represent a power spectrum whose amplitude and crossover frequencies are estimated by visually minimizing the variance of the periodogram around the estimated power spectrum.

In order to extend our analyses to higher frequencies we have carried out power-spectral analyses on data for atmospheric temperature variations from weather stations. One of the longest available records is the average monthly temperature in cen-

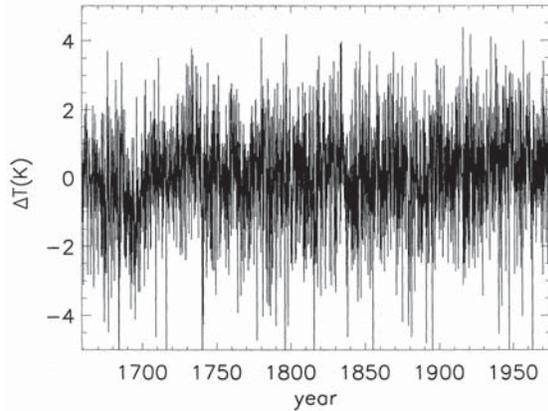


Fig. 2. Average monthly atmospheric temperature for central England [11] with the yearly periodicity removed.

tral England, 1659–1973. The data are tabulated in Manley [11]. The yearly periodicity was removed from these data by subtracting from each value the average temperature of that month for the entire record. The resulting time series is given in Fig. 2. The time series exhibits rapid fluctuations from year to year superimposed on more gradual, lower-frequency variations. The power spectrum estimated as the square of the coefficients of the fast Fourier transform (FFT) with a Hanning taper is presented in Fig. 3 along with a least-square power-law fit to the data with $S(f) \propto f^{-0.47}$. We have also determined the average power spectrum of the time series of monthly mean temperatures from 94 stations worldwide with the yearly trend removed. We obtained

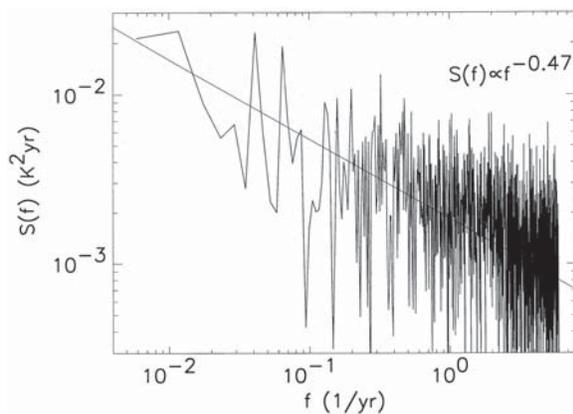


Fig. 3. Power spectrum of the time series of central England temperatures.

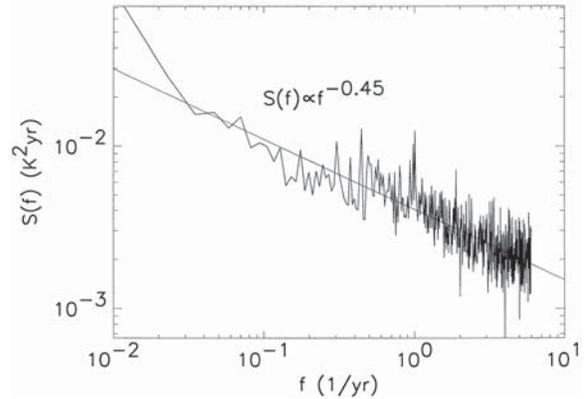


Fig. 4. Average power spectrum of 94 complete monthly temperature time series from the data set of Vose et al. [18] plotted as a function of frequency in yr^{-1} . The power spectrum, S , is given as a function of frequency for time scales of 2 months to 100 yr.

the power spectra $S(f)$ of all complete temperature series of lengths greater than or equal to 1024 months from the climatological data base compiled by Vose et al. [18]. The yearly trend was removed by subtracting from each monthly data point the average temperature for that month in the 86 yr record for each station. All of the power spectra were then averaged at equal frequency values. The results are given in Fig. 4. The data yield a straight-line on a log-log plot with slope close to -0.5 indicating that $S(f) \propto f^{-1/2}$ in this frequency range.

Finally we consider the average power spectrum of time series of daily mean temperature (estimated by taking the average of the maximum and minimum temperature of each day) from 370 continental and 91 maritime stations over 4096 days. Maritime stations are sites on small islands far from any large land masses. Continental stations are well inland on large continents, far from any large bodies of water. We chose all stations from the complete records (those with greater than 4096 nearly consecutive days of data) provided by the Global Daily Summary database compiled by the National Climatic Data Center [12]. The yearly periodicity in these data were removed by subtracting from each station's temperature time series a converged least-squares fit to a function of the form:

$$T_i = T_{\text{av}} + A \cos\left(\frac{2\pi}{365}i + \phi\right) \quad (1)$$

where i is the number of the day in the year, and T_{av} ,

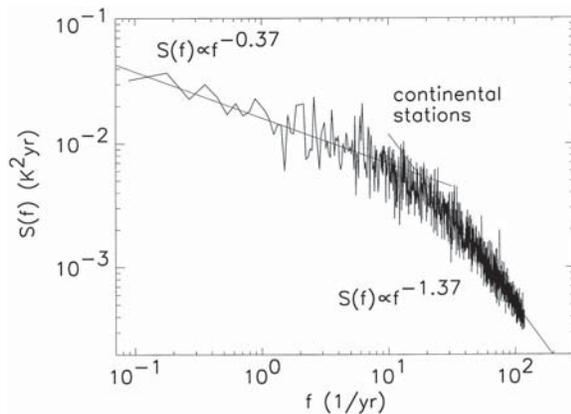


Fig. 5. Averaged power spectrum of 370 continental daily temperature time series from the data set of the National Climatic Data Center [12] as a function of frequency in yr^{-1} . The power spectrum, S , is given as a function of frequency for time scales of 2 days to 10 yr. The crossover frequency is $f = 1/\text{month}$.

A , and ϕ were the fitting parameters. This procedure is a standard one for subtracting the yearly periodicity in a meteorological time series [7]. The results are given in Figs. 5 and 6. Continental stations (Fig. 5) correlate with a $f^{-3/2}$ high-frequency region. Maritime stations (Fig. 6) correlate with a $f^{-1/2}$ scaling up to the highest frequency. The crossover frequency for the continental spectra is $f \approx 1/(\text{1 month})$. The difference between continental and maritime stations results from the air mass above maritime stations

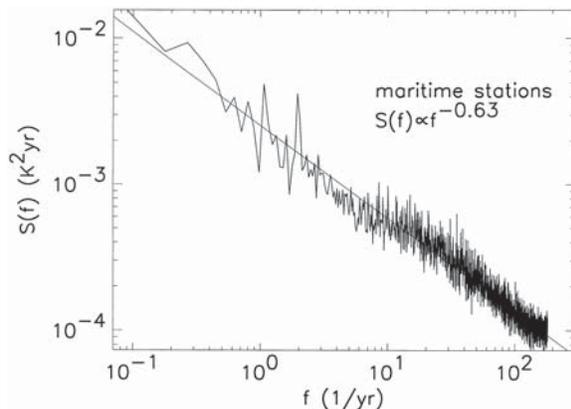


Fig. 6. Averaged power spectrum of 91 maritime daily temperature time series from the data set of the National Climatic Data Center [12] as a function of frequency in yr^{-1} . The power spectrum, S , is given as a function of frequency for time scales of 2 days to 10 yr.

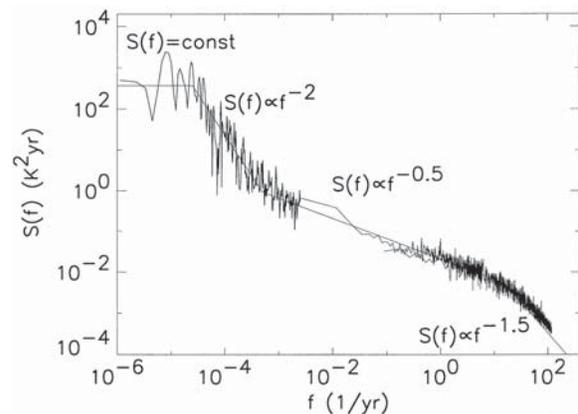


Fig. 7. Composite power spectrum of local atmospheric temperature from time scales from instrumental data and inferred from ice cores from time scales of 10^{-2} to 10^6 yr. The high-frequency data are for continental stations. Piece-wise power-law trends are indicated.

exchanging heat with both the atmosphere above and the oceans below while the air mass above continental stations exchanges heat only with the atmosphere above it. The three spectra have been combined in Fig. 7 to give a continuous spectral behavior of local atmospheric temperature from time scales of 10^{-2} to 10^6 yr.

3. Model

We can interpret these results in terms of the vertical turbulent transport of heat energy in the atmosphere, its radiation of energy to space, and its exchange of heat with the ocean. In our model, vertical turbulent transport is modeled as a stochastic diffusion process. If the convection of heat energy in the atmosphere is dominated by eddies small compared to the scale height of the atmosphere, the dispersion of heat energy by turbulent eddies is analogous to the diffusion that results from the random motion of fluids at the molecular scale. Due to the stochastic nature of turbulent flow, heat energy is dispersed a different way when placed in a turbulent flow at different times even if the mean concentration is described by the diffusion equation. Therefore, it is appropriate to model turbulent transport by the diffusion equation with a random noise in the flux

term:

$$\rho c \frac{\partial \Delta T}{\partial t} = -\frac{\partial J}{\partial x} \quad (2)$$

$$J = -\sigma \frac{\partial \Delta T}{\partial x} + \eta(x, t) \quad (3)$$

where J is the heat flux, ΔT is the fluctuation in temperature from equilibrium, ρ is the density, c is the heat capacity per unit mass, σ is the thermal conductivity, and η is a Gaussian white noise in space and time. Eq. 2 is conservation of energy. Eq. 3 is Fourier's law of heat transport with random advection of heat superimposed. The noise term is chosen to be Gaussian white noise because velocity fluctuations in atmospheric turbulence are observed to be Gaussian and white above a time scale of minutes. We will compute the power spectrum of temperature fluctuations in a layer of width $2l$. We consider an infinite space in which Eqs. 2 and 3 are applicable. We focus our attention on a layer of width $2l$ in this space. The presentation we give is similar to that of Voss and Clarke [19]. The variations in total heat energy in the layer of width $2l$ is determined by the heat flow across the boundaries. Fig. 8a illustrates the geometry of the layer exchanging thermal energy

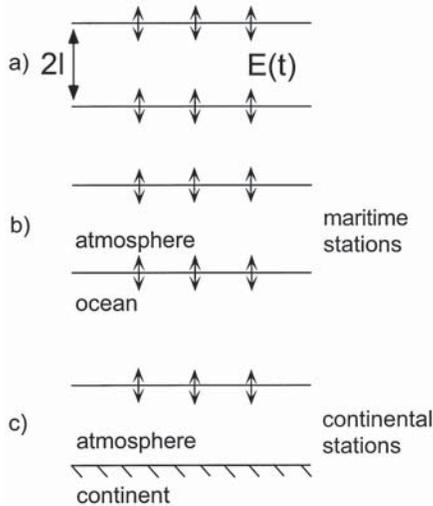


Fig. 8. (a) Geometry of the one-dimensional diffusion calculation detailed in the text. (b) Boundary conditions appropriate to the air masses above the ocean (maritime stations), where the ocean acts as a thermal conductor. (c) Boundary conditions appropriate to the air masses above the continents (continental stations), where the continents act as a thermal insulator.

with diffusing regions above and below it. A diffusion process has a frequency-dependent correlation length $\lambda = (2D/f)^{1/2}$ [19]. Two different situations arise as a consequence of the length scale, $2l$, of the geometry. For high frequencies $\lambda \ll 2l$ the fluctuations in heat flow across the two boundaries are independent. For low frequencies $\lambda \gg 2l$ and the fluctuations in heat across the two boundaries are in phase. First we consider high frequencies. Since the boundaries fluctuate independently, we can consider the flow across one boundary only. The flux of heat energy is given by Eq. 3. Its Fourier transform is given by:

$$J(k, \omega) = \frac{i\omega\eta(k, \omega)}{\alpha k^2 + i\omega} \quad (4)$$

where $f = 2\pi\omega$ and $\alpha = \sigma/(\rho c)$ is the diffusivity. The flux of heat energy out of the layer at the boundary at $x = l$ (the other boundary is located at $x = -l$) is the rate of change of the total energy in the layer $E(t)$: $dE(t)/dt = J(l, t)$. The Fourier transform of this equation is:

$$E(\omega) = \frac{i}{(2\pi)^{1/2}\omega} \int_{-\infty}^{\infty} dk e^{ikl} J(k, \omega) \quad (5)$$

Therefore, the power spectrum of variations in $E(t)$, $S_E(\omega) = \langle |E(\omega)|^2 \rangle$, is:

$$S_E(\omega) \propto \int_{-\infty}^{\infty} \frac{dk}{\alpha^2 k^4 + \omega^2} \propto \omega^{-3/2} \quad (6)$$

In the above expression, the noise term η does not appear because, since it is white noise in space and time, its average amplitude is independent of ω and k , i.e. it is just a constant. Since $\Delta T \propto \Delta E$ the power spectrum of temperature has the same form as S_E and $S_T(\omega) \propto \omega^{-3/2}$.

If we include the heat flux out of both boundaries, the rate of change of energy in the layer will be given by the difference in heat flux: $dE(t)/dt = J(l, t) - J(-l, t)$. The Fourier transform of $E(t)$ is now:

$$E(\omega) = \frac{1}{(2\pi)^{1/2}\omega} \int_{-\infty}^{\infty} dk \sin(kl) J(k, \omega) \quad (7)$$

Then,

$$S_T(\omega) \propto S_E(\omega) \propto \int_{-\infty}^{\infty} \frac{dk \sin^2(kl)}{\alpha^2 k^4 + \omega^2} \propto \omega^{-3/2} [1 - e^{-\theta} (\sin \theta + \cos \theta)] \quad (8)$$

where $\theta = (\omega/\omega_0)^{1/2}$ and $\omega_0 = \alpha/2l^2$ is the frequency where the correlation length is equal to the width of the layer. When $\lambda \ll 2l$, the above expression reduces to $S_T(f) \propto f^{-3/2}$. When $\lambda \gg 2l$, $S_T(f) \propto f^{-1/2}$ [19]. In this limit the boundaries fluctuate in phase, and heat that enters into the region from one boundary can diffuse out of the other boundary. The result is a sequence of fluctuations which are less persistent than the single boundary $f^{-3/2}$ case.

The difference in the spectral behavior between continental and maritime stations at high frequencies can be interpreted in terms of the diffusion model presented above. The power spectrum of temperature variations in an air mass exchanging heat by one-dimensional stochastic diffusion is proportional to $f^{-1/2}$ if the air mass is bounded by two diffusing regions and is proportional to $f^{-3/2}$ if it interacts with only one. The boundary conditions appropriate to maritime and continental stations are presented in Fig. 8b and c, respectively. The layer considered is taken to have an upper boundary embedded in the atmosphere and a lower boundary at the earth's surface. For maritime stations heat is transferred across this lower boundary into the oceans so it is equivalent to the case $\lambda > 2l$ and therefore the power spectrum of temperature variations is $S(f) \propto f^{-1/2}$. For continental stations the lower boundary is insulating (conducts very little heat compared to the atmosphere and the ocean) so it is equivalent to the case $\lambda < 2l$ and therefore $S(f) \propto f^{-3/2}$. At low frequencies, horizontal heat exchange between continental and maritime air masses limits the variance of the continental stations. This crossover should occur at the time scale when the air masses above continents and oceans become mixed. The time scale for one complete Hadley or Walker circulation which mixes the air masses is approximately 1 month, the same time scale as the observed crossover. Details are presented in Pelletier [13].

The same sequence of power-law behaviors from 10^{-2} to 10^6 yr is also observed in the spectrum of temperature fluctuations in a metallic film in thermal contact with a substrate. Van Vliet et al. [17] studied the temperature fluctuations in a two-layer system with heat transport by electrons given by Eqs. 2 and 3. The boundary conditions were zero heat flux out of the bottom layer, continuity of temperature and

heat flux at interface between the two layers, and a radiation boundary condition at the top of the metallic film. This geometry and boundary conditions may also be applicable to the coupled ocean–atmosphere case. Pelletier [13] applied this model to the climate system. He showed that at low frequencies the spectrum of temperature variations in the model exhibits the flat spectrum observed at time scales greater than 40 ka and a f^{-2} spectrum for time scales between 2 ka and 40 ka. For time scales greater than 1 month but less than 2 ka, fluctuations in outgoing heat energy from the atmosphere by radiative cooling causes temperature variations in the atmosphere which can be damped by the oceans. At frequencies lower than 2 ka, the time scale of vertical ocean mixing, the atmosphere and oceans are in thermal equilibrium. The oceans can no longer absorb thermal fluctuations in the atmosphere resulting from fluctuations in radiative emission at this time scale. This is consistent with the results of Bender et al. [2], who identified 2 ka as the time scale for global thermal equilibration based on a comparison of isotope records from Greenland and Antarctica. The variance in temperature of the atmosphere and oceans is then determined solely by the radiation boundary condition. The fluctuating temperature at the cloud layer will result in a white noise flux out of the atmosphere–ocean system. The average temperature of the atmosphere and oceans at these time scales will be given by the sum of a white noise, a Brownian walk. This is observed in the Vostok data between time scales of 2 ka and 40 ka.

The power spectrum of temperature variations flattens out at frequencies lower than $f \approx 1/(40 \text{ ka})$ as a result of a negative feedback mechanism: as the coupled atmosphere and oceans warm up (or cool down) due to nonstationary fluctuations resulting from the random heat flux out into space, the system will radiate, on average, more (or less) radiation, limiting the variance at low frequencies. This can be described by a linear damping equation for the global temperature difference from equilibrium:

$$\frac{\partial \Delta T}{\partial t} = -\frac{1}{\tau_0} \Delta T + \eta(t) \quad (9)$$

where τ_0 is the radiative time scale of the coupled

earth system, given by [13]:

$$\tau_o = \frac{w_1 c \rho + w_2 c' \rho' (1 + g w_1 / \sigma)}{g} \quad (10)$$

where c' is the heat capacity per unit mass of the oceans, w_2 is the mean depth of the oceans, g is the thermal emissivity of the atmosphere, w_1 is the scale height of the atmosphere, ρ' is the density of the oceans, and σ is the vertical conductivity of the atmosphere by turbulent transport. We will use the value $g = 1.7 \text{ W m}^{-2} \text{ K}^{-1}$ as used by Harvey and Schneider [4].

The temperature variations ΔT from this equation have a spectrum which is a Lorentzian with a crossover time scale of τ_o . This can be shown with Fourier transforms. The Fourier transform of Eq. 9 is given by:

$$\Delta T(\omega) = \frac{\eta(\omega)}{\tau_o^{-1} + i\omega} \quad (11)$$

The power spectrum $S_{\Delta T}(\omega) = \langle |\Delta T(\omega)|^2 \rangle$ is then given by:

$$S_{\Delta T}(\omega) = \frac{1}{f_o^2 + f^2} \quad (12)$$

Now, we must consider whether the observed low-frequency crossover time scale of 40 ka is consistent with the model prediction Eq. 10. If we neglect the heat capacity of the atmosphere relative to that of the ocean, Eq. 10 reduces to:

$$\tau_o = \frac{c' \rho' w_2}{g} + \frac{w_1 w_2 c' \rho'}{\sigma} \quad (13)$$

The first term is the time scale for radiative damping of the heat energy of the coupled atmosphere–ocean system into space from clouds. The second term is the time scale for transport of the heat energy of the ocean through the atmosphere to the cloud layer. If the time scale for one of these processes is much larger than the time scale for the other, the crossover time scale will be determined by that rate-limiting step. For the earth's climate system, the transport of the oceans' heat through the atmosphere appears to be the rate-limiting step. This process takes a long time because the atmosphere has a low heat capacity compared to the oceans and is therefore a poor heat conductor. The time scale of radiative damping is estimated to be 600 yr using an ocean depth of 4 km,

$g = 1.7 \text{ W m}^{-2} \text{ K}^{-1}$, and the density and heat capacity of water. The time scale for vertical transport of the oceans' heat through the atmosphere can only be estimated to within an order of magnitude since this time scale is linearly dependent on the average vertical diffusivity of the atmosphere. Only rough estimates are available for this parameter. Estimates of $1 \text{ m}^2/\text{s}$ for this parameter have been given by Pleune [14] and Seinfeld [16]. In order for the time scale of vertical advection of the oceans' heat through the atmosphere to be 40 ka, a vertical diffusivity of $0.1 \text{ m}^2/\text{s}$ is required, a factor of 10 lower but roughly in agreement with the values quoted above.

4. Conclusion

We have presented evidence that the power spectrum of atmospheric temperatures exhibits four regions of power-law behavior characterized by different power-spectral exponents. We presented a model of van Vliet et al. [17] used to study temperature fluctuations in a metallic film (analogous to the atmosphere) supported by a substrate (analogous to the ocean) that matches the observed frequency dependence of the power spectrum of temperature fluctuations. The difference between the high-frequency spectra of continental and maritime stations may be interpreted in terms of the fact that air masses above maritime stations interact with the atmosphere above and the ocean below while air masses above continental stations interact with only the atmosphere above. [CL]

References

- [1] T.P. Barnett, A.D. Del Genio, R.A. Ruedy, Unforced decadal fluctuations in a coupled model of the atmosphere and ocean mixed layer, *J. Geophys. Res.* 97 (1992) 7341–7354.
- [2] M. Bender, T. Sowers, M.-L. Dickson, J. Orchard, P. Grootes, P.A. Mayewski, D.A. Meese, Climate correlations between Greenland and Antarctica during the past 100,000 years, *Nature* 372 (1994) 663–666.
- [3] J.W. Galbraith, C. Green, Inference about trends in global temperature data, *Climatic Change* 22 (1992) 209–221.
- [4] L.D.D. Harvey, S.H. Schneider, Transient climate response to external forcing on 10^0 – 10^4 year time scales, I. Experiments with globally averaged, coupled, atmosphere and

- ocean energy balance models, *J. Geophys. Res.* 90 (1985) 2191–2205.
- [5] K. Hasselmann, Stochastic climate modeling, I. Theory, *Tellus* 6 (1971) 473–485.
- [6] Intergovernmental Panel on Climate Change, Climate Change, The IPCC Scientific Assessment, J.T. Houghton, B.A. Callendar (Eds.), Cambridge Univ. Press, New York, 1995.
- [7] I.M. Janosi, G. Vattay, Soft turbulent state of the atmospheric boundary layer, *Phys. Rev. A* 46 (1992) 6386–6389.
- [8] J. Jouzel, C. Lorius, J.R. Petit, C. Genthon, N.I. Barkov, V.M. Kotlyakov, Vostok ice-core: A continuous isotope temperature record over the last climatic cycle (160,000 years), *Nature* 329 (1987) 403–407.
- [9] J. Jouzel, D. Merlivat, Deuterium and oxygen 18 in precipitation: modeling of the isotopic effects during snow formation, *J. Geophys. Res.* 89 (1984) 11749–11757.
- [10] M.A. Komintz, N.G. Piasias, Pleistocene climate: deterministic or stochastic?, *Science* 204 (1979) 171–172.
- [11] G. Manley, Central England temperatures: monthly means 1659–1973, *Q. J. R. Meteorol. Soc.* 100 (1974) 389–405.
- [12] National Climatic Data Center, Global Daily Summary: Temperature and Precipitation 1977–1991, Version 1.0, 1994.
- [13] J.D. Pelletier, Analysis and modeling of the natural variability of climate, *J. Climate* 10 (1997) 1331–1342.
- [14] R. Pleune, Vertical diffusion in the stable atmosphere, *Atm. Environ. A* 24 (1990) 2547–2555.
- [15] W.H. Press, S.A. Teukolsky, W.T. Vetterling, B.P. Flannery, *Numerical Recipes in C: The Art of Scientific Computing*, 2nd ed., Cambridge Univ. Press, New York, 1992.
- [16] J.H. Seinfeld, *Atmospheric Chemistry and Physics of Air Pollution*, Wiley, New York, 1986.
- [17] K.M. Van Vliet, A. van der Ziel, R.R. Schmidt, Temperature-fluctuation noise of thin films supported by a substrate, *J. Appl. Phys.* 51 (1980) 2947–2956.
- [18] R.S. Vose, R.L. Schmoyer, P.M. Stewer, T.C. Peterson, R. Heim, T.R. Karl, J.K. Eischeid, The global historical climatology network: long-term monthly temperature, precipitation, sea-level pressure, and station pressure data, *Environ. Sci. Div. Publ.* 392, Oak Ridge National Laboratory, 1992.
- [19] R.F. Voss, J. Clarke, Flicker (1/f) noise: equilibrium temperature and resistance fluctuations, *Phys. Rev. B* 13 (1976) 556–573.
- [20] P. Yiou, J. Jouzel, S. Johnsen, O.E. Rognvaldsson, Rapid oscillations in Vostok and GRIP ice cores, *Geophys. Res. Lett.* 22 (1995) 2179–2182.