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Temporal climatic variabilities of global atmospheric, oceanic, and land surface parameters

SERGUEI G. DOBROVOLSKI* and BHASKAR J. CHOUDHURY

Laboratory for Hydrospheric Processes, Hydrological Sciences Branch, NASA/Goddard Space Flight Center, Greenbelt, MD 20771, USA

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RESUMEN

Se investigan los cambios en el estado de dos grupos de subsistemas climáticos mayores terrestres: (a) subsistemas "gruesos" (atmósfera y océano) y (b) subsistemas "delgados" (vegetación, nieve y hielo marino). Estos dos grupos se distinguen respecto a las variaciones temporales de su extensión espacial que son dinámicas para los subsistemas "delgados" y estables para los "gruesos".

Se evalúan unos modelos estocásticos que gobiernan las anomalías mensuales y de más largo plazo de los parámetros promediados globalmente, registrados remotamente e *in situ* de dichas capas, según una nueva modificación del método de máxima entropía y con truncación de frecuencia del espectro normalizado. Los análisis muestran que las variaciones temporales globales de los subsistemas "gruesos" están regidos por procesos estocásticos del tipo Bernoulli-Wiener, mientras que los de los subsistemas "delgados" globales se describen mediante el proceso Markoviano de primer orden con valores intermedios de los coeficientes.

Se avanza la hipótesis que las variaciones estocásticas en las cubiertas globales de hielo marino, nieve y vegetación dependen de sólo un número pequeño de variaciones independientes regionales y, por consiguiente, muestran todavía las características Markovianas de las últimas.

Con base en los análisis de datos observacionales y modelado estocástico, los autores proponen un concepto de mecanismos de variaciones climáticas que toman en cuenta subsistemas climáticos diferentes así como escalas diferentes de tiempo y espacio.

La hipótesis propuesta explica la estacionalidad de cambios en el área de cubiertas globales "delgadas" y el carácter no estacionario de los cambios de temperatura globales del tipo de marcha al azar (es decir, sin tendencias determinísticas).

ABSTRACT

Changes in the states of two groups of major earth climatic subsystems are investigated: (a) "thick" subsystems (atmosphere, and ocean), and (b), "thin" subsystems (vegetation, snow, and sea ice). These two groups are distinguished with respect to temporal variations of their spatial extent which is dynamic for the "thin" subsystems and stable for "thick" subsystems. Stochastic models governing monthly and longer-term anomalies of remotely sensed and *in situ*, globally averaged parameters of these layers (tropospheric temperature, sea surface temperature, vegetation, snow and sea ice cover areas) are evaluated according to a new modification of the maximum entropy method with frequency truncation of normalized spectra. The analysis shows that global temporal variations of global "thick" subsystems are governed by the Bernoulli-Wiener type stochastic processes, while variabilities of the

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* On leave from Water Problems Institute, Moscow, Russia

global "thin" subsystems are described by the first order Markov processes with intermediate values of coefficients. It is hypothesized that stochastic variations in global sea ice, snow, and vegetation covers depend on only a small number of independent regional variations and therefore still show the Markov characteristics of the latter.

On the basis of the analysis of observational data and stochastic modeling, the authors propose a concept of mechanisms of climatic variations which takes into consideration different climatic subsystems as well as different temporal and spatial scales. The proposed hypothesis explains the stationarity of changes in the area of global "thin" covers, and the non-stationary, random walk (i.e. without deterministic trends) character of global temperature changes.

1. Introduction

In many recent papers and books, month-to month, year-to-year, and longer changes in the climatic system are considered and described as almost exclusively deterministic. Consequently, deterministic experiments on GCMs, based on the description of physical processes at relatively small temporal scales, are often supposed to be the best tool for the development of climate theory. It is usually hypothesized explicitly or implicitly that detailed description of all specific physical processes within the climatic system (El-Niño, Northern and Southern Oscillations, greenhouse effect, strange attractors, etc.), and detailed information on the parameters of climatic variables should inevitably bring substantial improvement of the understanding and forecast of climatic changes.

At the same time, we know that although the speed of computers has increased by thousands of times since the first experiments with coupled ocean-atmosphere models, although the information on the climatic system has increased by millions of times, and although substantial progress has been made in understanding of many specific climatic mechanisms, the overall quality of climatic forecasts with lead time of 1 month and more, is approximately the same as 30 years ago. That means that there is no guarantee that further increasing of computer speed and information will bring a breakthrough as for understanding of something essential about climatic variability using the above approaches. Moreover, there is no guarantee that the diagnosis of the "phase" of recent and past slow natural global climatic changes is, in principle, possible with the use of GCMs.

Thus, there is an impression that the (stochastic) nature of climatic variations as well as the most essential facts about the theory of climatic variations and variability were discovered, and foundations of this theory were put many years ago by the Hasselmann's "Stochastic climate models theory" (1976)*, and by the works of Adem (1964, 1970, 1991, etc.) based on the description of processes of genuine climatic scale of about 1 month and longer. Unfortunately, later these approaches were somewhat faded by other works which were or still are "à la mode" but do not bring as much progress in understanding of climatic variabilities as cited papers did.

Let us remind that, according to the Hasselmann's Stochastic climate models theory, the most part of climatic variability (1 month and longer) is generated by stochastic, "white" (uncorrelated in time, at climatic frequencies), "pure" atmospheric weather forcing of more stable and inertial components of climatic system, ocean, land, and glaciers. In turn, the output of slow climatic subsystems affects the atmosphere. Finally, observed climatic changes represent the superposition of the initial atmospheric white noise and its transformations by different climatic subsystems. Two types of such a transformation are the most important: the first order Markov process (if there is a negative, linear in the first approximation, feedback between the atmosphere and the slow subsystem), and the Wiener process (random walk) if this feedback is small.

* Stochastic nature of climatic changes was also demonstrated by Mitchell (1966), Privalsky (1982, 1985), Polyak (1986), Gordon (1991), and by other researchers

In the present paper authors would like to return to these beginnings of the climate theories based on the genuine climatic time scales, and to demonstrate what improvements can be made within these approaches, using new sets of observational data and investigations of stochastic structure of climate. We were especially interested in comparing of results of unified analyses of different remotely sensed and conventionally observed global parameters and in relating these results to fundamental properties of the climate subsystems. The present paper is a stochastic analysis of the observed variabilities within major subsystems in terms of their dimensions, relative heat and water capacities, and type of spatial extent, in the light of the characteristic spatial dimensions of synoptic variability in the atmosphere which affects its interactions with the other subsystems.

In an attempt to develop stochastic approach to the climate studies, Dobrovolski (1992,1994) examined numerous time series of conventionally measured atmospheric, oceanic, hydrological parameters at various spatial and temporal scales, and suggested appropriate 0D, 1D, 2D dynamic-stochastic and stochastic models. These studies also took into consideration important monographs on the stochastic modelling of several hundred hydrological (Ratkovich, 1976) and climatic (Privalsky, 1985) variables.

As a result, the following concept was developed. At the local scale, changes in annual values or monthly anomalies of both primary variables (i.e. parameters which can be "accumulate": air and sea temperatures, sea level, atmospheric water content), and transfer-parameters (precipitation, evapotranspiration, solid glacial runoff, etc.) are described by first order Markov processes with various coefficients (or by somewhat more sophisticated models):

$$X(t) - E = k[X(t-1) - E] + a(t) \tag{1}$$

where X is a random process; t is time in months or in years; E is a mathematical expectation (mean value) of the process; k is a constant coefficient; $a(t)$ is a sequence of uniformly distributed normal, non correlated values with zero mean (white noise).

But when we move from the local scale to continental and global scales (i.e. if we take time series of spatially averaged parameters), the best fitting stochastic models tend to two "polarized" variants: white noise (zero coefficient k, transfer processes), and the Wiener descrete process (coefficient k equals unity, primary variables), while intermediate variants of models disappear. It was hypothesized that the limited radius of spatial correlation of synoptic processes in the atmosphere could eventually be the reason of this phenomenon.

Thus, the behavior of the stable, "thick" climate subsystems with relatively large water and heat capacities (ice sheets, oceans, and the atmosphere which is connected to them) seemed to be elucidated. Nevertheless, monthly data on the above processes were still insufficient. The temporal variabilities of "thin" subsystems with limited capacities, and limited and changing aerial extension, were not evaluated because conventional observations could not describe their climatic dynamics.

Remote sensing has provided time series of global vegetation cover (Choudhury, and Di Girolamo, 1994), snow cover (Robinson *et al.*, 1993; Foster and Chang, 1993), sea ice (Gloersen *et al.*, 1992; Parkinson and Gloersen, 1993; Gloersen, 1995), and tropospheric temperature (USDC, 1994). Also, further improvement of parametric techniques of time series analysis has occurred and computer programs have become available (Privalsky *et al.*, 1992; Privalsky and Jensen, 1993). Investigation of temporal variability of these earth covers using advanced time series analysis thus becomes possible.

Specific objectives of the present study were to:

- reevaluate the local-global polarization of stochastic models for the primary variables monitoring "thick" subsystems (ocean, and atmosphere) using remotely sensed and blended remotely sensed/in situ data sets (air and sea temperatures);
- identify stochastic models for the remotely sensed data related to changes in the "thin" climatic subsystems (vegetation, snow and sea ice covers) and assess the parameters of these models;
- make appropriate modifications of the concept of the local-global polarization of stochastic models, taking into consideration results for "thin" layers.

2. Data and methods

2.1. Time series observations

In the present work we have analyzed tropospheric temperatures (channel 2R) from (USDC, 1994). The length of the series was 180 months (Jan. 1979 - Dec. 1993). The time series of monthly sea surface temperature (SST) anomalies are *in situ*, blended and remotely sensed data from Parker (1995) and Reynolds *et al.* (1989); the length of the record being 171 (Jan. 1981 - March 1995), 84, and 84 months (Jan. 1982 - Dec. 1988), respectively. Development of data is documented by Parker *et al.* (1994).

The following monthly satellite based data sets for the "thin" covers have been used: (1) globally (0°- 40°N and 55°S - 75°N) averaged 37 GHz polarization differences related to the density of vegetation cover (Choudhury and Di Girolamo, 1994); (2) land snow area for the Northern Hemisphere (IPCC, 1992); (3) global sea ice area (Gloersen, 1995). Detailed description of appropriate methods of obtaining and processing remote sensing data can be found in Choudhury (1989), Foster and Chang (1993), Gloersen *et al.* (1992), Parkinson and Gloersen (1993), and Robinson *et al.* (1993). The lengths of these series are respectively 84 months (Jan. 1979 - Dec. 1985), 228 months (Jan. 1973 - Dec. 1991), and 96 months (Jan. 1979 - Dec. 1986).

2.2. Stochastic analysis

The above mentioned time series of 37 GHz polarization differences, sea ice area, snow cover area were analyzed in 9 variants: (1) in initial form; (2) after removing the seasonal means by converting monthly mean values to monthly anomalies; (3) after removing the seasonal mean and variability by dividing the monthly anomalies by their standard deviations; (4 - 9) divided into two segments of equivalent lengths. Series of primary variables of "thick" subsystems - global microwave tropospheric temperatures and oceanic temperatures (*in situ*, remotely sensed and blended) were already in the form of monthly anomalies and, thus, variants (2-3, 5-6, 8-9) were used in this case. In addition, for the parameters of "thick" climatic subsystems month-to-month increments of series were also analyzed to yield information on "transfer" processes. In turn, for each variant of analyzed series about 40 sets of parameters of stochastic models were proposed. Thus, all in all 60 variants of main time series were investigated, more than 2,000 variants of stochastic models were tested, and more than 1,000,000 scalar parameters of stochastic and statistical models were visually checked and compared.

The basic principles of stochastic analysis used in this study are described in (Dobrovolski, 1992). Each time series, $x(t)$ was considered as a segment of a realization of a stationary random process, $X(t)$:

$$X(t) = k_1 X(t-1) + \ldots + k_m X(t-m) + a(t) \tag{2}$$

Here t is time in months (or, for the series of annual values, in years); k are coefficients of autoregression (AR) of orders $1, \ldots, m$; $a(t)$ is a sequence of uniformly distributed normal, non correlated values (residual white noise). In equation (2) for the purpose of simplicity we considered the mean value (mathematical expectation) of the process $X(t)$ as being zero. Let us remind that equation (2) describes the first order Markov process (the first order AR process) if $m = 1$.

This method of describing time series, also known as the "maximum entropy method" (MEM) is thus reduced to fitting autoregressive processes to the time series, i.e. to the determination of the order m, coefficients k, and variance (or standard deviation) of residual white noise, a(t). Appropriate formulae, including the above parameters, describe the spectral densities, autocovariances, and other important functions related to the investigated processes.

Changes made to our variant of the time series analysis are as follows:

- we have modified the Burg scheme of obtaining a first trial coefficient of the models: it is calculated using an iterative procedure which provides a best forecast of the process;
- the characteristic equation of autoregression is solved for each set of trial coefficients in order to check the stationarity of the model;
- five criteria (Akaike's, Akaike's Information Criterion, Parzen's, Schwarz-Rissanen's, and Hannan-Quinn's) are used for the stochastic models identification (Privalsky and Jensen, 1993);
- a procedure of frequency truncation was used while obtaining normalized spectra.

The last modification deserves some comments. In order to obtain spectra in logarithms, and to compare spectra of different processes one usually calculates normalized spectra dividing the spectrum by the variance of the process:

$$S_n(f) = S(f)/\sigma^2 x \tag{3}$$

But it is known that estimations of the variance of the process (as well as estimations of its mean value) contain considerable errors. This difficulty is especially serious with respect to the part of the variance generated within the range of low frequencies in the case of realizations close to the nonstationary ones. As a result, the calculation of normalized spectra using Eq. (3) leads to a spread of errors of spectral estimations from low frequencies to the whole frequency range. Thus, some procedure excluding values of spectra at low frequencies is needed. For this purpose, we divided the spectra not by the total variance but by the part of the latter which is generated within the range of frequencies between a chosen "threshold" one and the Nyquist frequency (0.5 cycles/discreteness interval). In other words, we take only that part of the variance which is located to the right of the threshold frequency, under the curve $S(f)$. Finally, calculations are made using following equation for the "truncated" spectrum, S_t :

$$S_t(f) = S_n(f)^*(s_l + s_h)/s_h \tag{4}$$

where

$$s_l = \int_0^{f_t} S_n(f)df; \quad s_h = \int_{f_t}^{0.5} S_n(f)df.$$

3. Results

In Figure 1A traditional MEM spectral estimations of parameters of state for the atmosphere and the ocean are shown (curves 1 - 4). In the same figure a traditional spectrum estimation for the realization of the Wiener process is presented (line 5, the length of this realization is equal to the length of the longest observed series, i.e. 180 points; the sample estimation of the coefficient k_1 is equal to 0.99 in this case). It is evident that the difference between spectra is great, and substantially exceeds confidence intervals for individual spectra (as shown in Fig. 1B). For the

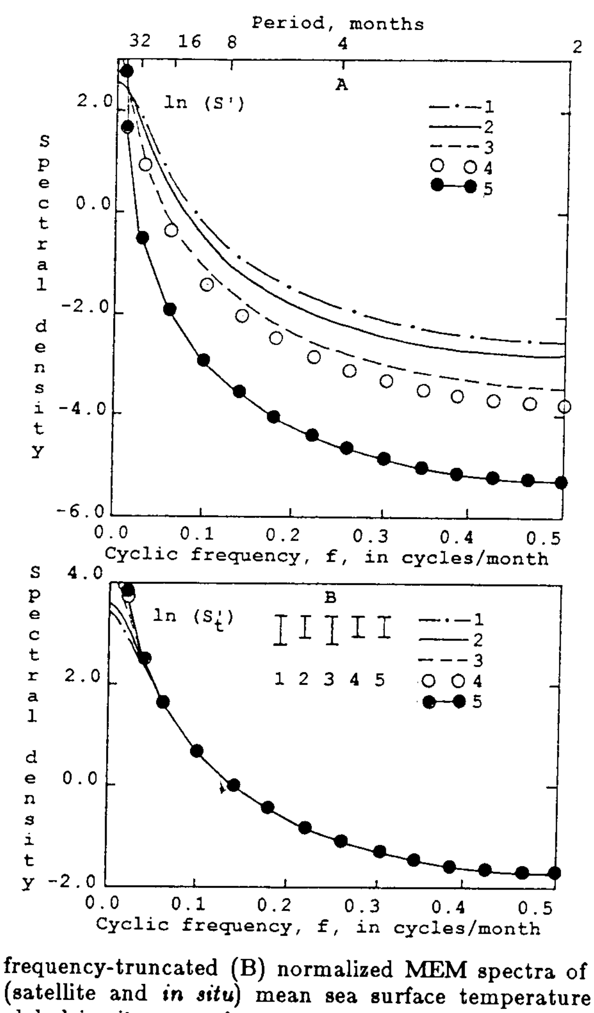

Fig. 1. Traditional (A) and frequency-truncated (B) normalized MEM spectra of parameters of global "thick" subsystems. 1 - blended (satellite and *in situ*) mean sea surface temperature anomalies for the Northern Hemisphere (0°- 60°N); 2 - global *in situ* sea surface temperature anomalies; 3 - satellite sea surface temperature anomalies for the Northern Hemisphere (0°-60°N); 4 - global microwave mean tropospheric air temperature; 5 - realization of the Wiener process (180 points). Vertical lines on Fig. 1B denote 95% confidence limits. Data from: Reynolds *et al.* (1989) (1, 3); Parker (1995) (2); USDC (1994) (4).

truncated spectra we used threshold frequency, f_t equal to 6/(series length in number of points), from Dobrovolski (1996). In order to obtain comparable spectral estimations and not to lose much information about traditional spectra we used, in the formula for f_t , the same value of series (maximal) length, 180 months, for each spectrum.

Truncated normalized spectra are shown in Figure 1B. For the most part of the frequency range spectra practically coincide. Calculations show that Cromer confidence intervals (in the top part of the Fig. 1B) can be used in this case not only for individual spectra but also as absolute ones. The divergence among spectra becomes statistically significant for the frequencies less than approximately 0.03 - 0.04, i.e. only for 6%-8% of the total frequency range where estimations of the individual spectra become unreliable because of limited length of time series.

Figure 1B shows that spectra of different records of global atmospheric and oceanic temperatures obtained using different methods coincide with each other and with the spectrum of the discrete Wiener process:

$$X(t) = X(t-1) + a(t) \tag{5}$$

In order to check the applicability of model (5) to the description of investigated processes, one must also process series of appropriate month-to-month increments. In accordance with the Wiener model they are expected to be non-correlated in time, i.e. demonstrate zero spectrum. Indeed, calculations show that time series of increments relating to "thick" climate subsystems are described by the white noise process, $a(t)$ or are close to it. As an example, the spectrum of month-to-month increments of global mean sea surface temperature anomalies (data from Parker, 1995) is shown in Figure 2.

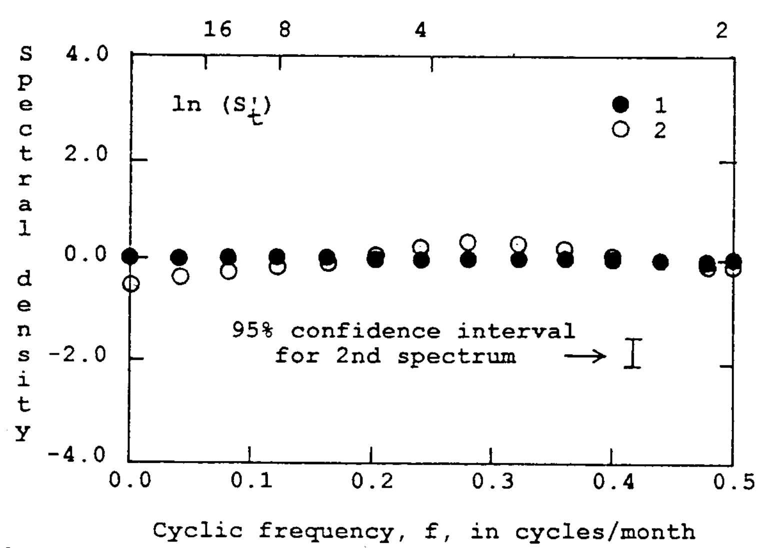

Fig. 2. Spectra of month-to-month increments of global mean sea surface temperatures. Line 1 - zero order model ("white noise"); line 2 - 2nd order model. Data from Parker (1995).

Truncated normalized spectra of parameters of the thin subsystems, snow area in the Northern Hemisphere (which includes almost all seasonally variable snow cover on the Earth), mean global vegetation cover density (37 GHz PD), and global sea ice area are shown in Figure 3. As in the case of "thick" subsystems, the spectra for the "thin" subsystems in Figure 3 coincide with each other taking into consideration the confidence limits of estimates, apart from zero and neighboring frequencies. Monthly anomalies of the investigated parameters of global thin subsystems , unlike those for thick subsystems parameters, are satisfactorily described by the Markov first order process with coefficient 0.5 - 0.6.

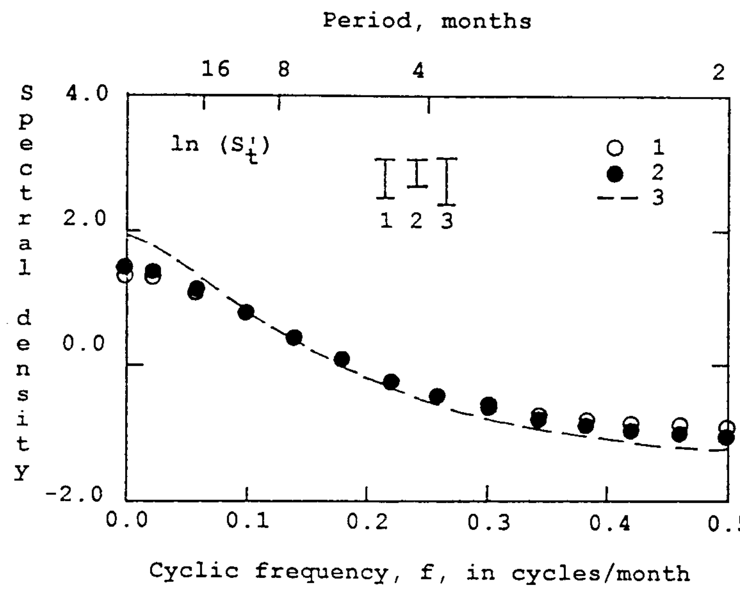

Fig. 3. Spectra of changes in parameters of "thin" climatic subsystems and related parameters: 1 - North Hemispheric snow area; 2 - global 37 GHz polarization difference (density of vegetation cover); 3 - global sea ice area. Vertical lines denote 95% confidence intervals. Data from: IPCC (1992) (1); Choudhury, Di Girolamo (1994) (2); Gloersen (1995) (3).

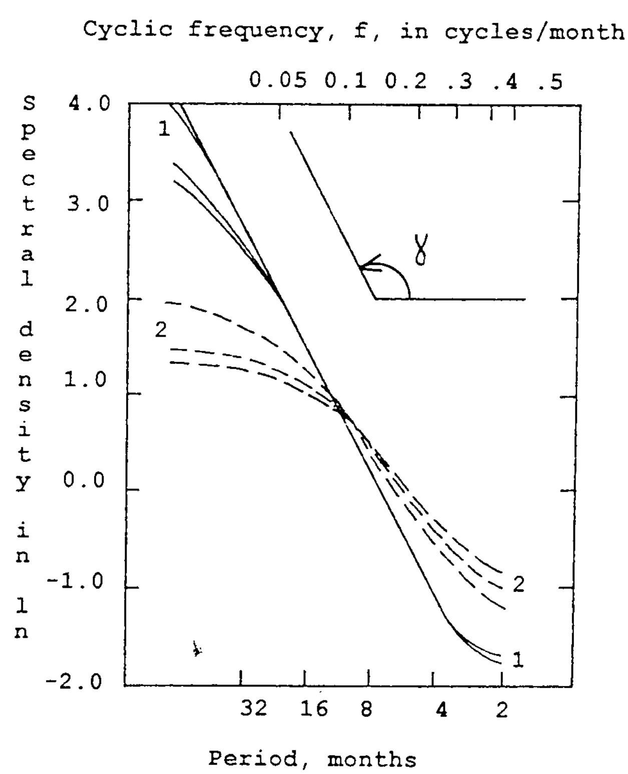

Fig. 4. Spectra of parameters of "thick" subsystems (1, solid lines, from Fig. 1), and parameters of "thin" subsystems (2, dashed lines, from Fig. 3) in bilogarithmic coordinates. γ denotes an angle with tangent -2.

Spectra of both "thick" and "thin" subsystems are presented in Figure 4 in bilogarithmic coordinates. It is evident that the first spectra as well as the spectra of the Wiener process are close to the straight line with tangent -2 and thus corroborate the "minus two" law of climatic variability without feedback (Hasselmann, 1976) and the concept of global climatic temporal scale invariance. At the same time "thin" subsystem spectra give distinctly differing pattern.

4. Modeling the dynamics of "thin" subsystems

In order to illustrate and understand the main features of local and global changes in the major state parameters of "thin" climatic subsystems, the following simple stochastic model can be suggested.

The main "thin" subsystems are confined to one (snow cover) or two (sea ice) latitudinal belts. In principle, global vegetation cover also can be presented as two latitudinal belts: within midlatitudes of the Northern Hemisphere and within equatorial and tropical latitudes. We shall consider as a first example snow cover which has only one (southern) variable boundary.

Consider a meridional strip i (Fig. 5) with east-west width, ΔX approximately equal to the characteristic dimension of atmospheric synoptic eddies, i.e. about 1,000 km. Investigations summarized in Dobrovolski (1992) show that the correlation radii of monthly anomalies of atmospheric temperature, precipitation, air humidity, and horizontal atmospheric vapor transport are of similar magnitude. Assume that within this strip and near the climatological snow boundary, conditions at the land surface are homogeneous.

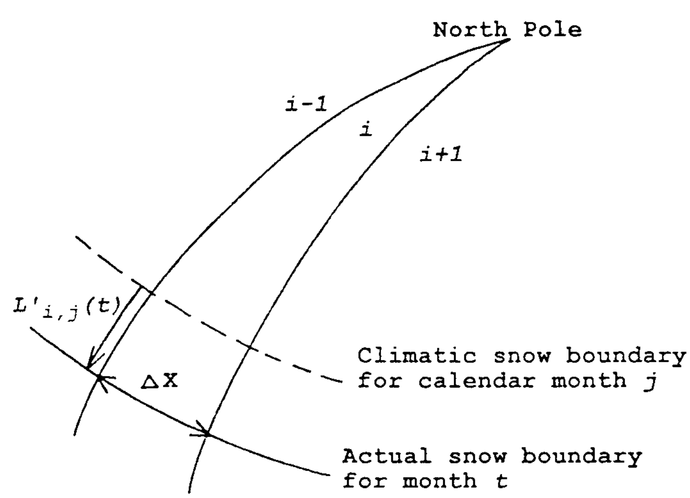

Fig. 5. Schematic diagram for the simple stochastic model of snow cover.

Initially, ignore seasonal variations and suppose all months to be specific calendar month j. Thus, we are looking for stochastic equation for the anomaly of the distance between the North Pole and the southern edge of the snow cover, $L'_{i,j}(t)$ where t is time. As stated above, "intermediate" (about 1 month) and longer time scales are investigated, so 1 month is taken as the temporal interval, Δt.

On intermediate time scales there are considerable negative feedbacks in the interactions between the atmosphere and the underlying surface (Adem, 1991). If the boundary of snow, sea ice, or vegetation cover in a specific month differs substantially from its climatology, there will be a tendency to "return" towards normal through the mechanisms of radiation processes, evaporation and evapotranspiration, vertical turbulent heat exchange, large scale ocean turbulence (for sea ice), snow and ice melting or formation, vegetation degradation or expansion (Piterbarg, 1985; Dobrovolski and Rybak, 1992). For the snow cover as well as for the other main "thin" climatic subsystems, this feedback in its simplest linear form can be described by the following equation:

$$L'_{i,j}(t) = kL'_{i,j}(t-1) \tag{6}$$

where values of approximately constant coefficient k lie between 0 and 1 .

Now let us superimpose an atmospheric forcing onto (6). Following the concept of two-scale weather-climate separation* by Hasselmann (1976), we shall consider an atmospheric forcing term, $a_{i,j}(t)$ at intermediate temporal scale as a temporally uncorrelated random variable. The mathematical expectation of $a_{i,j}(t)$ (white noise) for specific i and j will be zero.

A simple model of hemispheric snow cover proposed by Koster (1995) and empirical models of the same type for other "thin" subsystems can give an idea of how to construct specific equations for $a_{i,j}(t)$. For example, in Koster's model, the temperature-related forcing of the snow cover is either the area (or extent) of anomalous air temperatures where there has been no snow, or the snow area under the air with positive temperatures. In turn, the meridional extent of air temperature anomaly can be written in terms of the mean climatological air temperatures and the magnitude and sign of the actual anomalies at each latitude. Yet, as stated above, we try not to construct specific exact equations for each "thin" subsystem but to present a common simple approach to the type of equations governing their variabilities.

Thus, the equation for the snow extension anomaly for month t will be:

$$L'_{i,j}(t) = kL'_{i,j}(t-1) + a_{i,j}(t) \tag{7}$$

As ΔX is chosen to be approximately equal to the horizontal correlation radius of atmospheric processes on intermediate temporal scales, the $a(t)$ for different longitudes will be uncorrelated and will result in a global atmospheric forcing, $A(t)$ which is supposed to be white noise also. The result of the integration along the lateral boundary of the cover can be represented as follows:

$$S'_j(t) = kS'_j(t-1) + A_j(t) \tag{8}$$

where S'_j- is the monthly anomaly of hemispheric snow cover area; the global variance of atmospheric forcing, $A_j(t)$ is equal to the sum of local variances, $a_{i,j}(t)$.

Table 1. Seasonally changing coefficients of autoregression, k of global monthly snow area anomalies (Eq. 8). Calculations were made using data from (IPCC, 1992)

Season	Coefficient, k	Coefficient minus mean annual value
Winter	0.55	-0.06
Spring	0.74	0.13
Summer	0.74	0.13
Fall	0.45	-0.16

* It is shown in (Hasselmann, 1976) that internal atmospheric forcing at frequencies less than about 1 cycle/month can be represented, in the first approximation, as the white noise

It is evident that, in reality, (8) is an equation of a stochastic process with seasonally varying parameters. Nevertheless, analysis of observational data suggests that seasonal changes in coefficients k are not dramatic, at least they have an order of the error bars on k itself. As an example, estimations of coefficient k for global snow area anomalies and for different seasons are presented in Table 1. It is clear that the differences between the estimations of k for different seasons and its mean annual value are comparable with the mean error of these estimations (about 0.14). Seasonal dynamics in standard deviations of $S_j(t)$ and $A_j(t)$, can be removed using the procedures described in Part 2. Finally, we obtain an equation for seasonally-normalized monthly anomalies of the areas of "thin" subsystems:

$$s'(t) = ks'(t-1) + a'(t) \tag{9}$$

The form of equation (9) is thus identified by us as appropriate for the description of changes in the global area of climatic "thin" subsystems and related parameters. Using the above considerations we can understand the difference between models for the main parameters of "thin" and "thick" subsystems.

5. Local-global polarization phenomenon

Two conditions are necessary for the existence, at the global scale, of stochastic processes with negative feedback related to the interactions of the atmosphere and underlying surface at intermediate and longer time scales. Firstly, a negative feedback of the above type must exist at the local and regional scale; secondly, these processes must not be "lost" in the process of global averaging; in other words, the relative range of global variations governed by the negative-feedback mechanisms must be significant.

It is convenient to measure the relative variability of the global processes by their coefficient of variation, i.e. the standard deviation of the parameter divided by its mean value. After the global summarizing of Eq. (7) which describes regional snow area anomalies, the standard deviation of the anomaly will increase by $(B/\Delta X)^{1/2}$, and the mean value of the snow area will increase by $B/\Delta X$ where B is the characteristic length of the variable part of the snow cover boundary. Thus, the coefficient of variation will decrease by:

$$CVD = (B/\Delta X)^{1/2} \tag{10}$$

Estimations of mean values of parameter CDV for the snow cover area as well as for the sea ice area and the vegetation aerial density (calculated using the same formula) are presented in Table 2. Coefficient of variation for these climatic subsystems will decrease only by 3 - 5 times, and process (9) will still be "visible".

The local and regional variations of the state parameters of "thick" climatic subsystems like sea surface temperature anomalies or air temperature anomalies, at intermediate and larger temporal scales, are in the first approximation described by the same type of models with negative feedback (Adem, 1970, 1991; Mendoza *et al*, 1996; Dobrovolski and Yarosh, 1980; Privalsky, 1982; Privalsky, 1985; Dobrovolski, 1992, 1994). However, global integration of the ocean surface temperature and air temperature (more precisely, the heat content of the troposphere and oceanic upper layer) is carried out over the whole surface of the ocean. In this case approximate formula for the parameter of decreasing of variation coefficient will be:

$$CVD = (A/S_x)^{1/2} \tag{11}$$

where A is the total area of the troposphere or the total oceanic area, and S_x is the area of a circle with diameter equal to the characteristic integral spatial correlation scale, ΔX (approximately 10^3 km). Estimations of CVD for "thick" subsystems are also presented in Table 2, and are equal to 21 and 26.

Table 2. Parameters of global decrease of coefficient of variability, CVD, and CVD^2 for different climatic subsystems. Estimations using Eq. (10) for "thin", and Eq. (11) for "thick" subsystems.

Subsystem	CDV	CDV^2
A. "Thin" subsytems:		B/ΔX
Snow cover area	3.5	12.2
Sea ice area	4.3	18.3
Vegetation cover spatial density	5.1	26.0
B. "Thick" subsystems		A/S_x
Upper oceanic layer heat content	21.3	454
Tropospheric heat content	25.5	649

Thus, the globally-averaged stochastic variations of primary variable of the "thick" subsystems such as mean oceanic and atmospheric temperature, of the negative-feedback type (9) are very weak and, perhaps, almost nonexistent at the investigated time scales. In this case other mechanisms (radiative processes, volcanic eruptions, processes involving deep oceanic layers or global El Niño type events, different nonlinear processes, anthropogenic effects) can play important roles. Yet, for the temporal scale under consideration, all these mechanisms, together with the processes of interactions within the climatic system, appear to create only small relative changes in the globally averaged "thick" subsystems whose behavior is therefore described, in the first approximation, by the no-feedback Wiener type processes (Hasselmann's "-2 degree" law of climatic variability without feedback, Figure 4).

In other words, the behavior of the parameters describing the "thin" and "thick" climatic subsystems can be explained as follows. In the case of "thin" climatic subsystems, only a few groups of cyclones and anticyclones near the lateral frontier of the cover are important. That is why temporal changes of globally averaged parameters (e.g. area of snow cover) have the same features as the changes at regional level within the size of individual eddy. At the same time, the mean states of the ocean and the atmosphere at the intermediate and longer temporal scales are determined by such a large number of atmospheric and oceanic eddies that the variation of the mean state is related to the specific features of none of these groups of eddies.

Consequently, the characteristic time-scale or "integral time scale" (Yaglom, 1987) of the changes in land snow cover area and in the areas of other "thin" subsystems (Yaglom, 1987) is equal to approximately 1.5 months and is obviously related to the intermediate time scale (or a frequency at which spectrum of synoptic variability becomes flat), whereas the characteristic time of changes in mean global temperature seems to be at least of the order of dozens of years*.

* We do not exclude that in certain cases local events of, say, El-Niño type, volcanic eruptions, etc. can affect global temperature at other time scales.

Strictly speaking, in the real climatic system atmospheric forcing at local scale, $a(t)$ slightly differs from white noise because of inertia within the combined atmosphere- surface system. Nevertheless, it can be shown that the different character of spatial integration over "thick" and "thin" subsystems would lead to the above described difference in local-global patterns in this case also.

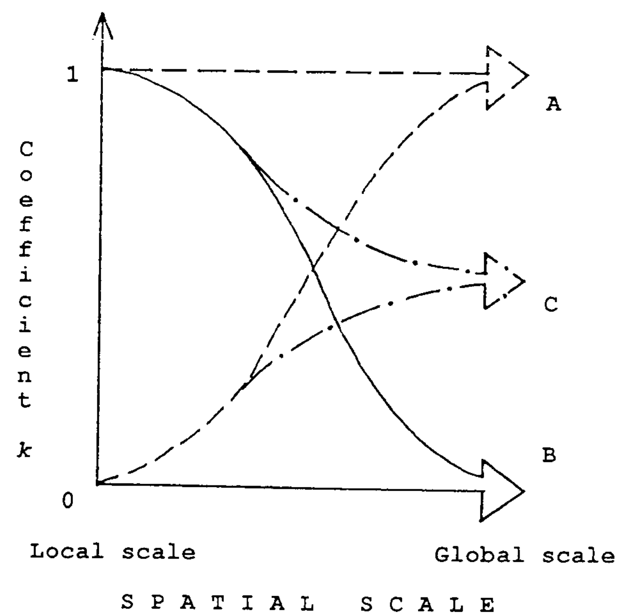

Fig. 6. Illustration of the local-global contrasts in temporal changes in the climatic subsystems. A - bulk (temperature, water equivalent) parameters of "thick" and "thin" subsystems; B - "transfer" parameters of thick and thin subsystems; C - area or density of thin subsystems; k - coefficient of the Markov first order process (1).

Finally, Figure 6 hypothetically illustrates the local-global contrasts for "thin" and "thick" subsystems. At local and regional scales, intermediate-scale (monthly) temporal variations of different parameters of climatic subsystems are described in the first approximation by the first order Markov process with different values of coefficient k in the range (0,1) (Eq. 1). As we increase the scale of spatial averaging of the process, k tends to 1 for bulk (accumulation) processes, e.g. temperature of ocean and atmosphere, perhaps water equivalent of land snow cover (last hypothesis is suggested by the results of stochastic experiments on 3D model of the global hydrological cycle, Dobrovolski *et al.*, 1995), to 0 for "transfer" processes (increments of accumulated water or heat), and to some intermediate value for the area of "thin" covers (land snow, sea ice, vegetation cover). Respectively, model (1) tends to the model of Wiener process, to white noise or remains the first order Markov process.

6. Discussion and conclusion

The above analysis demonstrates, from our point of view, the existence of the global climatic subsystems of the "third type" whose behavior differs from that of both the atmosphere and the ocean: snow cover, vegetation cover, and sea ice. These three Earth covers are not as thick as the ocean and the atmosphere, areas of their spatial extent are relatively small, and their boundaries are extremely unstable. Unlike the ocean and the atmosphere, their most significant global parameter is not temperature, heat or water content, but area of spatial extent. Although at small temporal and spatial scales these climatic subsystems are governed by different physical processes, their behavior at climatic time scales and at the global spatial scale has essential common features. The temporal variability of their global areas is much more stationary than that of the variability of global air and ocean temperature, and relative role of high frequencies of their oscillations is much more important than that of global temperature variations.

Only analysis of remotely sensed data on the state of "thin" climatic subsystems can bring answers to important questions: what are the best fitting stochastic models for the description of their variabilies? Does their behavior differ significantly from the behavior of the global temperature, and what is stochastic model for the latter? How does the whole climatic system work? We hypothesize that the answers to these questions can be as follows.

1. The fundamental model of the stationary first order Markov process with different values of the coefficient (globally averaged mean value is 0.4 - 0.6) governs local monthly anomalies of the "thin" climatic subsystems area variations. The mean integral time scale (correlation time) of snow cover, sea ice, and vegetation cover local variations is about 1.5 moths, and is evidently related to the characteristic time scale (about 1 month) of the internal atmospheric forcing. Surprisingly enough, the same stochastic model of the first order Markov process, with the same mean value of the coefficient, is valid for the description of temporal variability in the monthly anomalies of globally averaged areas of "thin" subsystems. Thus, a local-global "invariance" (or the lack of "local-global contrast") of the models and their coefficients takes place in the case of the "thin" subsystems.

2. On the contrary, a sharp contrast between, on one hand, local and, on the other hand, global variabilities of primary (accumulation) parameters of all "thick" climatic subsystems, takes place. At the local scale time series of monthly anomalies of air and sea surface temperatures are governed, in the first approximation, by the same, stationary first order Markov process model with different values of coefficient (for local air temperatures this coefficient is relatively small). But time series of globally averaged sea surface temperature and global air temperature are essentially non-stationary, and are very close to the realizations of the discrete Wiener process model. By the way, no statistically significant deterministic monotonous trends were found by us in the analyzed global and local time series; there were only nonstationarities in the mean variances of the processes, of the type of random walk.

3. Mechanism which governs the above processes can be described as follows. In accordance with the Hasselmann's two-scale weather-climate separation hypothesis, synoptic processes in the atmosphere create at climatic frequencies (1 cycle/month and less, i.e. far from main synoptic frequencies, 1 cycle/several days) a variability which has much more flat spectral density in comparison with the form of spectra near synoptic frequencies. Correspondingly, this random "weather" forcing at *climatic scale* possesses only small temporal correlation.

Specific mechanisms at the sea and land surface, which transform this initial input forcing, involve different physical processes: evaporation and evapotranspiration, vertical turbulent heat exchange, snow melting, etc. But the result of the transformation of the atmospheric forcing, at intermediate (1 cycle/month) and lower frequencies, in the first approximation, is eventually similar: input and output "signals" form a sort of negative feedback. Thus, anomalous heating of the sea surface during specific month (for instance, by anomalous winds and currents from south) causes anomalous evaporation, vertical turbulent cooling of the surface, anomalous radiative vertical fluxes. In turn, these mechanisms tend to smooth and, eventually, to eliminate the initial anomaly.

Just the same happens at the land surface with the "thin" climatic subsystems. Abnormal snow or vegetation cover eventually get under climatic conditions which are not favorable for maintaining snow or vegetation and tend to return the situation to the norm by the mechanisms of, respectively, snow melting and droughts.

Mathematically speaking, these processes of, say, linear negative feedback, are described by the first order Markov process (first order autoregression process) with coefficient between 0 and 1. Because of relatively small area of the most important in these cases, lateral parts of "thin" subsystems (covers), global averaging of snow or vegetation variations involves only few groups of independent atmospheric eddies which can not totally eliminate the anomalous effects of each other. Finally, the negative feedback mechanism and appropriate stochastic model is still "seen" and works at the global scale.

On the contrary, global averaging of the sea surface and air temperature is performed over the whole area of the ocean or the globe. During specific month, hundreds of independent anomalous groups of atmospheric (and oceanic) eddies form global mean air and sea surface temperature, and eventually almost eliminate the anomalous effect of each other. At the same time, heat capacity of the ocean and the atmosphere grows almost linearly with the area of the part of "thick" subsystems under consideration. Under these conditions, during the procedure of global averaging, global negative feedback mechanism of the above type governs only small relative anomalies. Finally, at the global scale, it is practically destroyed by the great number of random local nonlinearities. Residual global temperature anomalies, without sufficient feedbacks, follow the pattern of the random walk with independent temporal increments.

At first sight, realization of this nonstationary process resembles a nonstationary process with monotonous deterministic trend. However, the difference between these two types of nonstationarity is crucial, and is easily detected using the analysis of temporal increments. The mean value of the temporal first increments (differences) of the process with monotonous deterministic trend is statistically significant, i.e. the absolute value of the mean increment is more than the error of its estimation. At the same time, the mean value of the first increments of discrete Wiener process does not differ statistically from zero.

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