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Global Correlation between Surface Heat Fluxes and Insolation in the 11-Year Solar Cycle: The Latitudinal Effect

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Abstract—Because of the small amplitude of insolation variations (1365.2–1366.6 W m^{-2} or 0.1%) from the 11-year solar cycle minimum to the cycle maximum and the structural complexity of the climatic dynamics, it is difficult to directly observe a solar signal in the surface temperature. The main difficulty is reduced to two factors: (1) a delay in the temperature response to external action due to thermal inertia, and (2) powerful internal fluctuations of the climatic dynamics suppressing the solar-driven component. In this work we take into account the first factor, solving the inverse problem of thermal conductivity in order to calculate the vertical heat flux from the measured temperature near the Earth's surface. The main model parameter—apparent thermal inertia—is calculated from the local seasonal extremums of temperature and albedo. We level the second factor by averaging mean annual heat fluxes in a latitudinal belt. The obtained mean heat fluxes significantly correlate with a difference between the insolation and optical depth of volcanic aerosol in the atmosphere, converted into a hindered heat flux. The calculated correlation smoothly increases with increasing latitude to 0.4–0.6, and the revealed latitudinal dependence is explained by the known effect of polar amplification.

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1. INTRODUCTION

One of the fundamental problems of present-day science consists in the study of controlling actions on the global climate. It is difficult to describe in detail considerable local disturbances related to a number of nonlinear processes in the atmosphere and ocean. The indices of the global surface temperature (see, e.g., (Smith and Reynolds, 2005)) averaged over the surface have considerable dispersion, and the reconstructions of such global indices in the past based on tree rings substantially differ from one another; thus, the correctness of these indices is doubtful (Makarenko et al., 2013). As a result, it is difficult to interpret the general pattern of global climatic changes and to search for the theoretically predicted (Budyko, 1968; Stevens and North, 1996) response to quasiperiodically small external actions such as the 11-year solar activity cycle in real data (Scafetta and West, 2007; Zhou and Tung, 2010). Strictly speaking, temperature (in contrast to energy) is not an additive quantity: it cannot be summed up or, consequently, averaged. The near-surface temperature physically results from many factors: it primarily depends on thermal inertia and the reflectivity of the underlying surface. It was indicated (Volobuev, 2013, 2014) that consideration of thermal inertia in the inverse problem of heat conductivity results in the presence of a pronounced 11-year cycle in the near-surface heat flux converted from the average annual temperature at Vostok station at the Antarctic ice dome center. The variation amplitude is approximately three times as large as the average response to the 0.1% change in insolation predicted by global circulation models. This effect is probably explained by the known phenomenon of polar amplification, i.e., by positive feedback: a decrease in temperature results in an increase in the polar ice area, and the corresponding increase in albedo leads to an even more considerable temperature reduction. The numerical estimates of this effect approximately correspond to observed variations in the surface temperature. Thus, the high sensitivity to small flux variations is explained by an accurate (averaged over a year) balance of the incident and reflected heat. This balance removes the main (constant) part of the heat flux in dynamics models. On the other hand, even in the case of a quasi-homogeneous underlying surface and regular climatic conditions in the central Arctic Regions, it can be shown that the surface temperature does not correlate with insolation in the 11-year cycle and that the correlation is significant for the heat flux (Volobuev, 2014). The absence of correlation in this case is most probably related to the nonlinear shape of the 11-year solar cycle. For heat fluxes calculated from the averaged global temperature indices, a correlation is also absent (Volobuev, 2013) for all solar activity cycles except the strongest ones (cycles 19 and 21).

The motivation of this work is an attempt to convert local surface temperatures into heat fluxes. We thereby hope to obtain physically reasonable global climate indices that can be correctly summed up and averaged at least within an individual latitudinal belt. The paper is organized as follows: Section 2 presents the physical backgrounds and approach to the calculation of apparent thermal inertia (ATI) from the measured albedo and seasonal temperature variation amplitude and the formulas for calculating the heat flux density based on the inverse heat conductivity problem. Section 3 indicates the sources and specific features of the data used in this work. The results are discussed and the conclusions are made in Section 4.

2. CALCULATION OF ATI AND HEAT FLUXES

We now consider heat propagation in a homogeneous half-space heated at the boundary. In such a case, we arrive at the one-dimensional inverse problem of heat conduction

$$\frac{\partial T}{\partial t} = \frac{\lambda}{\rho c}\frac{\partial^2 T}{\partial z^2}, \tag{1}$$

with a time variable boundary condition

$$\lambda\frac{\partial T}{\partial z}\bigg|_{Z=0} = q(t), \tag{2}$$

where λ, c, and ρ are heat conductivity, heat capacity, and density of a medium, respectively. Assuming that heat flux $q(t)$ is unknown (it depends on the balance between the flux coming from above and the flux from below, dependent on the heating prehistory), we can write the solution to inverse problem (2)–(3) in the form (Beck et al., 1985):

$$q_i = 2\sqrt{\frac{\lambda\rho c}{\pi\Delta t}}\sum_{j=1}^{i}[T_{j-1} - T_j][\sqrt{i-(j-1)} - \sqrt{i-j}], \tag{3}$$

where Δt is the time step and i is the reading number in time. Thus, we can convert a time series of temperatures measured on the surface (in K) into a heat flux through the unit surface (W m^{-2}) following, e.g., Putzig and Mellon (2007), assuming that

$$\sqrt{\lambda\rho c} = \mathrm{ATI}. \tag{4}$$

In this case $\sqrt{\lambda\rho c}$ is usually called thermal inertia, and ATI is apparent thermal inertia; it was for the first time introduced in (Price, 1985). Thermal inertial is measured in TIU (1 TIU = 1 Jm^{-2} °Cs$^{-1/2}$.

It is convenient to use ATI because we can calculate this parameter using only data of remote satellite sensing. Thus, ATI was calculated for the Mars surface (Putzig and Mellon, 2007). Researchers also try to use ATI calculations in order to reveal geological anomalies (e.g., (Nasipuri Majumdar and Mitra, 2006)). The restriction is that Eq. (4) is usually invalid under actual conditions: ATI is not equal to thermal inertial for a number of reasons. The first reason is that the brightness temperature is merely the temperature of a rather thick tropospheric layer rather than the surface temperature, and it is very difficult to determine brightness temperature long-term trends (Mears and Wentz, 2005). This difficulty can easily be avoided when measurements are performed on the ground using data of weather stations, as we perform in the present work. The second reason consists in water freezing and evaporation, which is ignored in such a simple model. A more complex model (Xue and Cracknell, 1995) makes it possible to take into account evaporation but not freezing. At the same time, direct measurements indicate that ATI, even calculated from diurnal variations, weakly depends on the water content of soil (Bennett et al., 2008). This should be particularly valid for ATI calculated by us from the seasonal variation, which takes into account deeper soil layers. Finally, the third reason is the underlying surface inhomogeneity and topography. The last two reasons cannot be taken into account by a simple model; therefore, our estimates are far from true thermal inertia, which depends on the geological structure. On the other hand, it is precisely apparent thermal inertia that can be calculated from the known amplitude of the periodic (diurnal or seasonal) variation in the surface temperature and albedo and is responsible for the formation of a heat flux of interest. Bennett et al. (2008) used similar assumptions when they calculated the global map of surface thermal inertia from diurnal temperature variations, using the soil classification code. Simple estimations of thermal inertial in the subsurface thin layer can be made from measured heat fluxes and temperatures (Wang et al., 2009). In our case we should estimate thermal inertia by using seasonal variation, which covers thicker soil layers, and data on the albedo. Such thermal inertia should be responsible for the interannual climate variability, specifically, the response of the surface temperature to the variations in solar and volcanic activity. Following (Price, 1985; Xue and Cracknell, 1995; Sobrino et al., 1998), we estimated apparent thermal inertia as

$$\mathrm{ATI} = C(\varphi)S_0\frac{1-A}{\Delta T\sqrt{\omega}}. \tag{5}$$

Here, $C(\varphi)$ is the proportionality factor, which depends on a local latitude and takes into account insolation variations at different latitudes and cloudiness; ΔT is the amplitude of seasonal temperature variations; A is surface albedo for fine days; $S_0 = 1367/4$ (W m^{-2}) is the average solar heat flux per unit surface; and $\omega = (24 \times 3600 \times 365.25)^{-1}$ is the frequency corresponding to the seasonal variation with a period of one year. From (5) it follows that the ATI measuring unit is equal to the thermal inertia unit (TIU). We do not consider here the form of function $C(\varphi)$, assuming that this function is constant within 10° latitudinal belt.

3. DATA

The average albedo value for five years on the 0.25° grid (Csiszar, 2009) was used for areas without snow. If an albedo value in the grid cell was absent, the average albedo value in the corresponding latitudinal belt was assigned to the station albedo value. The daily values of the Global Historical Climatic Network (GHCN, Correlation between the surface heat flux, averaged in a latitudinal belt, and the controlling action $F = \mathrm{TSI} + 0.5VF$

Latitudinal belt	Correlation (F, q)	The earliest year	Maximal number of weather stations in a belt
0–10	–0.11152	1880	15
10–20	0.01695	1880	35
20–30	0.10843	1913	38
30–40	0.26543	1901	67
40–50	0.4555	1880	86
50–60	0.47113	1880	44
60–70	0.43113	1908	37
70–80	0.42057	1927	4

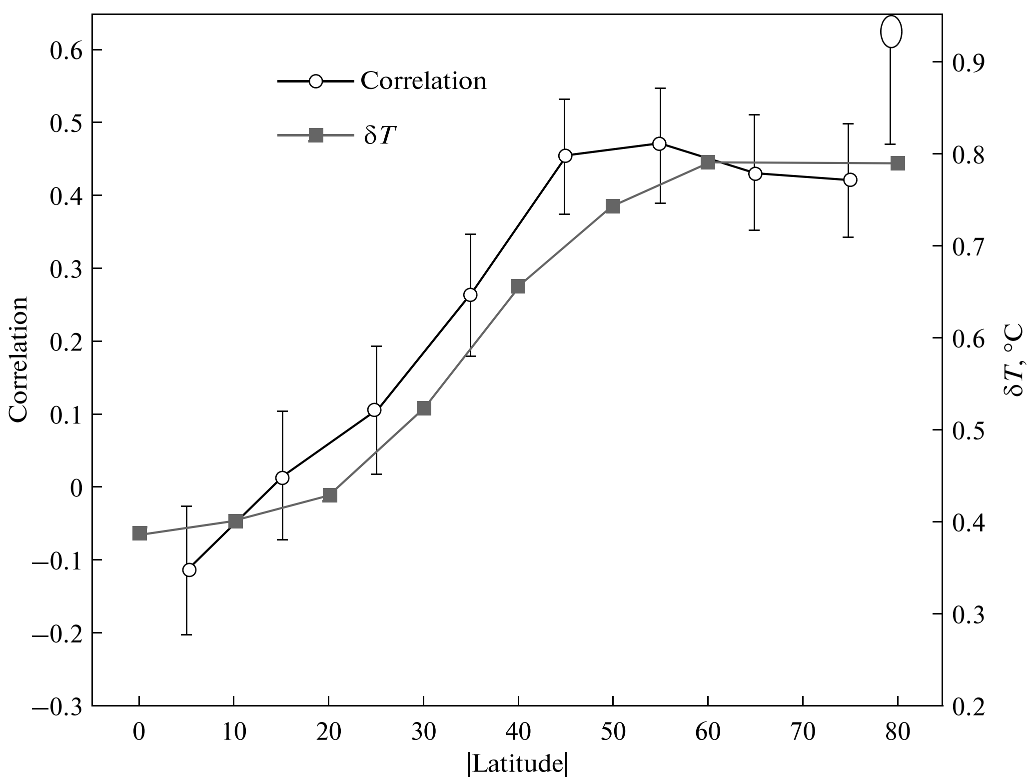
ftp://ftp.ncdc.noaa.gov/pub/data/ghcn/daily/), from which the seasonal variation amplitude was taken in order to calculate ATI using formula (5) and the average annual values were taken in order to calculate the heat flux long-term variation using formula (3), were the temperature data. The daily average temperatures averaged over a year were used to calculate the heat flux local variation. Only the observations that continued for more than ten years and had less than 20% of gaps were used in this case.

4. RESULTS AND CONCLUSIONS

We applied expression (3) to the average annual temperatures measured at each weather station in a fixed latitudinal belt. As in (Volobuev, 2013, 2014), the average annual temperatures for each weather station were approximated by a spline in order to stably calculate the derivative in formula (3). The calculated heat flux variations in a latitudinal belt were reduced to the epoch of 1990 and were averaged. The correlation between the controlling action (F) and heat fluxes, averaged in a latitudinal belt, was subsequently calculated (table). The controlling action was combined as a weighted sum from the reconstructed total solar irradiance TSI (Wang et al., 2005) and the flux absorbed by volcanic aerosol VF (Sato et al., 1993)

$$F = \mathrm{TSI} + 0.5VF, \tag{6}$$

with coefficient 0.5, which provides for the maximum correlation between the controlling action and the heat flux variation in the central Antarctic Regions (Volobuev, 2014). Volcanic forcing is converted from variations in the aerosol optical depth τ (nm) into the hindered heat flux, using the recommended scaling factor $VF = -23\tau$ (W m^{-2}) (Sato et al., 1993). The latitudinal dependence of correlation F and heat fluxes is presented in figure. The correlation coefficient confi-



The latitudinal dependence of the correlation coefficient of the controlling action on climate and vertical heat fluxes near the Earth's surface (circles) as compared to the model temperature response to the 0.1% change in the solar irradiance variations (squares). The correlation (0.63 ± 0.22) for Vostok Antarctic station is shown by a large circle (Volobuev, 2014).

dence interval in figure was estimated using the standard formula $s = \sqrt{\frac{1-r^2}{n-2}}$, where r is the correlation coefficient, and n is the sample length.

Thus, we can consider that the correlation coefficients larger than 0.2 are significant; i.e., the effect at latitudes lower than 30°–40° is absent, which is apparently related to climatic system noise and the absence of polar amplification. The second curve indicates the latitudinal dependence of the temperature response to solar irradiance variations in the 11-year cycle, which was calculated using the model presented in (Gal-Chen and Schneider, 1975) and describes the polar amplification effect.

Based on the above presentations, we can formulate the following conclusions:

• The response of climate to its natural control by means of the 11-year variations in insolation and volcanic activity has a pronounced latitudinal effect of amplification toward poles (polar amplification), which is confirmed by the observational data.

• A weak response to the control signal at low latitudes is apparently masked by climatic system noise.

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