Climate Fluctuations and Climate Sensitivity

Gerald R. North and Kuor-Jier Joseph Yip
Climate System Research Program
Texas A&M University
College Station, TX 77843
Telephone: 409-845-8083

May 18, 1990

Abstract

Some evidence is presented that the main part of the atmospheric climate system is such that small forcings in the heat balance lead to linear responses in the surface temperature field. By examining first a noise forced energy-balance climate model and then comparing with a long run of a highly symmetrical general circulation model, one finds a remarkable connection between spatial autocorrelation statistics and the thermal influence function for a point heat source. These findings are brought together to indicate that this particular climatological field may be largely governed by linear processes.

1 Introduction

The sensitivity of the ensemble averaged (climatological) surface temperature field to changes in the outside forcing such as the solar insolation or carbon dioxide loading of the atmosphere is a topic of obvious importance. Similarly, one wants to know as much as possible about the temporal delays in such responses. In this paper it is argued that certain aspects of the climate response, mainly the surface temperature, are controlled by effectively linear processes. By this I mean that say a 2% change in solar constant will lead to twice the response as a 1% change. In the following some evidence will be produced indicating that this is so and in turn how a study of the fluctuations in the system can shed some light on the sensititivity issue.

First, consider why such an approach is important. The controversy over the magnitude of global warming to be expected in the next 60 years is well known. Climate models differ markedly in their projections for conditions with an effectively doubled CO_2 concentration in the atmosphere. The models give a range of from 2°C to about 5°C for their responses. Much of the uncertainty seems to be traceable to the differing ways the models include the effects of clouds (Cess et al., 1989). When we discuss the transient climate problem, i.e., the response of the system to a gradual (say 1% per year) increase in the effective CO_2 concentration we have the question of how the slower parts of the system (mainly oceans) share and effectively sequester the heat away from the surface and thereby cause a substantial lag in the surface response. The problem of cloud feedback and oceanic lag are the first order problems of forced climatic change; after that we must begin to look at the way the general circulation of the atmosphere and the oceans might be distorted by small changes in the forcing. These may be amplified in subtle ways by how the system disposes of latent heat through precipitation. Precipitation is notoriously poorly simulated by today's general circulation models (GCMs).

The above problems illustrate the need to occasionally step back and look at the bigger picture in search of general principles that can help us in cross-checking our results. Getting a better feeling for the climate sensitivity has enormous practical implications as well. The difference in policy response to a 2°C versus a 5°C scenario could be quite dramatic. It might mean the difference between an adaptive versus an interventionist strategy. Hence, it is very important to establish whether we are on the high or low end of the sensitivity range. Is there any evidence from the natural fluctuations in the system to tell us about the climate's sensitivity to external changes?

We are not the first to enter such a program since it was first opened by Leith (1975) and examined in some fluid dynamical model calculations by Bell (1980, 1985) in the context of the so-called Fluctuation Dissipation Theorem. The theorem states that in certain nonlinear fluctuating systems the sensitivity to changes in external forcing can be related to temporal autocorrelation properties. The example discussed in this paper is a trivial application of the theorem to a linear system forced by noise. Our task is to see how the all-important surface temperature field in the real world can be modeled by such a linear noised forced model. We postpone consideration of the real world for lack of data just now and concentrate on the available data set generated by a long run of a GCM with highly symmetrical boundary conditions.

In the following, some results from linear energy balance climate models (EBMs) will be reviewed along with a discussion of their sensitivity to changes in external forcing at various frequencies of forcing. Then some interesting properties of a noise forced linear EBM will be discribed as an analog of the real fluctuating climate system. Then I will discuss a long run of a GCM for an idealized planet. By examining the results of the natural fluctuations of this system and some simple experiments a preliminary assessment of the linearity question will be given.

2 Linear EBM Review

The EBM used in the present study is derived from one introduced by North, Mengel and Short (1983) and recently updated by Hyde et al., (1989). It can be completely characterized by the governing energy balance equation

$$C(\hat{\mathbf{r}})\frac{\partial}{\partial t}T(\hat{\mathbf{r}},t) - \nabla\cdot(D(\hat{\mathbf{r}},t)\nabla T(\hat{\mathbf{r}},t)) + A + BT(\hat{\mathbf{r}},t) = QS(\hat{\mathbf{r}},t)a(\hat{\mathbf{r}}) \tag{1}$$

where

$\hat{\mathbf{r}}$ = point on the sphere

t = time of year

$T(\hat{\mathbf{r}},t)$ = surface temperature at point $\hat{\mathbf{r}}$ and time t

A, B = phenomenological infrared radiation constants, from satellite data

$C(\hat{\mathbf{r}})$ = location dependent heat capacity (large over ocean:60, small over land:1)

$D(\hat{\mathbf{r}})$ =latitude (only) dependent thermal diffusion coefficient

Q=solar constant/4=340W/m^2

$S(\hat{\mathbf{r}},t)$ = normalized solar insolation at the top of the atmosphere

$a(\hat{\mathbf{r}})$=local coalbedo, here given a smooth latitude dependence from satellite data

For a discussion of the parameters and their physical interpretations see the references cited. Suffice it to say that there are a small number of adjustable parameters (three in $D(\hat{\mathbf{r}})$ and one in $C(\hat{\mathbf{r}})$) which are adjusted to obtain the best possible fit to the present climate. Since the system is linear as posed, it is convenient to develop the solution into harmonics of the annual cycle, a rapidly converging series. Maps of the modeled and observed annual harmonic and phase show a good degree of similariy. For these and other comparisons the reader is refered to Hyde et al., (1989). The semiannual harmonic is everywhere outside the polar regions less than 2°C and its pattern looks like that of the data. This is already extremely strong evidence that the system is behaving linearly at least at time scales in the frequency band around the annual cycle.

2.1 Noise Forcing

Next consider the departures $T'(\hat{\mathbf{r}}, t)$ from the equilibrium seasonal cycle by fluctuations induced by a noise forcing (see Hasselmann, 1976; North and Cahalan, 1981; Leung and North, 1990). The governing equation of the fluctuations is

$$C(\hat{\mathbf{r}})\frac{\partial}{\partial t}T'(\hat{\mathbf{r}}, t) - \nabla \cdot (D(\hat{\mathbf{r}}, t)\nabla T'(\hat{\mathbf{r}}, t)) + BT'(\hat{\mathbf{r}}, t) = F(\hat{\mathbf{r}}, t) \tag{2}$$

where $F(\hat{\mathbf{r}}, t)$ is a white noise forcing in both space and time. That is, its autocorrelations vanish unless they are taken at equal time and position

$$\langle F(\hat{\mathbf{r}}, t)F(\hat{\mathbf{r}}', t')\rangle = \sigma_F^2\delta(\hat{\mathbf{r}} - \hat{\mathbf{r}}')\delta(t - t') \tag{3}$$

The noise forcing may be thought of as a combination of imbalances due to transport related eddies or such local influences as cloudiness fluctuations. In what follows we take $D(\hat{\mathbf{r}})$ and $C(\hat{\mathbf{r}})$ to be independent of position for simplicity.

Next is a list of results that follow directly from the equations. The global average temperature is a first order Markov process with autocorrelation time $\tau_0 = C/B$. One can also show that the higher spherical harmonic mode amplitudes have characteristic times

$$\tau_{nm} = \frac{\tau_0}{(n(n+1)D/B + 1)} \tag{4}$$

which are also the decay times of the unforced system. The spatial statistics for this system are also readily computed. Consider the autocorrelation function in space for equal times (we use the plane tangent to the sphere to obtain closed form solutions with familiar functions)

$$\rho(s, \tau = 0) = K_0\left(\frac{s}{\ell}\right) \tag{5}$$

where s is the distance separating the points, K_0 is the Bessel function and $\ell^2 = D/B$ is a characteristic length in these climate models. Its interpretation is as follows: a thermal anomaly is carried away from a point by random walk a distance $\sqrt{Dt/C}$; the appropriate time t is the radiation damping time C/B; insertion gives for the characteristic distance ℓ. Another spatial autocorrelation of interest is the low frequency limit (i.e., the time series is low pass filtered before the autocorrelation in space is computed)

$$\rho(s, f \to 0) = \frac{s}{\ell}K_1\left(\frac{s}{\ell}\right) \tag{6}$$

The Green's function for a point heat source was given by North (1984). Its form is simply

$$G(s) = gK_0\left(\frac{s}{\ell}\right) \tag{7}$$

where s is the distance of the heat source to the point being examined and g is the strength of the heat source. Note that the shape of the Green's function is the same as the equal time spatial autocorrelation function. The latter shape equivalance is a general property of a large class of linear heat transport operators of which diffusion is an example. This property of linear models tells us that we can learn about the Green's function by examining certain spatial autocorrelation properties of the natural system.

Finally, we show observed data from the paper by Hansen and Lebedeff (1987) for correlations between separated stations of annual average surface temperatures for various latitude bands. Since the real earth contains oceans (5 year time constant) as well as land (one month time constant) near the stations considered, these spatial correlations represent something in between the equal time case and the low frequency case. However, we are certainly left with the impression that a length scale between 1000 and 2000 km is involved. This is in substantial agreement with the EBM calculations described above.

It is important to note that the length scale in the nfEBM is controlled by the diffusion parameter and the radiation parameter. This is different from the length scale often encountered in dynamics, the Rossby Radius of Deformation, which has nothing to do with radiation damping. We argue that the length scale in the Hansen-Lebedeff data is the climate length scale which is controlled by radiation because it comes from annual averages. One way to eliminate the RRD from the present length scale is that the RRD should have a much stronger latitude dependence ($1/\sin(latitude)$) than is found in the data.

2.2 Selected GCM Results

We have conducted a long run (15 years) of the NCAR CCM0 (R15 version) on the Texas A&M Cray Y-MP computer with the assistance of Dr. Robert Chervin of NCAR. We have saved all output from the run once per model day but are particularly interested at the moment in the surface temperature field. In addition, to facilitate comparison with simpler models and to avoid confusion with too many variables, we have simplified the boundary conditions in the CCM0 as follows: 1) The obliquity is set to zero so that the climate system is forced by perpetual equinox solar insolation. 2) All mountains have been removed so that the planet has only sea level topography. 3) Oceans and other zonal symmetry breaking features have been removed so that it is an all-land planet. The planet has no snow to change the local albedo. Statistically, every month is equivalent to every other month, all longitudes are statistically equivalent and the planet is north-south symmetric statistically. The longest time scale in the model (autocorrelation time of global

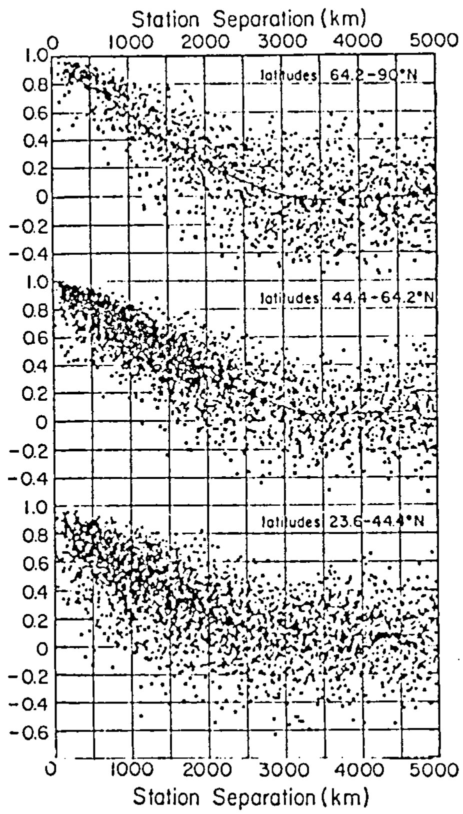

Figure 1: The spatial autocorrelation between sites separated by distance s for annually averaged temperatures. The three panels show the relationship for three different latitude bands. From Hansen and Lebedeff(1987)

average surface temperature) is about 30 days. Hence, 15 years represents $15 \times 12 = 180$ such units, an extremely long time series. By contrast the corresponding time scale when a mixed layer ocean is present is 5 years, which would require a 900 year run for equivalent statistics. Furthermore, since each longitude is statistically equivalent and the hemispheres are (statistically) reflection symmetric, the data can be pooled to have an even greater sample. While this configuration has the weakness of ignoring certain important forcing mechanisms (such as mountains and lagging heat sources such as seasonally varying ocean surfaces) it is a physically realizable planet and should yield some interesting insights into the mechanisms of climate dynamics. We refer to the planet as *Terra Blanda.*

It is important to emphasize from the beginning that the CCM0 version of the climate system (R15, noninteractive soil moisture) is far from realistic in many respects. It is nonetheless close to models being used in routine simulations today and therefore to better understand it is a worthy goal considering the policy implications of present day model simulations. The T42 (CCM2) version is now being tested for experiments by the community, and we intend to employ it in some of our later experiments to check that our conclusions are insensitive to resolution.

As part of our preliminary work with CCM0 and its simulation of surface temperatures for *Terra Blanda* we are working out the complete climatology for the planet. We report here only those aspects of the climatology relevant to the research plan to be discussed in the next section.

First consider the autocorrelation times for various spherical harmonic mode amplitudes of the CCM0 surface temperature data. We obtain these by projecting out these mode amplitudes and examining the corresponding time series. The lowest mode (0,0) has a time constant τ_0 of 29 days. The autocorrelation function is reasonably exponential in shape (in agreement with a noise-forced linear model). Higher modes also exhibit exponential autocorrelation functions with a sequence of time scales in rough agreement with those suggested by the linear model.

Next we turn to the question of spatial correlations in the CCM0 data. Consider first the equal time spatial correlations, which are shown in Fig. 2 for a point centered at 60° latitude. The horizontal scale on the map has been stretched so that distances in either direction are in locally equal length units (a small circle on the sphere should look approximately like a small circle here). The contours in the map show equal correlation lines. Note that the equal time autocorrelation length is about 13° of latitude on the map ($\approx$ 1300 km). Note first how remarkably isotropic the correlations are. Figure 3 shows the same contours for a low pass filtered ($\approx$ 60 day moving average). We see an east-west elongation of the contours and a small stretching effect not unlike the effect predicted in Fig. 1 for low-pass filtering. The anisotropy is easily identified with advection, since correlations lagged by one day show the maximum of the contours moving to the east at about 2.5 m/sec. However, the advected correlation decays very quickly with an e-folding time of its peak of only about 2.8 days. Hence, while advection is clearly present, it may be of little practical importance for the *surface temperature* in climate applications. (It is clearly the very essence of weather forecasting).

As another test of linearity in the surface temperatures generated by the CCM0 we consider the case of the response of the temperature field to a steady point (one grid site) heat source injected at the surface. The standard deviation of the surface temperature at a point is about 15° C with an autocorrelation time of about 3 days. Hence, in order to graph the spatial climatological response to a steady 400 W/m^2 heat source (local time averaged response $\approx 6.2°C$) requires several years of model generated data. After some experimentation, we exploited the symmetry in our model to combine results from a "picket fence" of point sources placed 45° apart around a latitude circle in both hemispheres. After pooling the data, the resulting contour map of responses can be found and is shown in Fig. 4. Note how nearly isotropic the map is. The distortion from isotropy is surely due to the small advection alluded to earlier. The scale of the contours is in agreement with the scale of the equal-time autocorrelations as predicted by the noise-forced EBM. Finally, we changed the strength of the heat source to 2/3 that used above. The result (not shown) was identical within sampling errors to that shown in Fig. 4 except for the scaling.

3 Conclusion

The above preliminary results suggest that forced surface temperature climate change is essentially a linear response to external forcings such as changes in the solar constant. This is remarkable considering that heat transport on the sphere is governed by extremely nonlinear interactions. We are apparently seeing the result of statistical ensemble averaging at its most beneficial. The advection of heat which is dominant for the purposes of weather forecasting is essentially indistinguishable from thermal diffusion for ensemble averaged climate. The linearity has shown up in the spatial and temporal autocorrelation statistics as well as in the scaling of point-heat-source strength/response characteristics. Of course, our analysis leaves out the rather obvious nonlinearity associated with snow-albedo feedback and other simplifications which could also induce nonlinear response such as mountain wave locking are similarly omitted. The main importance of the results is that the sensitivity of the real climate system may be relatable to the natural fluctuation statistics. This could be a key element in a strategy for deciding between competing models whose sensitivities are grossly different.

References

◇ Bell, T. L., 1985: Climatic sensitivity and fluctuation-dissipation. in Turbulence and Predictability in Geophysical Fluid Dynamics and Climate Dynamics. Corso. Soc. Italiana di

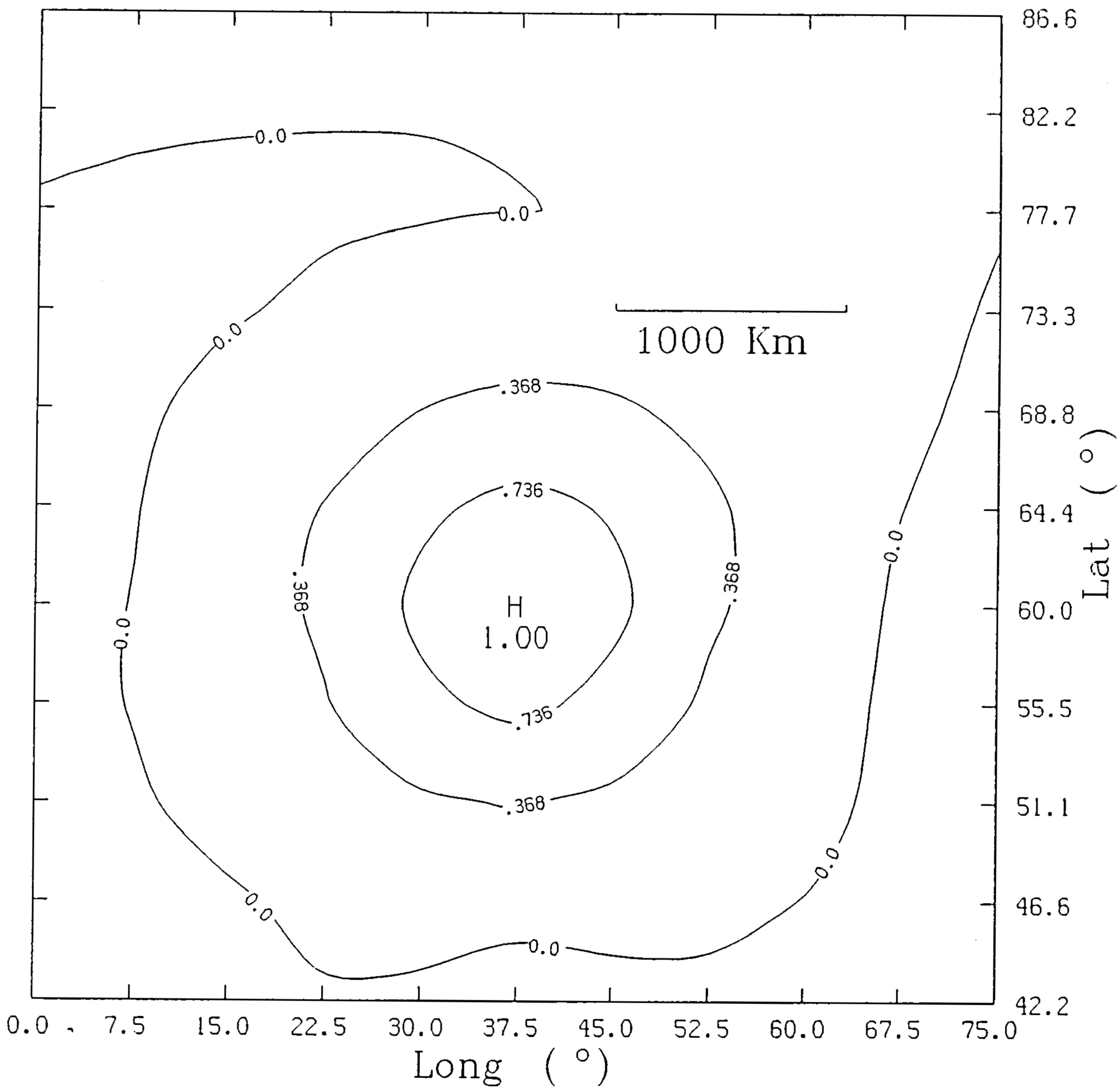

Figure 2: Equal time spatial correlations with the surface temperature at a point located at 60.0° latitude for *Terra Blanda* as simulated in the CCM0. Horizontal and vertical scales are approximately equal for linear distances.

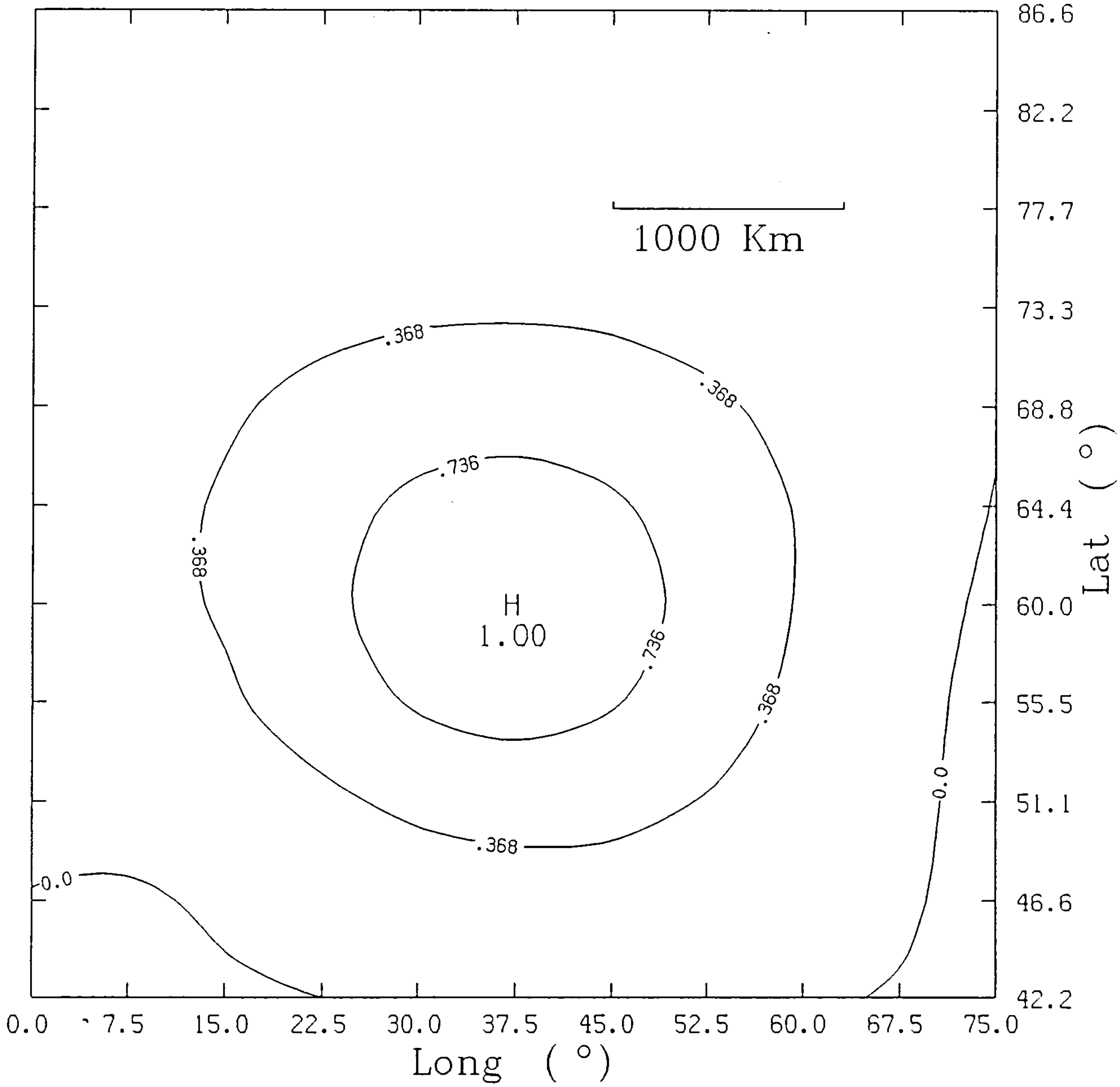

Figure 3: Same as Fig. 2 except that the time series has been low pass filtered (61 day moving average).

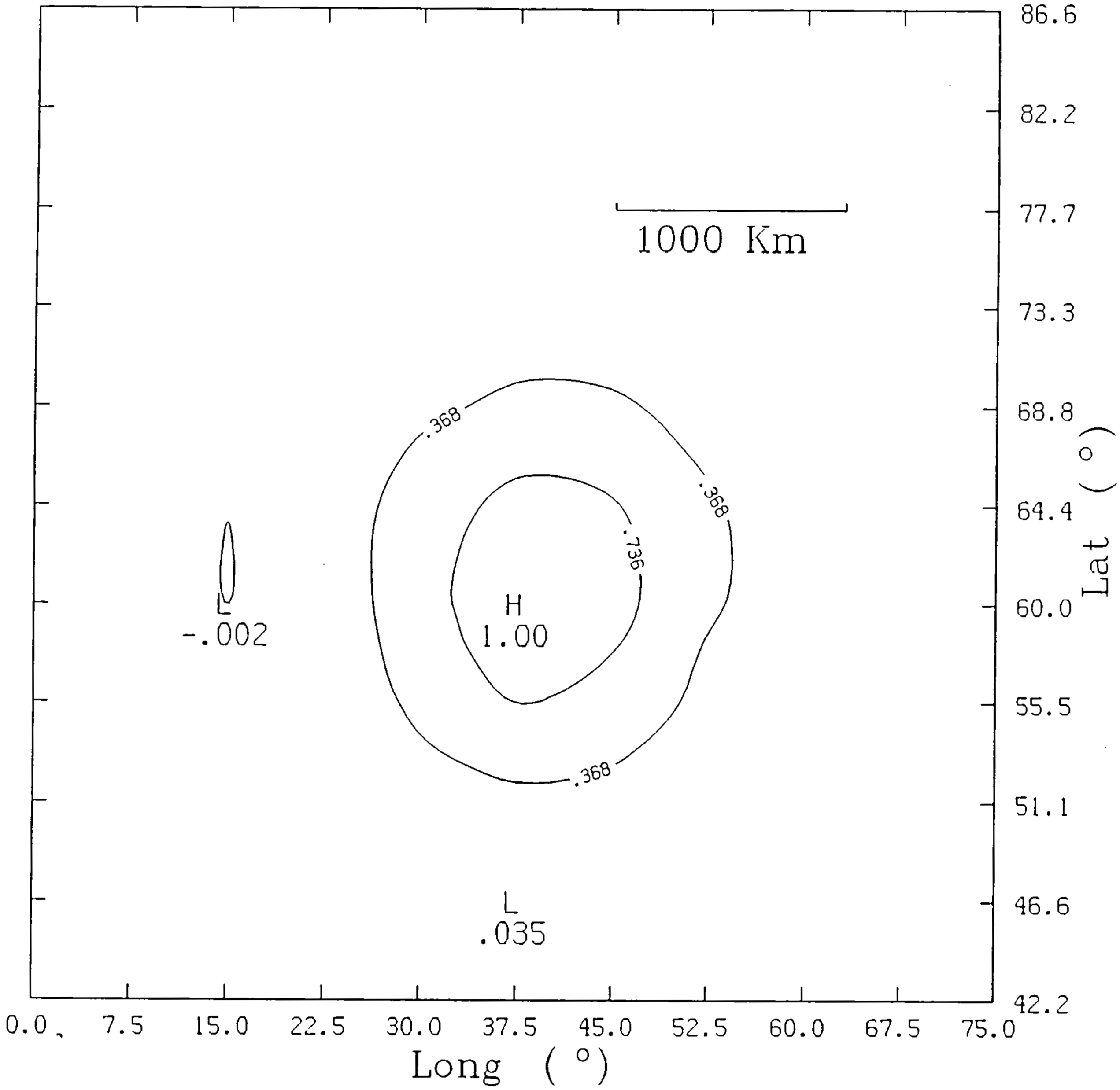

Figure 4: Time averaged surface thermal response to a heat source at a single grid point of strength of 400 W/m^2.

Fisica. 1985.

◇ Bell, T. L., 1980: Climatic sensitivity and fluctuation-dissipation: Some simple model tests. J. Atmos. Sci., 37, 1700-1707.

◇ Cess, R. D, (and 19 coauthors), 1989: Interpretation of cloud-climate feedback as produced by 14 atmospheric general circulation models. Science. 245, 513-516.

◇ Hansen, J. and S. Lebedeff, 1987: Global trends of measured surface air temperature. J. Geophys. Res., 92, 13345-13372.

◇ Hasselmann, K., 1977: Stochastic climate models. Part I: Theory. Tellus. 28, 473-484.

◇ Hyde, W. T., T. J. Crwoley, K.-Y. Kim and G. R. North, 1989: Comparison of GCM and energy balance model simulations of seasonal temperature changes over the past 18000 years. J. Clim., 8, 864-887.

◇ Leith, C. E., 1975: Climate response and fluctuation dissipation. J. Atmos. Sci., 32, 2022-2026.

◇ Leung, L.-Y. and G. R. North, 1990:; Information theory and climate prediction. J. Clim., 3, 5-14.

◇ Mengel, J., D. A. Short and G. R. North, 1988: Seasonal snowline instability in an energy balance model. Clim. Dynam., 2, 127-131.

◇ North, G. R. and R. F. Cahalan, 1981: Predictability in a solvable stochastic climate model. J. Atmos. Sci., 38, 504-513.

◇ North, G. R., J. G. Mengel and D. A. Short, 1983: Simple energy balance model resolving the seasons and the continents: Application to the astronononomical theory of the ice ages. J. Geophys. Res. 88, 6576-6586.

◇ North, G. R., 1984: The small ice cap instability in diffusive climate models. J. Atmos. Sci., 41, 3390-3395.