Assessing Surface–Atmosphere Interactions Using Former Soviet Union Standard Meteorological Network Data. Part II: Cloud and Snow Cover Effects

PAVEL YA. GROISMAN

Department of Geosciences, University of Massachusetts, Amherst, Massachusetts

EUGENE L. GENIKHOVICH

Main Geophysical Observatory, St. Petersburg, Russia

RAYMOND S. BRADLEY

Department of Geosciences, University of Massachusetts, Amherst, Massachusetts

BORIS M. ILYIN

Main Geophysical Observatory, St. Petersburg, Russia

(Manuscript received 29 February 1996, in final form 2 January 1997)

ABSTRACT

Groisman and Genikhovich developed a method to obtain direct estimates of surface turbulent heat fluxes. The authors now apply it to the territory of the former Soviet Union using the 3-/6-h data of 257 stations for the past several decades to assess the sensitivity of sensible heat flux to cloud and snow cover. This property was quantified for bare soil landscapes over the entire country. During the day, the presence of clouds is associated with low values of sensible heat flux from the surface to the atmosphere. At night (and during the day in winter in high latitudes), the sign of the effect is different, but because the direction of sensible heat flux is also different (from the atmosphere to the surface), the presence of clouds again reduces the turbulent heat exchange between the bare soil and the atmosphere. The estimates of "overall cloud effect" on summer sensible heat flux are compared with similar estimates from five general circulation models to assess the abilities of these GCMs to reproduce the response of this flux to cloud cover change. Snow on the ground is associated with temperature depression. When the effect of this depression is excluded, the presence of snow on the ground is generally associated with less water vapor in the lower troposphere under clear-sky conditions, while the evaporation rate and sensible heat flux are higher than average.

1. Introduction

This paper is a continuation of empirical studies of cloud and snow cover effects (Groisman and Zhai 1994, 1995; Groisman et al. 1994a,b, 1995; Groisman and Genikhovich 1997) based on a blend of observational meteorological data for the past several decades. It employs the concept that the analysis of climate variability observed during the period of intensive instrumental observations can provide "overall estimates" of these effects.[1] The method of evaluating sensible heat fluxes from routine meteorological data was suggested by Genikhovich and Osipova (1984) and tested on different datasets by Groisman and Genikhovich (1997, hereafter GG). The complete description of the algorithm employed is presented in GG.

Surface turbulent heat fluxes, which are considered in this paper, are intensively used in hydrology and climate modeling but are not observed (or are poorly observed) by existing observational systems (Anderson 1973; Kuchment et al. 1983; Anan'ev 1986; Famiglietti

[1]By the term overall effect, we mean the summarized result of all forcings and feedbacks that are associated with the change of a given internal climate variable observed during a sufficiently long time in the climatic system. Thus, the change in cloud cover, being itself a result of the climatic system behavior, "forces" radiation and heat balance changes, which feed back to the cloud conditions. The results of these changes are by no means forcings or causal relationships. They represent the changes of the state of the climatic system with the presence of clouds or the overall effect of cloudiness that we observe and, thus, can estimate (Groisman et al. 1996a, b).

Corresponding author address: Dr. Pavel Ya. Groisman, National Climatic Data Center, 151 Patton Ave., Asheville, NC 28801.
E-mail: pgroisma@ncdc.noaa.gov

and Wood 1991; Dekic et al. 1995; Phillips 1994; Mihailovic et al. 1995). These deficiencies make verification of the output of climatic and hydrologic models difficult, which in turn affects our ability to reliably foresee the consequences of climate changes on the hydrologic cycle (Peck and Schaake 1990; IGBP 1991; Henderson-Sellers et al. 1993). Knowledge of heat fluxes at the surface and of their variability and behavior under different climate conditions (global warming, snow, and cloud cover changes, etc.) is essential in developing realistic general atmospheric circulation and/or hydrologic models to meet the needs of water resource and natural resource modelers (cf. Sellers and Hall 1992; Famiglietti et al. 1992; Robock et al. 1995; IPCC 1996). Currently, such fluxes are not well understood. Only turbulent heat fluxes (sensible and latent) at the land–atmosphere interface are discussed in this paper. Methods used for turbulent heat flux estimation are given in Rosenberg et al. (1983). Most of them require special observations not available from the standard meteorological network (e.g., fast response measurements of meteorological elements or measurements of their mean vertical profiles), or they do not provide satisfactory results when diurnal cycles and high accuracy are required (e.g., those methods based on empirical relationships between mean values of meteorological elements and their fluxes). New remote sensing methods are being intensively tested in several international experiments (the Global Energy and Water Cycle Experiment, the International Geosphere–Biosphere Programme, the Hydrologic Atmospheric Pilot Experiment in the Sahel, the Terrestrial Initiative in Global Environment Research, and the Kursk Area Hydrological–Meteorological Experiment), but, as yet, the requirements of the scientific community concerning the accuracy and reliability of these flux estimates have not been satisfied. To remedy this problem, GG suggested a method of estimating turbulent heat fluxes (primarily sensible heat fluxes) using standard meteorological network data. The standard meteorological network was not designed to measure these fluxes, and therefore our estimates are quite inaccurate at each point in a given moment of time. But at the same time, they possess the advantage of a large timescale that can be extended into the past for several decades, they provide us with clues about interannual variability and possible long-term changes, and they can be used to assess relationships with other climatic variables. Additionally, time averaging over several decades allows us to substantially reduce random errors in our estimates of long-term mean turbulent heat fluxes (GG).

The approach to the estimation of near-surface turbulent heat fluxes based on data available from the former Soviet Union (hereafter, fUSSR) standard meteorological network is given in GG. Therefore, we start with the description of the data used for estimation of turbulent heat fluxes, and then outline the concept of the overall effect for internal climate variables as it was defined by Groisman et al. (1996a) and apply it in an assessment of the changes of surface turbulent heat fluxes when cloud and/or snow cover are changing or (otherwise) are preset. Clouds and snow cover are internal components of the climatic system, and the entire system responds to changes in these components. This is the case when satellite observations (cf. Harrison et al. 1990; Sohn and Robertson 1993), as well as the output of the general circulation models (cf. Cess et al. 1991; Potter et al. 1992), are analyzed. This is also the case for our estimates of the cloud and snow cover effects described below (see also Groisman and Zhai 1995; Groisman et al. 1996a).

The cloud and snow cover effects on turbulent heat fluxes are functions of many variables, and thus, after their estimation from empirical data, they can be a subject of further analyses of the physics that governs the revealed behavior of the climatic system. These effects are typical for current climate conditions, but we anticipate that in a 2 × CO_2 climate, for example, they will be different. In the changed climate conditions, the only way to study these effects is with the appropriate GCM runs. But before such a study is possible, these effects should be reproduced by the GCMs for modern climate conditions. To secure this, we are checking our empirical estimates versus output from Atmospheric Model Intercomparison Project (AMIP)[2] runs of several GCMs (Groisman et al. 1996b).

2. Data

Figure 1 shows the stations used in the current stage of the study (223 primary and 34 supplementary stations from the fUSSR). The data characteristics are given in Table 1. In our estimation of the turbulent heat fluxes from standard meteorological observations, we use data only from the fUSSR because the observational practice there included the state of the ground code and ground surface temperature T_g that are not observed in the national networks of other countries or are not available to us in digital form. Therefore, we describe below some specifics of this network that are important for our purposes.

The primary meteorological stations in Russia (currently about 1800, but more than 4000 10 years ago) are located on dedicated plots of land (usually 30 m by 30 m), with instruments installed in a prescribed order and position. Everything that can affect measurements (e.g., trees, brush, etc.) has been removed. This uniformity has some advantages but may create problems with extrapolation to contrasting environments (e.g., heavily wooded areas).

Wind speed and direction are measured and archived

[2] AMIP was launched by the international modeling community for validation of the general circulation models (Gates 1992). The AMIP runs are GCM runs for the same 10-yr period (1979–88), with sea surface temperatures prescribed from observations.

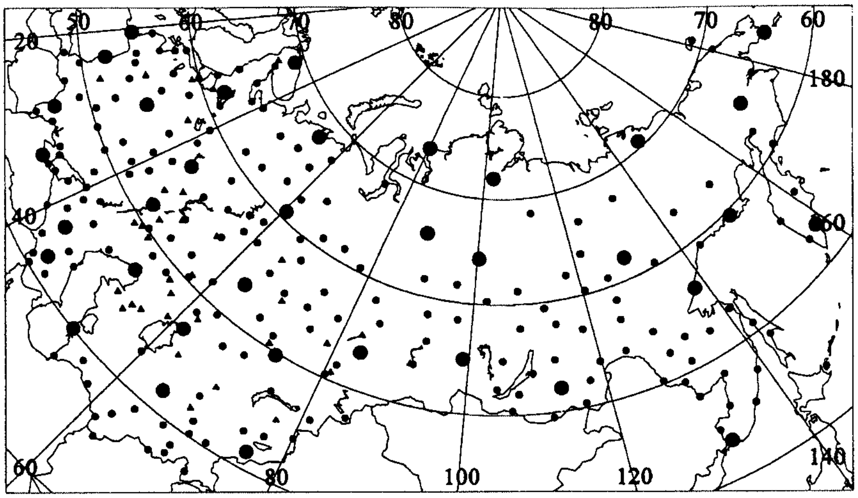

FIG. 1. Stations used in the analysis of turbulent heat fluxes over the fUSSR: primary (●) and supplementary (▲) networks. A subset of 35 stations for which data were accumulated up to 1990 and used in our first analyses (Groisman et al. 1995) are shown by enlarged dots.

at all meteorological stations of the fUSSR. Anemometers are usually installed at 10 to 12 m above the ground. In special instances with overprotected exposures, the installations are higher. Each wind datum is a 10-min average value (State Committee of the USSR on Hydrometeorology and Environmental Control 1985). The accuracy of a single measurement by standard wind gauges is about ± 0.5 m s^{-1}. Therefore, the data archived are rounded to integer values (Razuvaev et al. 1995). The accuracy of the measurements under low wind conditions is not satisfactory. All low wind speeds (less then 0.5 m s^{-1}) are ascribed the value 0, and this seriously affects the calculations of turbulent heat fluxes under unstable stratification. To resolve this problem, GG developed a special algorithm to replace the 0 value of "measured" wind speed with a nonzero value that provides an estimate of the eddy diffusivity K_1 equal to its boundary limit value, derived from their equation for K_1 with a nonzero wind speed.

A code for "the state of the ground" in the fUSSR archive provides a subjective estimate of the fraction of the meteorological site covered by snow, the type of snow, and, when there is no snow, the general estimates of the ground wetness (wet, moist, flooded, or dry) and the temperature (frozen or glazed with ice). These or similar codes are used in many countries at synoptic meteorological stations. We use this information in two ways. The information about presence of snow cover and its type gives us clues about seasonal changes of the aerodynamic roughness of the ground surface. The information about the surface wetness helps us to estimate the absolute humidity at the surface and then the vertical latent heat flux as described in GG. When the surface is wet, moist, or snow covered, we suggest that relative humidity at the surface is 100% and calculate its absolute humidity from the ground surface temperature.[3] This gives us a second measurement of humidity in addition to that at the level of the instrument shelter. Unfortunately, when the surface is dry, we cannot accurately quantify this second humidity "measurement" and have to use different hypotheses.

The accuracy of the measurements of T_g is estimated by the Russian Meteorological Service as $\pm 1°$C, but radiation errors in the mercury thermometer can bias the measurement by up to 10°C in a desert environment on sunny days (Gorodetsky 1970). These measurements are performed in controlled bare soil conditions throughout the country, and in the cases with snow on the ground, thermometers are placed partially below the surface of the snow. Instructions require observers to

[3] With ground surface temperature as used here and throughout this paper, we mean the temperature at the atmosphere–land surface boundary: bare ground or the snow surface. This element is measured at the fUSSR meteorological network.

TABLE 1. Characteristics of the Russian near-surface dataset of meteorological observations (Razuvaev et al. 1995). The map of stations is shown in Fig. 1.

Number of stations	Time increment	Period	Comments
35	3-h (6-h before 1966)	1936–90	Primary
188	3-h (6-h before 1966)	1936–84/6	Primary
34	3-h	1966–84	Supplementary

select for these measurements a 4 m by 6 m plot of land that is free of grass and nonshadowed to cultivate it every spring (after snow cover has melted) to the depth of 25–30 cm and keep the plot flat and without dry soil crust during the entire warm season. On this plot, three ground surface mercury thermometers (standard, minimum, and maximum) are placed with their reservoirs directed eastward. Although the observer is required to report the reading of these thermometers to 0.1°C, the increments of the instrument scale are only 0.5°C, and in the digital archives, this element is rounded to integer values of degrees Celcius. Thus, the T_g measurements may significantly differ from the ground surface temperatures of the surrounding environment, which can be, for example, grassland. Generally, we would not expect that these measurements would be close to the "active layer" surface temperatures derived from satellite and/or airborne remote sensors, which measure the "skin" temperature, with the possible exception of a flat bare soil terrain (Gol'tzberg 1969). However, it appears that for the estimation of turbulent heat fluxes the T_g measurements can be used (Budyko 1948; Genikhovich and Osipova 1984).

To evaluate the accuracy and biases in the standard T_g measurements at the network, Zubenok (1947) performed an elaborate test using as a "ground truth surface temperature" the records of temperature derived from a platinum wire net placed on the ground near the installation of mercury thermometers. Observations were performed on two different soil types in daytime summer conditions, when the measurement biases of the mercury thermometer are most pronounced, together with other meteorological parameters (the air temperature, T_b, profile in the lower 1.5 m and the wind speed). She derived an adjustment of the readings of mercury thermometers to the "ground surface true temperature" for unstable stratification of the form

$$T_g(\text{"true"}) = BT_g(\text{measured}) - (B - 1)T_b(\text{measured at 1.5 m}), \qquad (1)$$

where the value of B varies from 1.4 (sand) to 1.56 (black soil). We use this formula to recalculate T_g and evaluate ground surface saturation humidity. In the worst cases (in a desert environment in the afternoon), these adjustments still provide reasonable accuracy in the mean monthly turbulent heat fluxes (see the appendix and GG).

3. Concept of the overall effect for internal climate variables

One way to evaluate the parameterization of meteorological elements is to compare the modeled statistical structure (mean, standard deviation, and spatial and temporal correlation) with empirical data from the world climate observational system (Weare and Mokhov 1995; Stouffer et al. 1994; Del Genio et al. 1996). Groisman et al. (1996a,b) suggested a further step in this assessment. They consider the parameterization of the meteorological element in the model adequate if, together with a proper reproduction of the mean and variability of this element, the model properly reproduces the internal relationships between this element and other climatic variables. For example, parameterization of cloud cover should lead to a proper change in the temperature field when clouds are changing, etc.

The statistics of overall effect (OE) and "partial overall effect" (POE) are simple and easy to interpret. Groisman et al. (1996a) suggested using the following statistics for a set of elements "f" to check the proper parameterization of the element "y" using empirical data:

$$\mathrm{OE}(f \mid y \in D) = E(f) - E(f \mid y \in D), \qquad (2)$$

where E is a mathematical expectation and $E(\ \mid\ )$ is a conditional expectation. To study the proper parameterization of the element "z" when several other internal factors are involved, the following statistics can be used:

$$\begin{aligned}\mathrm{POE}(f|z|x \in C,\ y \in D) &= E(f \mid z = a,\ y \in D,\ x \in C) \\ &\quad - E(f \mid z = b,\ y \in D,\ x \in C). \qquad (3)\end{aligned}$$

First examples of the use of these statistics were presented by Groisman and Zhai (1995) and Groisman et al. (1996a,b). The further applications of Eqs. (2) and (3) are presented in the next sections of this paper.

Observations under clear skies play a key role in contemporary studies of cloud and snow cover feedbacks and forcings (cf. Harrison et al. 1990; Cess et al. 1991; Potter et al. 1992; Gaffen and Elliott 1993; Groisman et al. 1994; Chanine 1995). Comparison of climate conditions under clear skies with average conditions [i.e., statistics described by Eq. (2)] gives estimates of overall cloud effect. Sorting by equal ranges of surface temperature with preselected cloud conditions provides a dataset for studying the overall effects of snow-covered versus bare ground conditions on other meteorological elements (Groisman and Zhai 1994, 1995; Groisman et al. 1996a).

4. Overall cloud effect on the surface turbulent heat fluxes

Figures 2 through 7 give our pilot estimates of *overall cloud effect* (OCE) on the surface sensible heat fluxes f over the fUSSR territory. We apply statistics $\mathrm{OE}(f \mid y \in D)$ as a main research tool. Estimates pertain only to the territory of the fUSSR and were calculated as a difference between an average flux and an average flux under clear-sky conditions. The preliminary estimates of these effects presented earlier by Groisman et al. (1995) corresponded to a coarse resolution of the network selected at that stage of analysis. Now we apply our estimates to the data of another 200 fUSSR stations from the archive (Razuvaev et al. 1995), expand the

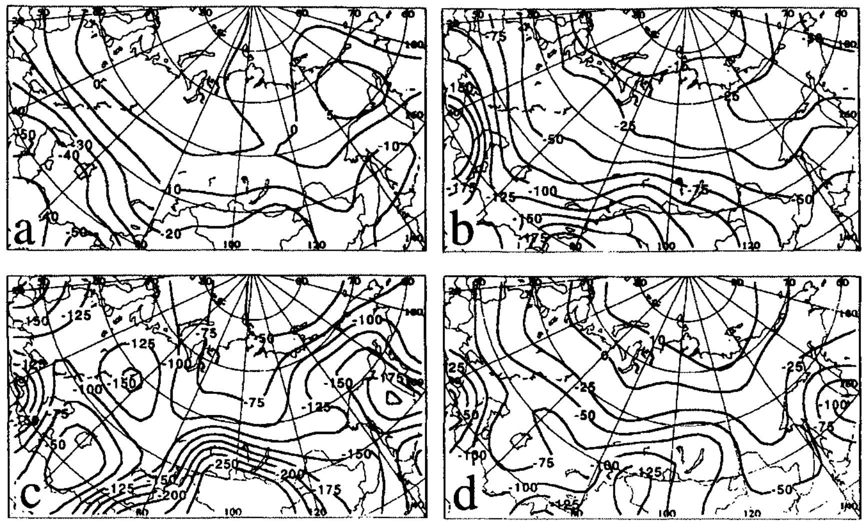

FIG. 2. Pilot estimates of overall cloud effect (W m^{-2}) on the afternoon sensible heat flux in (a) January, (b) April, (c) July, and (d) October for $f_{avg} - f_{clear\,sky}$. Negative values indicate that near-surface heat flux under cloud cover is less than that under clear skies.

period of time averaging in the past back to 1936, and present our most recent version of these estimates. We, nevertheless, designate them as pilot (or preliminary) estimates because they represent only one type of landscape—bare soil.

The fUSSR spans 11 time zones and while constructing the maps presented in Figs. 2 and 3, we chose the time closest to 1500 LT, (afternoon) and 0300 LT (night) for each station. The initial field of the OCE estimates from the stations was interpolated onto a coarse grid using a kriging procedure and then smoothed. Since, at this stage, we are using a rather dense network of stations [257 stations instead of 69 and 35, respectively, in Groisman and Zhai (1995) and Groisman et al. (1995)], a simple kernel smoothing (without kriging) delivers similar results. The effects of this smoothing, as well as of using different sets of stations (i.e., 35, 69, and all 257 stations) instead of a sparser network, appear to be of minor importance. The mapped fields of OCE are based on meteorological station data and thus represent only one type of landscape, or the landscape of standard meteorological ground: leveled terrain with bare soil and (maybe) short grass. Further studies are required to transform these fields to fit other landscapes and, therefore, to characterize OCE over forest zone and rough terrain.

Figure 2 shows our pilot OCE estimates on the afternoon sensible heat flux in January, April, July, and October. This OCE is mostly negative and indicates that the presence of clouds is associated with sensible heat

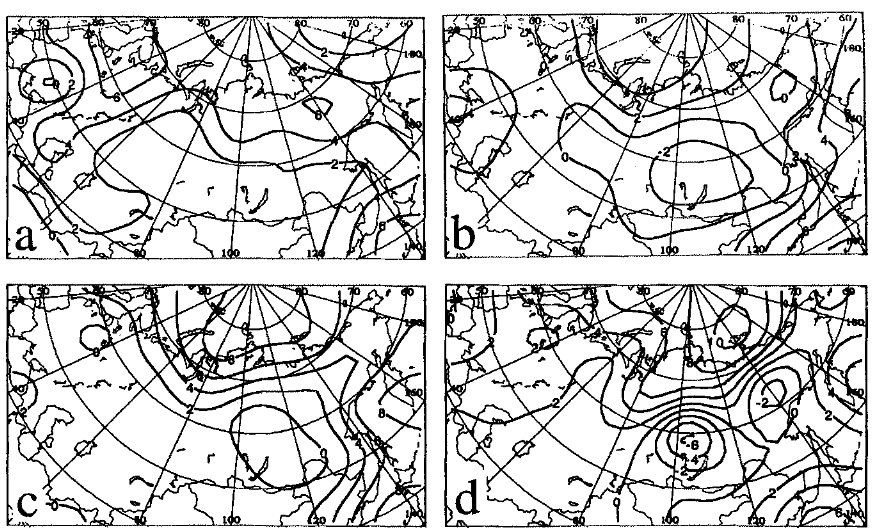

FIG. 3. The same as Fig. 2 but for night sensible heat flux.

fluxes less than those under clear-sky conditions. The extremely high differences of sensible heat fluxes between average and clear-sky conditions along the southern border of the fUSSR are suspicious. These regions are mountainous, and the legitimacy of spatial extrapolation of our estimates here is questionable. In the warm season, we obtain a mostly zonal pattern of this effect: its magnitude increases from north to south in the range -50 to -150 W m^{-2}. But over the deserts and semideserts of central Asia in summer (July), the distinction between clear-sky conditions and average conditions (which include more than 70% of days with clear skies) diminishes. This is also seen in the decrease of cloud effect in this region to -50 W m^{-2} or less (Fig. 2c). The uncertainty in sensible heat flux estimates (related to different methods of handling the bias in the ground surface temperature) does not affect the pattern of the daytime OCE on these fluxes and changes its amplitude little. So replacement of the B value in Eq. (1) from 1.4 (sand soil) to 1.56 (black soil) in the dry steppe environment (Saratov, Russia, midday in July, with clear skies) increased the ground surface temperature by ~3 K but changed the single values of sensible heat fluxes by approximately 15%.

At night (Fig. 3) and in winter over northern Eurasia, there is a weak positive cloud effect on sensible heat fluxes up to 5 W m^{-2}. This is correlated to the nighttime (including polar night) cloud "warming" of the surface (Groisman et al. 1996a) and to strong inferences between this warming and surface turbulent heat fluxes.

During the day, sensible heat flux is mostly directed upward and the negative values of OCE indicate that average values of sensible heat flux are less than those for clear-sky conditions. At night, this flux is mostly directed downward, and the positive values of OCE again indicate that the absolute values of turbulent heat exchange between the atmosphere and land surface are weaker with clouds than under clear skies. Thus, in almost all situations, the presence of clouds is associated with a decrease of the exchange of sensible heat between bare soil and the atmosphere. This in turn affects convective processes and cloudiness itself.

The variability of the daytime sensible heat fluxes in the warm period is enormous (Table 2). Therefore, the estimates presented in Fig. 2 may be inaccurate when the sample size (i.e., the number of clear-sky observations) is not sufficient. We select the worst case (July afternoon with turbulent heat fluxes) and assess in Fig. 4 the accuracy of determination of the sign of the overall cloud effect on these fluxes. We intentionally used only half of the period of observations (1966–90) with a subset of 35 stations that were well distributed over the fUSSR territory. Student's t scores plotted in this figure show that the sign of the cloud effect on the sensible heat flux is defined well beyond the 99% level of confidence (t scores are above or near 3, with the effective number of degrees of freedom varying from 500 in the desert to 20 in the forest zone). Nevertheless, the confidence intervals are quite wide. We can reduce this uncertainty by (a) adding more points and spatially averaging them over appropriate grid cells, thereby reducing the variance of the estimates and increasing the t scores (this is a standard way to handle the situation, and the complete network of 257 stations provides us with this opportunity), and (b) introducing a less strict definition of clear-sky conditions, such as adding the partly cloudy observations with a cloud cover of 1 octa to "clear skies" (we did this in the next section when we assessed the performance of five GCM AMIP runs in the characterization of July OCE on sensible heat flux). The main reason for the low t scores is the small sample size: clear-sky summer afternoon conditions in high latitudes are quite rare (Fig. 4b; see also Gaffen and Elliott 1993; Warren et al. 1986; Hahn et al. 1994). The situation, however, stays manageable due to the long period of standard meteorological observations, which allows time averaging. Table 3 contains the estimates of standard deviations of the differences shown in Figs. 2 and 3. The values of these estimates are affected by the variance of sensible heat fluxes and by the number of situations with clear skies during the period of observations. In the forest and tundra zones, high values of these standard deviations in summer afternoons are related to a small number of the clear-sky situations, while in desert and steppe zones, high variability of afternoon heat fluxes is responsible for the magnitude of these values.

We would like to be able to assess the changes in time of turbulent heat fluxes for the entire period of data availability. Since the mid-1930s, synoptic observations in the fUSSR have been located mostly in airports—thus, in locations open to wind. This and a strict standard observational program determine the unaltered system of meteorological elements that is used for the evaluation of sensible heat fluxes in our method for the past several decades. However, there is one problem that still requires special attention. In 1966, the Soviet meteorological service converted its observational program at the primary network from the 6-h time step to a 3-h time step. But, at the same time, the hours of observations were shifted to be in line with Greenwich Mean Time. This introduces a homogeneity problem into the time series of each element that has a diurnal cycle. These inhomogeneities were assessed for temperature, cloudiness, etc., but not for sensible heat fluxes. In the summer season, these fluxes change by two orders of magnitude, and a minor shift (or change in the method of time averaging) may introduce a bias that will affect the conclusions about possible trends. Patterns of our estimates of the differences between average and clear-sky conditions for two periods before and after 1966 (Fig. 5) stay intact, but the magnitude of our estimates appear to be affected by this change. We observe an increase of our

TABLE 2. Examples of diurnal and seasonal changes of sensible heat fluxes over bare soil in different climatic zones of the fUSSR (W m^{-2}). Mean and standard deviations for the 25-yr period 1966–90.

	0000 LT		0300 LT		0600 LT		0900 LT		1200 LT		1500 LT		1800 LT		2100 LT	
Month	Avg.	Std.	Avg.	Std.	Avg.	Std.	Avg.	Std.	Avg.	Std.	Avg.	Std.	Avg.	Std.	Avg.	Std.
Tundra zone, Nar'yan-Mar, 67.7°N, 53.1°E																
1	−6	17	−7	14	−7	14	−6	19	−7	13	−8	16	−7	15	−6	15
2	−9	21	−9	14	−8	14	−6	13	0	14	−8	13	−10	13	−9	14
3	−11	14	−10	13	−9	11	3	18	16	31	2	18	−12	14	−12	13
4	−6	15	−6	13	0	16	29	47	43	62	25	46	−2	20	−8	17
5	−2	13	0	14	12	24	44	59	64	81	51	71	17	39	0	16
6	−2	9	0	9	17	22	78	81	124	112	107	87	40	43	0	11
7	−2	9	−1	8	16	22	79	79	129	119	114	98	46	60	0	11
8	−4	7	−3	6	5	10	43	55	77	90	61	66	13	25	−3	8
9	−6	7	−5	7	−3	7	10	18	29	39	15	30	−5	8	−6	7
10	−4	13	−3	21	−3	13	0	15	1	17	−4	13	−6	12	−5	16
11	−3	17	−2	26	−2	17	−3	16	−2	15	−4	16	−3	16	−3	16
12	−5	19	−4	20	−5	15	−5	20	−5	22	−7	13	−5	16	−5	14
Forest zone, Nizhnii Novgorod, 56.3°N, 44.0°E																
1	−4	11	−3	11	−2	13	−2	12	7	16	−1	11	−5	10	−4	13
2	−7	11	−5	13	−5	9	−2	9	20	33	3	16	−9	12	−9	12
3	−9	10	−7	10	−7	9	2	13	19	46	3	20	−10	12	−10	13
4	−9	10	−8	9	−5	9	18	32	54	70	41	74	−3	14	−9	12
5	−7	8	−6	8	0	8	84	75	158	148	123	103	25	43	−6	8
6	−4	7	−4	6	5	9	113	107	184	150	149	133	46	57	−2	7
7	−3	6	−3	5	3	6	95	73	161	123	138	113	42	40	−2	6
8	−3	16	−3	6	0	5	65	59	129	105	108	96	19	26	−3	6
9	−4	6	−3	6	−2	6	23	30	67	67	43	49	0	8	−4	7
10	−4	9	−3	9	−2	8	4	12	20	41	6	23	−5	9	−5	10
11	−1	11	−1	11	0	13	1	15	5	17	−1	12	−3	10	−2	10
12	0	15	0	15	0	17	0	13	7	25	0	18	−1	13	−1	13
Steppe zone, Simpheropol, 44.9°N, 34.2°E																
1	−7	17	−7	17	−7	17	−6	18	21	46	27	42	−2	17	−6	16
2	−7	15	−6	16	−7	17	−2	17	54	70	66	73	2	19	−4	16
3	−8	13	−8	14	−9	14	7	20	144	177	145	131	19	25	−6	13
4	−8	12	−10	14	−11	13	48	68	236	242	227	178	60	53	−3	12
5	−4	10	−6	10	−6	9	117	126	328	290	274	168	100	74	0	9
6	−4	7	−6	7	−5	8	143	143	364	333	301	183	128	80	2	12
7	−4	6	−6	6	−5	6	119	111	333	286	306	208	125	70	0	8
8	−5	6	−7	7	−8	7	78	78	302	257	265	137	89	50	−2	9
9	−7	7	−9	8	−9	8	30	32	246	210	225	172	40	35	−5	9
10	−9	9	−10	11	−10	11	4	14	153	163	143	147	0	14	−7	9
11	−10	13	−11	14	−10	13	−6	14	52	83	43	51	−5	15	−8	13
12	−8	14	−9	15	−8	15	−6	16	12	28	13	28	−5	13	−7	13
Desert, Tamdy, 41.7°N, 64.6°E (+1-h time shift, i.e., 1100 instead of 1200 LT)																
1	−6	16	−4	18	−4	20	−3	19	23	45	60	94	4	25	−4	23
2	−6	21	−6	21	−5	25	−5	21	48	78	103	118	20	46	−5	19
3	−8	13	−7	12	−7	11	−2	13	136	180	212	240	72	91	−6	15
4	−10	11	−10	12	−8	21	5	20	187	178	268	259	118	151	−6	20
5	−9	15	−8	14	−9	11	38	58	295	242	420	381	214	191	−1	27
6	−9	9	−10	9	−9	9	44	48	336	271	458	335	262	195	2	14
7	−9	10	−9	9	−8	19	32	45	320	259	430	323	249	194	3	21
8	−10	11	−9	12	−8	14	13	29	281	214	364	219	204	132	−4	21
9	−12	11	−11	10	−10	10	−2	13	244	218	340	243	141	115	−3	33
10	−12	12	−10	11	−10	15	−7	19	161	185	239	231	48	65	−9	12
11	−11	13	−10	16	−10	11	−8	13	50	71	106	125	1	21	−9	27
12	−6	16	−5	18	−4	22	−5	15	16	32	42	58	0	22	−5	14

OCE estimates by up to 50% between these two periods. Unfortunately, before any statement about the effects of "climatic change" on this OCE can be made, a thorough analysis of the diurnal cycle of the sensible heat flux over bare soil should be performed, and its contribution to the average from two sampling strategies (eight times per day and four times per day) should be assessed. Our comparison of average sensible heat fluxes for the periods before and after 1966 reveals statistically significant differences in mean daily fluxes in the southern part of the fUSSR (Fig. 6)—that is, in the regions where the potential bias mentioned above could be especially pronounced. This issue requires additional studies.

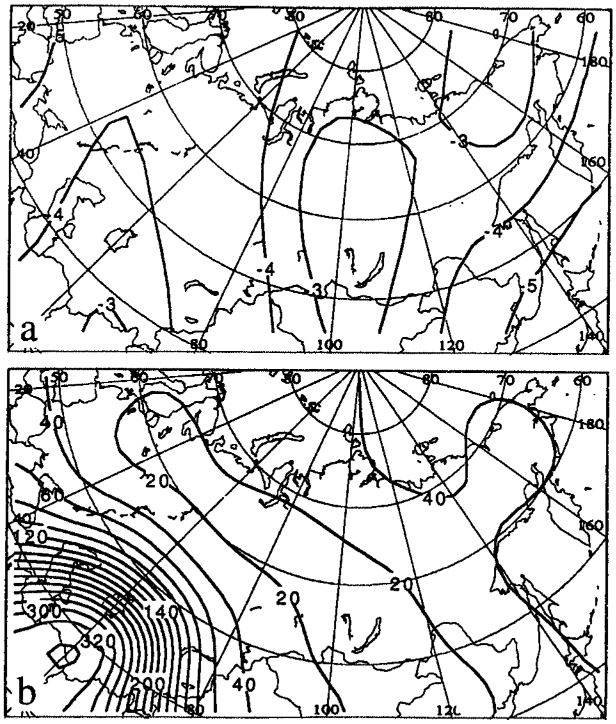

FIG. 4. (a) Student's t scores of the estimates of overall cloud effect on July afternoon sensible heat fluxes; estimates are based on 35 primary stations well distributed over the territory of the fUSSR (enlarged dots in Fig. 1). (b) The number of clear-sky afternoon observations in July from 1966 to 1990.

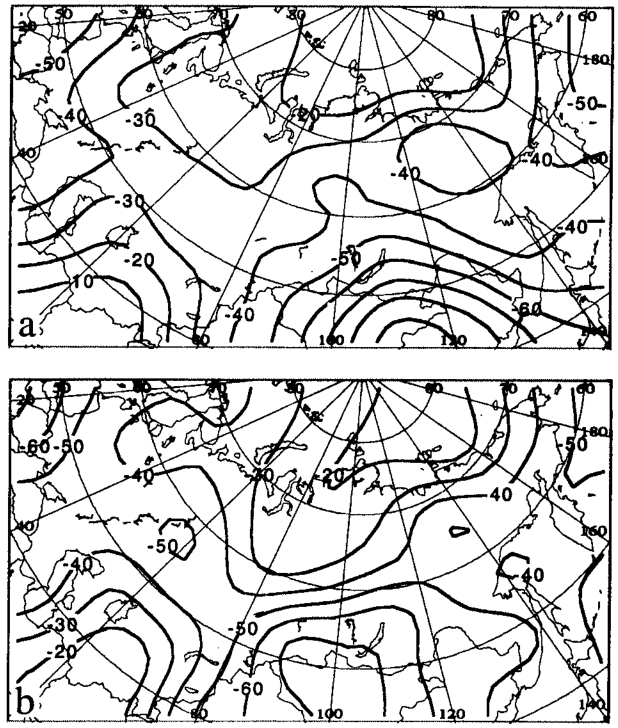

FIG. 5. Mean July estimates of the overall cloud effect on sensible heat flux (W m^{-2}) from the data of primary meteorological stations for (a) 1959 to 1965 and (b) 1966 to 1985.

Performance of five general circulation models in reproducing the mean July OCE on sensible heat flux

In this section we show the comparison of our estimates of OCE on sensible heat fluxes with similar estimates derived from five AMIP runs of the GCMs developed in the Canadian Climate Centre (CCC), the Goddard Institute for Space Studies (GISS), the National Centers for Environmental Prediction [NCEP, formerly the National Meteorological Center (NMC)], the University of Illinois at Urbana–Champaign (UIUC), and the Max-Planck Institute (MPI). The characteristics of these models are available in Phillips (1994). We received the output of these 10-yr (14 yr for NCEP[4]) runs with daily (GISS, NCEP, and UIUC), 6- (CCC), and 12- (MPI) hourly resolutions. In all five cases, we analyzed jointly total cloudiness and sensible heat flux fields and constructed mean monthly average fluxes for all cases and for clear-sky conditions only. Taking into account the short period of the AMIP run (10 or 14 yr) and spatially averaged output of the GCMs (i.e., model grid cells), we use a weakened condition in the definition of clear skies to include total cloudiness of less than 0.15. We made appropriate adjustments in our estimates of cloud effect on mean July sensible heat flux over the former USSR, but the results remain essentially unchanged.

Figure 7 presents the estimates of OCE on the mean July sensible heat flux from these five GCMs and our estimates. Generalized characteristics of accuracy (representative standard deviations in W m^{-2}) of these OCE estimates are shown in Table 4. Standard deviations of

[4] The NCEP AMIP run is a 14-year-long GCM run for the period from 1982 to 1995.

TABLE 3. Generalized characteristics of accuracy (representative standard deviations, W m^{-2}) of the estimates of OCE on the sensible heat fluxes over the fUSSR territory, derived from empirical data shown in Figs. 2 and 3. Standard deviations of mean differences $f_{avg} - f_{clear\ sky}$ were estimated from appropriate time series of these fluxes at each point.

	January		April		July		October	
	Afternoon	Night	Afternoon	Night	Afternoon	Night	Afternoon	Night
Tundra	2	1.5	5	2	15	4	2	4
Forest	1	1	10	2	20	0.7	4	1.5
Steppe	2	1	7	~0.5	30	~0.5	8	1
Desert	7	1	30	~0.5	15	0.2	15	0.2

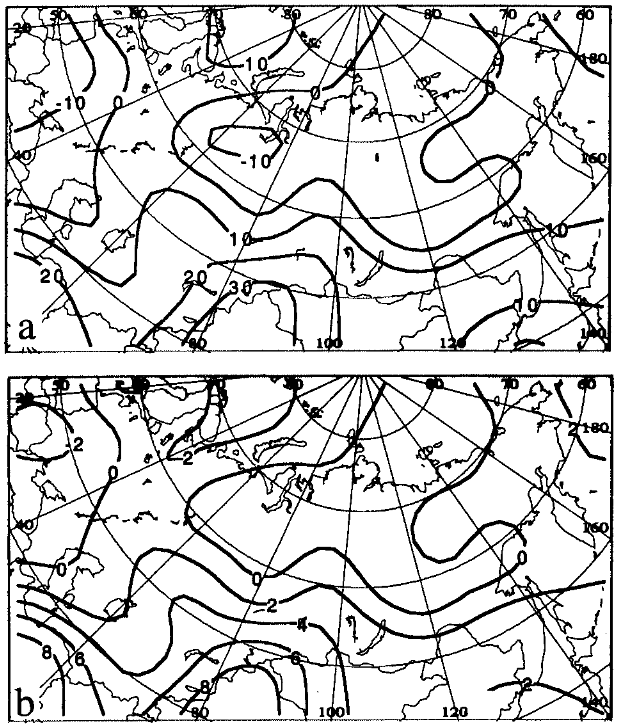

FIG. 6. (a) Difference (percentage of the mean value for period 1966–85) between mean July sensible heat flux estimates from the data of primary meteorological stations for 1966 to 1985 and 1959 to 1965 and (b) *t* scores of this difference (values above 2 corresponds to a two-tail criterion that the difference is statistically significant at the 95% level). To provide compatibility of daily mean flux estimates, we removed from the former estimates intermediate measurements every 3 h and left only the measurements made with 6-h time step, which was used before 1966. Nevertheless, the compatibility is not secured for the time zones 3, 5, 6, 8, 9, 11, and 12 because of the 1-h time shift in the observation schedule in 1966.

mean differences $f_{avg} - f_{clear\,sky}$ were estimated from appropriate time series of these fluxes at each point or grid cell. Thus, (a) they are conservative estimates of this accuracy because the spatial correlation of estimates is not taken into account, but (b) they do not incorporate biases in the methods of evaluation of the sensible heat fluxes in the GCMs (as well as in our method), which can be different for clear-sky and average climate conditions. GCM output provides grid cell average fluxes and empirical data provide point estimates of these fluxes. Analysis of the spatial correlation function of monthly sensible heat flux time series indicates that more than a half of the variance of these time series has local origin—that is, is not related to processes beyond a 50-km radius (Fig. 8). This function and the field of point variance of sensible heat fluxes can be used as an input for calculation of the variance of the area mean sensible heat fluxes (over the grid cell) that will be comparable to those in GCM outputs (Kagan 1979). So far, this was not done. As a result, summer variability of the GCM-derived sensible heat fluxes is two (in tundra) to five (in desert and steppe climatic zones) times less than our empirical *point* estimates shown in Table 2. This effect also contributes to a "superior" OCE accuracy of the GCM-derived estimates in Table 4.

The patterns of the OCE estimates on the mean July sensible heat flux shown in Fig. 7 are similar, but the magnitude of empirical estimates *for bare soil landscape* is at least twice as high as corresponding GCM estimates for average (mixed) landscapes over the fUSSR territory for CCC, GISS, and MPI GCMs. Point estimates of the accuracy of this OCE generalized in Table 4 often indicate that the differences between the OCE estimates from observations and the output of these three GCMs are statistically significant. For the NCEP and UIUC AMIP runs, the pattern and amplitude of the OCE are very close and do not differ statistically significantly from our estimates. Until we have a proper scale for our estimates of sensible heat fluxes, we cannot critically evaluate the performance of the GCMs in the reproduction of these fluxes. However, the sign and pattern of OCE in these five GCMs that conform to our empirical estimates indicate that the internal relationships in these models, which are responsible for cloud effects on the surface turbulent heat fluxes, work properly. Coincidentally, in desert, dry steppe, and tundra climate zones, where we believe that our estimates of sensible heat fluxes are close to area-averaged values, the magnitudes of the OCE on these fluxes estimated from the output of the GCMs and by our method are the same.

5. Snow cover "effects" on surface atmosphere interactions under clear skies

More than 100 years ago, Voeikov (1889) estimated an average reduction of surface air temperature associated with the presence of snow on the ground (−6 to −5 K). His findings were confirmed several times with new data (Lamb 1955; Pokrovskaya et al. 1965; Wagner 1973; Namias 1985; Leathers et al. 1995; Figs. 9 and 10), but the general approach did not change. The problem with this approach is that we cannot establish whether snow cover causes a temperature decrease (though it contributes to it) while implying that a temperature decrease is a cause of establishing permanent snow cover (although low temperatures are a necessary condition for this). For example, in Fig. 10, we do not see any effects of snow on the ground in winter 10-day periods, when over the fUSSR, the absence of snow cover does not necessarily mean *warm* weather, but may correspond to anticyclonic cold conditions over the steppe and deserts of central Asia, Kazakhstan, and southern Siberia. In spring, however, snow cover is a standard state of the ground over the entire country. Therefore, the absence of snow cover corresponds primarily to thaws, enhanced by a higher portion of absorbed radiation than with snow on the ground and, thus, to higher upward daytime surface turbulent heat fluxes. From Fig. 10, we may conclude that in spring snow cover "suppresses" turbulent heat exchange between the atmosphere and the surface. But in fact, temperature (and, in late spring, radiation) contributions dominate the snow cover effect on turbulent heat fluxes in this season. Some GCMs indicate that snow on the ground is associated with more stable atmospheric conditions and, thus, the troposphere gets less water vapor from the surface (Cess et al. 1991; Cohen and Rind 1991). The results obtained during the construction of Fig. 10 (not shown) support this statement for intermediate seasons.

The decrease of temperature with snow on the ground is so strong that all other effects of snow (including the difference in the heat fluxes) are mostly derivatives of this 5-K cooling. We believe that it is time now to further our understanding of this established relationship (cf. Figs. 9 and 10) to reveal the impact of snow on the ground on other meteorological elements *after* the effect of a temperature decrease is taken into account. Snow has properties that affect other meteorological elements: for example, it is an additional source of water vapor and produces a reduced surface roughness. These properties have a signature that is separate from the radiative effects of snow. The effects of these properties can be estimated when we "remove" the effect of a temperature decrease. Below, we present the diagnostic statistics that give information about the magnitude of the snow cover impact on the turbulent heat fluxes when the temperature contribution is eliminated/reduced. To achieve this goal, we selected in our analyses the data from the same surface temperature ranges R and then assessed the effect of snow cover on the turbulent heat fluxes separately within these ranges.

Because the effect of surface conditions on turbulent heat fluxes are greater under clear skies, we used the clear-sky conditions to delimit the possible effects of the presence of snow on the ground on these fluxes. We applied POE ($f|z|y \in D,\ T \in R$) statistics as a main research tool. To obtain the values of this POE, the estimates of heat fluxes at each primary station were sorted by surface air and/or ground temperature, cloudiness, and the presence of snow on the ground. Only those temperature ranges R (± 0.5 K) were selected for which both surface conditions (snow and no snow) were observed under clear skies at least 10 times during the analysis period. The difference ($f_{\mathrm{snow}} - f_{\mathrm{bare}}$) was then calculated for each selected temperature range R and for each station. Only saturated ground conditions were used for latent heat fluxes. The average values of these differences for temperatures below freezing are presented in Table 5. To obtain the values in Table 5, (a) estimates of turbulent heat fluxes (based on 3- and 6-h data of 223 stations, well distributed throughout the former USSR, for the past 50 yr) were sorted by *ground surface* temperature (in the ranges ± 0.5 K) and the presence of snow on the ground (we also considered separately the case for clear-sky conditions); (b) the average fluxes were calculated in each group separately, but we included in the analysis only those temperature ranges for which both snow and no-snow conditions occurred at least 10 times during the past 50 yr; and (c) the differences between turbulent fluxes in each range for snow–no snow conditions were arithmetically averaged. A total of about 10^6 observations over bare soil and 2×10^6 observations over snow-covered land from 223 stations for the period 1936–84 qualified to participate in this analysis. Because we excluded dry bare soil conditions from our analysis of latent heat flux sensitivity to the presence of snow on the ground, our sample size for latent heat flux over bare soil is 20% less than that for sensible heat flux. Accuracy (standard deviations) of these estimates varies from less than 0.2 W m^{-2} for average latent heat flux to 1 W m^{-2} for sensible heat flux under clear-sky conditions. The distribution of these differences with surface temperatures is shown in Fig. 11. This figure indicates that the largest differences are when the ground surface temperature is close to the freezing point and that the *signs* of partial snow cover effects (i.e., when temperature contribution was excluded) become different from those in Fig. 10.

Groisman et al. (1995) show that the presence of snow on the ground is associated with a water vapor decrease in the lower troposphere under clear-sky conditions for North America and for most of Eurasia. They found a 13% average decrease of water vapor content for the fUSSR when snow was on the ground as compared to bare soil conditions. Sorting by equal ground surface temperatures in the fUSSR refines and strengthens this effect to a 22% average decrease (Table 6). Note that this *decrease* of atmospheric water vapor pressure over snow-covered soil for the same surface temperatures corresponds to *an increase* of evaporation rates. We estimate the average increase of latent heat flux over saturated soil for *the same negative surface temperatures* to be about 6 W m^{-2} (but it is close to 20 W m^{-2} for clear-sky conditions) when snow is on the ground (Table 5, see also Fig. 11). Values presented in Table 5 indicate that the evaporation rate from snow cover is higher than that from bare soil for clear skies, as well as for average climate conditions, when temperature effects are "factored out." The removal of the "temperature factor" allows us to estimate the weak effects on turbulent heat fluxes of other snow properties, such as its reduced roughness and, related to this roughness, the increase of near-surface winds. The increase of turbulent heat fluxes with snow on the ground shown in Fig. 11 allows the following interpretation: this is a manifestation of the near-surface wind transformation by snow cover. In fact, we checked directly the changes in the wind field with snow on the ground. When the temperature effects are factored out, wind speed increases by 20% to 40% with the presence of snow on the ground.

6. Conclusions

Using the 3-/6-h data of 257 stations for the past several decades and the method developed by Groisman

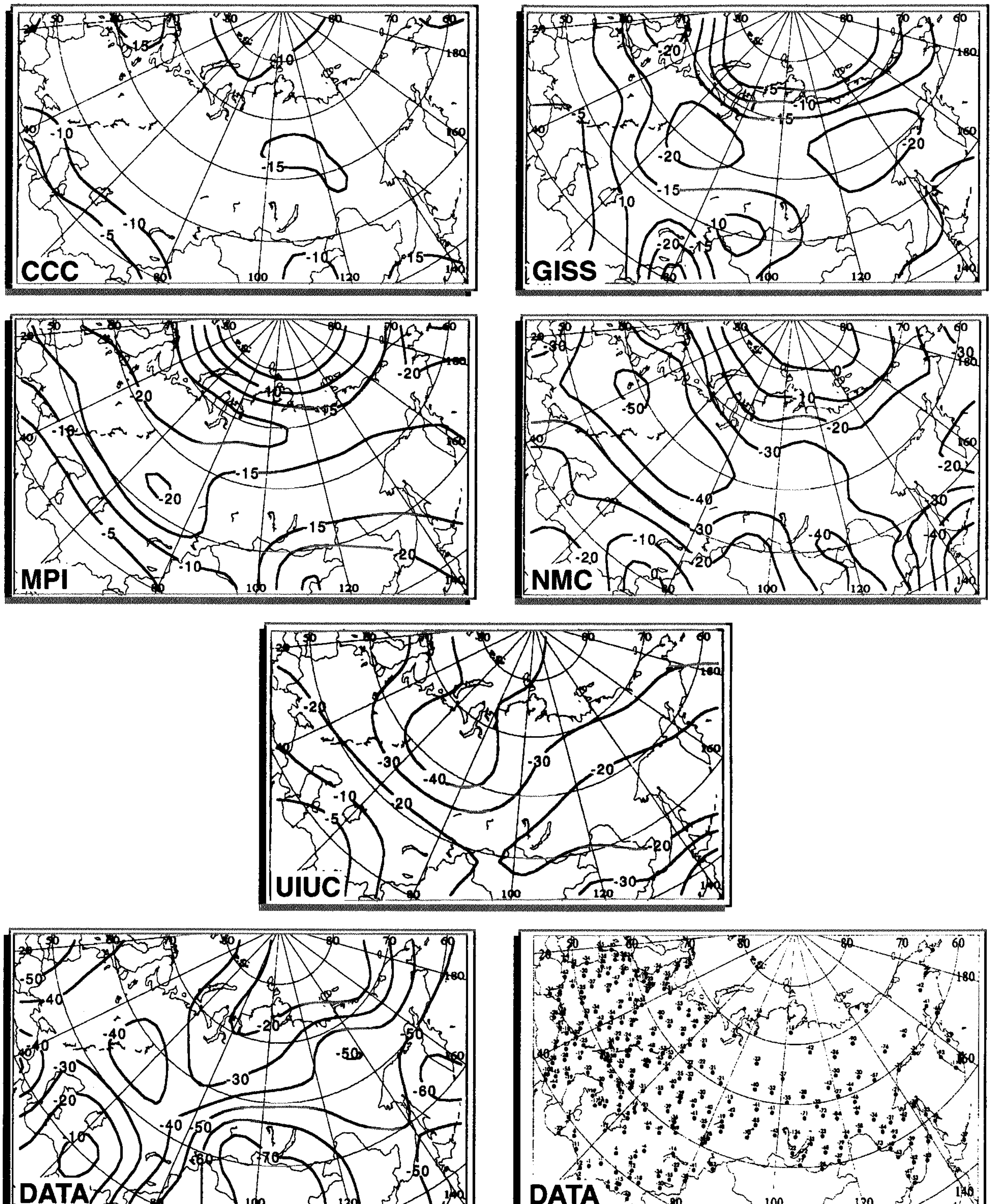

FIG. 7. Surface sensible heat flux overall cloud effect as estimated from the CCC, GISS, MPI, NCEP/NMC, and UIUC AMIP model runs over land areas only (upper three rows of maps and the enlarged map on the next page) and from empirical data (bottom row of maps) for mean July averages. Estimates of OCE are equal to $f_{\mathrm{avg}} - f_{\mathrm{clear\,sky}}$. Flux direction from the surface up is considered positive. A "clear sky" day is defined as a day when total cloudiness was less than 0.15.

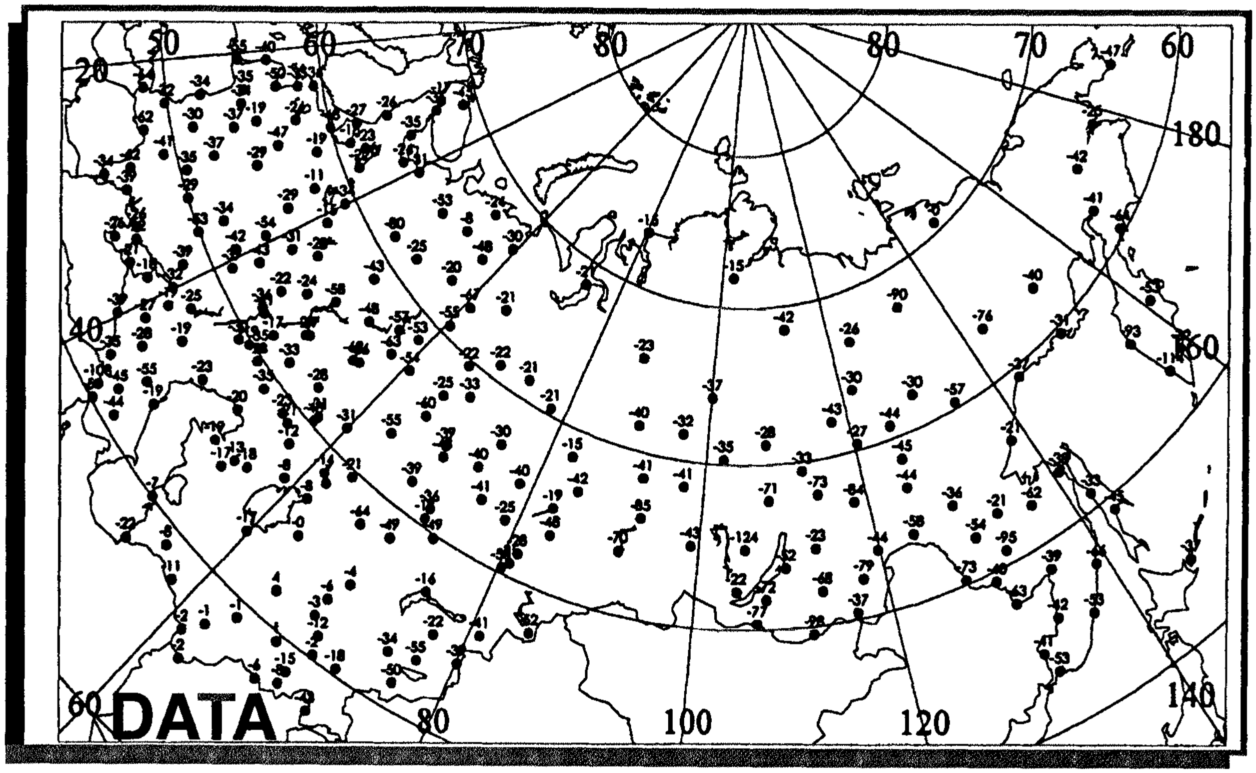

FIG. 7. (*Continued*)

and Genikhovich (1997), the sensitivity of sensible heat flux to cloud cover has been quantified for bare soil landscapes over the territory of the fUSSR. During the day, the presence of clouds is associated with low values of sensible heat flux from the surface to the atmosphere. At night (and during the day in winter in high latitudes), the sign of the effect is different, but because the direction of sensible heat flux is also different (from the atmosphere to the surface), the presence of clouds again reduces the turbulent heat exchange between the bare soil and the atmosphere.

The OCE estimates on summer sensible heat flux are compared with similar estimates from five AMIP GCM runs to assess the abilities of these GCMs to reproduce the response of this flux to cloud cover change. We found that the sign, the pattern, and the magnitudes of our OCE estimates coincide with similar estimates derived from the NCEP and UIUC AMIP runs, thus indicating that this feature of the behavior of the modern climatic system is well reproduced by the land-surface interaction schemes of these two models. Three other tested GCMs (CCC, MPI, and GISS) produce patterns and signs of the OCE that are similar to our estimates, but the magnitude of this effect in the forest zone is

TABLE 4. Generalized characteristics of accuracy (representative standard deviations, W m^{-2}) of the estimates of OCE on the July mean daily sensible heat fluxes over the fUSSR territory, derived from empirical data and from output of the four GCM AMIP runs (Fig. 7). Standard deviations of mean differences $f_{avg} - f_{clear\ sky}$ were estimated from appropriate time series of these fluxes at each point or grid cell.

	Data	NCEP	CCC	GISS	UIUC	MPI
Tundra	5	4	2	4	5	4
Forest	5	8	2	3	5	4
Steppe	10	3	3	1	2	3
Desert	3	~0.5	~0.5	~0.1	~0.2	~0.3

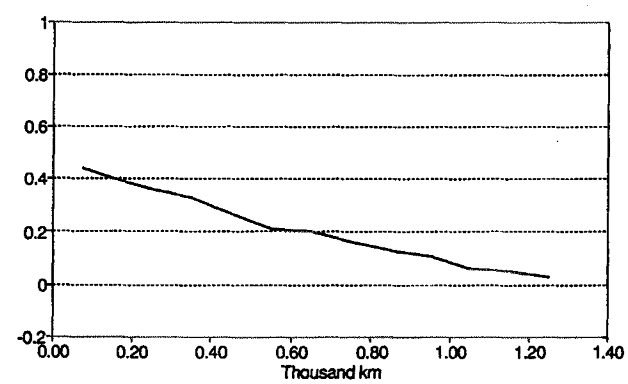

FIG. 8. Spatial correlation function of the mean monthly sensible heat fluxes over the fUSSR for July. Estimates are obtained from the 223 time series of monthly sensible heat fluxes at the fUSSR primary network (dots in Fig. 1) for the period 1966 to 1984 or 1986 under the assumption that the field of these fluxes is spatially isotropic and homogeneous.

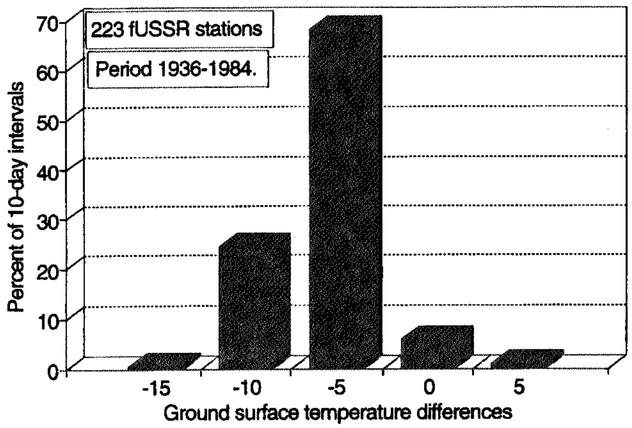

FIG. 9. Average distribution of the mean difference between ground surface temperatures associated with the presence of snow on the ground. It is a result of processing approximately 3×10^6 synoptic observations at the primary network (dots in Fig. 1) well distributed over the former USSR inside the same 10-day intervals when both "snow on the ground" and "no snow on the ground" situations were reported during the period 1936–84/6. Initially data were separated by climatic zones from tundra to desert and by day and night observations, but these separations produce little difference in results.

reduced twofold compared to our estimates for bare soil landscapes over the fUSSR territory.

Snow on the ground is associated with temperature depression. When the effect of this depression is excluded, the presence of snow on the ground is generally associated with less water vapor in the lower troposphere under clear-sky conditions, while the evaporation rate and sensible heat flux are higher than average.

The differences in turbulent heat fluxes with snow on the ground are often being suggested as the *cause* of special effects that overrun the radiative effects of snow (Cohen and Rind 1991). So far, we cannot support or disregard this point of view. Our diagnostic study indicates only that after a temperature decrease is taken

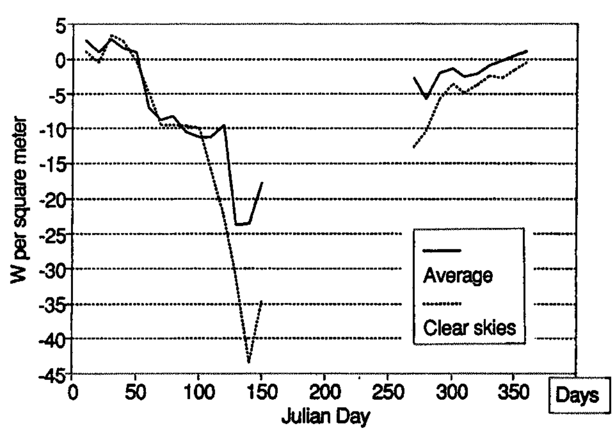

FIG. 10. Seasonal cycle of the differences between sensible heat fluxes with snow on the ground and bare soil conditions. Ten-day averages. Data of 223 primary USSR stations are used for the period 1936–84/6. The categorizing procedure is similar to that described in Fig. 9.

TABLE 5. Mean increase (W m^{-2}) of turbulent heat fluxes for the same ground surface temperatures with snow on the ground. Data pertain to the fUSSR territory (3- and 6-h observations from 223 primary stations for the period 1936–84/6).

	Average conditions	Clear-sky conditions
Sensible heat flux	11	29
Latent heat flux from saturated surface	6	18

into account, the sign of the effect of snow cover on turbulent heat fluxes changes {the appropriate OE, [Eq. (2)] and POE [Eq. (3)] statistics have different signs}. A comprehensive assessment of all components of the surface heat balance in cold environments similar to that made by Cohen and Rind (1991) with their GCM output but based on observations will be a relevant tool for future analysis of these effects.

Acknowledgments. This study has been supported by National Science Foundation Grant ATM-9501320 and NOAA Office of Global Programs and the Department of Energy Environmental Science Division Grant NA66GPO376. The authors express their gratitude to

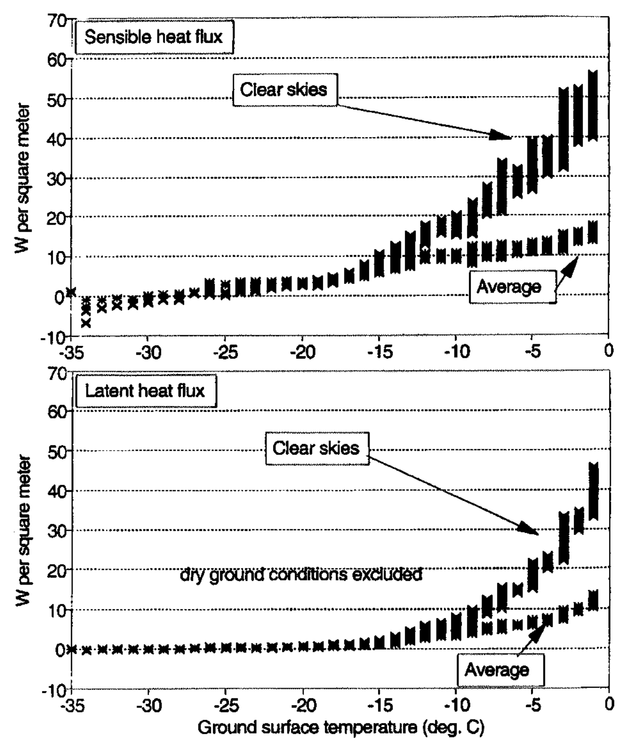

FIG. 11. Differences of sensible heat fluxes with snow on the ground and with bare soil condition ($f_{snow} - f_{bare}$) as a function of the atmosphere–land surface boundary temperature. The estimates are based on the entire dataset of primary stations for the period 1936–84/6. Each difference was calculated for a given temperature range at the station if there were more than 10 measurements of sensible heat fluxes for bare and complete snow cover conditions. In this figure, the differences are smoothed by a 200-point moving average.

TABLE 6. Mean decrease (percent) of the near surface water vapor content for the same surface air temperatures with snow on the ground as derived from 3-h data from 35 stations well distributed over the fUSSR territory. The data were compared in the same temperature ranges (±0.5°C) of surface air and ground temperatures (Groisman et al. 1995).

Sorting by surface	Average conditions	Clear-sky conditions
Air temperature	−4%	−13%
Ground temperature	−8%	−22%

the Meteorological Service of Russia for access to the hourly meteorological data, which made this study possible, and to modeler groups of the Canadian Climate Centre, Goddard Institute for Space Studies, National Centers for Environmental Prediction, University of Illinois at Champaign–Urbana, and Max-Plank Institute, who provided the output of their AMIP runs. We express our special thanks to Dr. Dale Kaiser from the Carbon Dioxide Information Analysis Center (Oak Ridge, Tennessee), who quality controlled the major part of the fUSSR meteorological data (Razuvaev et al. 1995) and organized it in an extremely convenient form for free access by the international research community, and to Mr. Frank Keimig from the Climatology Lab of the University of Massachusetts (Amherst, Massachusetts) for the invaluable help at different stages of this study.

APPENDIX

Effect of Irrigation on Turbulent Heat Fluxes: Desert versus Oases

This assessment is a pilot study that emerged during the process of verification of the GG algorithm of sensible heat flux evaluation in the desert environment. Groisman and Genikhovich (1997) initially planned to complement the verification of this algorithm by the comparison of the estimates of these fluxes from standard meteorological data at the first-order station Chardzhow (39.1°N, 63.6°E) and the heat balance station, (HBS) with the same name located nearby. However, analysis of the metadata from the HBS showed that this comparison would be valid only during the cold half of the year, when human activity does not interfere with the observational program at the HBS.

Chardzhow, Turkmenistan, is located in the Kara-Kum Desert on the bank of the Amu-Dar'ya River. Although the climate conditions are harsh [no precipitation from June to October; mean values of summer daytime surface air temperature around 30°C, while the surface ground temperature is in mid-50s (°C); severe dust storms], the Amu-Dar'ya River water provides means for agriculture (cotton production and gardening). The population of this city steadily grows, and the HBS, which was installed in the late 1950s on an open "natural" desert site, became surrounded by vegetable gardens 30 years later. These gardens are intensively irrigated from mid-April to the end of September. As a result in 1983, the HBS metadata reported sparse dry grass on the site from January to March and from November to December, but from April to September, green grass was reported (with the end of irrigation season in October, this grass became yellow and dried out in November). Due to the large distance from gardens and the specifics of the procedure of the ground surface temperature observations, meteorological station Chardzhow (in the airport) was not (or less) affected by irrigation. The values of daytime ground surface temperatures in the mid-50s (°C) and the near-surface temperature gradients about 15 K in the lowest 2 m indicate the absence of any evaporation at the Chardzhow meteorological station in the summer daytime.

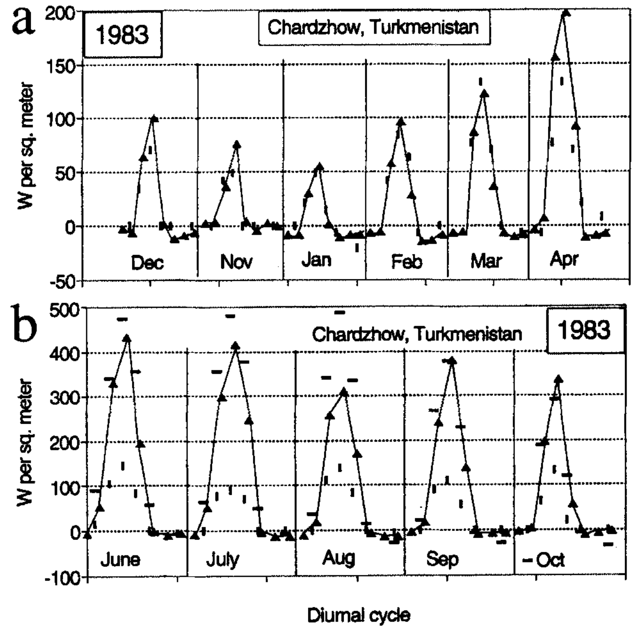

FIG. A1. Turbulent heat fluxes in the desert environment at Chardzhow, Turkmenistan. Mean monthly values for 1983 with preserved diurnal cycle. (a) Sensible heat fluxes for the "cold" season of the year. Estimates suggested by GG (triangles connected by solid lines) were obtained from the standard meteorological data at the primary station, the Chardzhow airport, with a 3-h time step starting with 0000 UTC; thus, the maximum in these estimates of sensible heat fluxes at 0900 UTC corresponds to 1400 LT. Vertical bars represent the estimates derived from the observations at the Chardzhow heat balance station by a standard method based on the station's gradient observations. These observations are made six times per day at 0100, 0700, 1000, 1300, 1600, and 1900 LT. (b) The same estimates for June through October, but additionally total turbulent heat fluxes (latent + sensible heat fluxes, horizontal bars) derived from the observations at the Chardzhow HBS are presented. The rationale for this addition is a high level of irrigation of vegetable gardens around the HBS that makes the comparison of sensible heat fluxes from the desert environment (airport) with the HBS data meaningless without accounting for latent heat fluxes. In May (not shown) in both locations, evaporation significantly contributes to the total surface turbulent heat flux.

Figure A1a shows the close correlation of the mean monthly sensible heat fluxes to preserved diurnal cycle (i.e., averaged over approximately 30 observations at the same time of day) derived from our method and at the HBS for the cold season. The cold season here corresponds to the mean monthly temperatures from slightly above the freezing point in January to 9°C in March. This validation of the GG method of estimation of sensible heat flux was extremely successful, but in the warm season, a new situation emerges.

The mean April temperature is 17°C, and in the second half of this month, the irrigation season starts. This immediately leads to a distortion of the correlation of the daytime sensible heat fluxes at the HBS and at the airport. In the afternoon, we estimate this flux at the airport on the order of 200 W m^{-2}, while the HBS gives only 130 W m^{-2}. The HBS mean monthly afternoon sensible heat flux was never higher than 170 W m^{-2} during this year, while the meteorological station mean monthly afternoon heat flux was at the level of 400 W m^{-2} during most of summer 1983. At the same time, the HBS reported significant summer evaporation, a process that ceased to exist in the bare soil conditions at the Chardzhow meteorological station. Figure 12b shows the turbulent heat fluxes at these two locations in the warm period of 1983. During this period, when precipitation is absent, the total daytime turbulent heat flux (sensible + latent) from the HBS site was close to the sensible heat flux at the airport (in June, September, and October) or exceeded it by 20% (July) to 50% (August). Thus, during the two hottest months of the year during the peak of irrigation, not only does a redistribution of the income radiation between two types of turbulent fluxes occur, but a total redistribution of all components of the surface heat balance is observed: there is an increase of absorbed radiation due to the reduction of the albedo of the soil, a decrease of outgoing longwave radiation due to the decrease of T_g, and a consequent release of the extra energy to the atmosphere via turbulent heat fluxes. Because of the short distance between the Chardzhow airport and the HBS, and a flat terrain, it is difficult to prescribe these changes to local climate specifics (e.g., clouds, etc.). The human impact, irrigation, drives the revealed differences in turbulent heat fluxes. It should be noted that the changes in radiation can be claimed only at the peak of the desert summer with enormous (and regretfully, reckless) irrigation. The same plots for June, September, and October hint that only redistribution between different types of turbulent heat fluxes occurs with the processes under consideration.

While writing this appendix, we did not pretend to assess the problem of land-use effects on the turbulent heat fluxes comprehensively. To do this, we must process a larger dataset than a 1-yr record at one station, involve in the analyses radiation measurements that are available at the same locations, touch base with the modern observational programs from space, etc. We also do not pretend to be pioneers in this subject. All contemporary textbooks on climatology more or less address the oases effects. But here we draft a new approach, which can deliver a quantitative and comprehensive assessment of the interaction of land use and surface heat balance using the rich and still not utilized sources of standard and nonstandard meteorological information available worldwide.

REFERENCES

Anan'ev, I. P., 1986: Remote sensing estimates of evaporation from forest and swamp landscapes of western Siberia (in Russian). *Mater. Meteor. Nablyudenii,* **10,** 109–115.

Anderson, E. A., 1973: National Weather Service river forecast system—Snow accumulation and ablation model. NOAA Tech. Memo. NWS HYDRO-17, U.S. Dept. of Commerce, Silver Spring, MD. [Available from National Technical Information Service, 5285 Port Royal Rd., Springfield, VA 22161.]

Budyko, M. I., 1948: *Evaporation under Natural Conditions* (in Russian). Gidrometeoizdat, 136 pp.

Cess, R. D., and Coauthors, 1991: Interpretation of snow–climate feedback as produced by 17 general circulation models. *Science,* **253,** 888–892.

Chanine, M. T., 1995: Observation of local cloud and moisture feedbacks over high ocean and desert surface temperatures. *J. Geophys. Res.,* **100** (D5), 8919–8927.

Cohen, J., and D. Rind, 1991: The effect of snow cover on the climate. *J. Climate,* **4,** 689–706.

Dekic, L. J., D. T. Mihailovic, and B. Rajkovic, 1995: A study of the sensitivity of bare soil evaporation schemes to soil surface wetness, using the coupled soil moisture and surface temperature prediction model, BARESOIL. *Meteor. Atmos. Phys.,* **55,** 101–112.

Del Genio, A. D., M.-S. Yao, W. Kovari, and K. K.-W. Lo, 1996: A prognostic cloud water parameterization for global climate models. *J. Climate,* **9,** 270–304.

Famiglietti, J. S., and E. F. Wood, 1991: Evapotranspiration and runoff from large land areas: Land surface hydrology for atmospheric general circulation models. *Surv. Geophys.,* **12,** 179–204.

——, ——, M. Sivapalan, and D. J. Thongs, 1992: A catchment scale water balance model for FIFE. *J. Geophys. Res.,* **97** (D17), 18 997–19 007.

Gaffen, D. J., and W. P. Elliott, 1993: Column water vapor content in clear and cloudy skies. *J. Climate,* **6,** 2278–2287.

Gates, W. L., 1992: AMIP: The Atmospheric Model Intercomparison Project. *Bull. Amer. Meteor. Soc.,* **73,** 1962–1970.

Genikhovich, E. L., and G. I. Osipova, 1984: Determination of the exchange coefficient using data of routine meteorological observations (in Russian). *Trans. Main Geophys. Observatory,* **479,** 62–69.

Gol'tzberg, I. A., 1969: *Microclimate of the USSR.* Israel Program for Scientific Translations, 236 pp.

Gorodetsky, O. A., Ed., 1970: *Observations at the USSR Meteorological Network. Definitions of Meteorological Elements and Estimates of the Accuracy of their Observations* (in Russian). Gidrometeoizdat, 89 pp.

Groisman, P. Ya., and P. M. Zhai, 1994: Climate under clear skies: An empirical benchmark for the cloud feedback problem. AGU 1994 Fall Meeting. *Eos, Trans. Amer. Geophys. Union,* (Suppl.),**74,** 219.

——, and ——, 1995: Climate variability under clear skies: Applications for the cloud and snow cover feedback problems. *Proc. Sixth Int. Meeting on Statistical Climatology,* Galway, Ireland, All Ireland Committee on Statistics and Amer. Meteor. Soc., 605–608.

—— and E. L. Genikhovich, 1997: Assessing surface–atmosphere interactions using former Soviet Union standard meteorological network data. Part I: Method. *J. Climate,* **10,** 2154–2183

——, T. R. Karl, and R. W. Knight, 1994a: Observed impact of snow cover on the heat balance and the rise of continental spring temperatures. *Science,* **263,** 198–200.

——, ——, ——, and G. L. Stenchikov, 1994b: Changes of snow cover, temperature, and the radiative heat balance over the Northern Hemisphere. *J. Climate,* **7,** 1633–1656.

——, P.-M. Zhai, and E. L. Genikhovich, 1995: Cloud and snow cover effects on the surface–atmosphere interactions. *Extended Abstracts, Joint Meeting of the Canadian Geophysical Union-Hydrology Section (CGU-HS) and International GEWEX Workshop on Cold-Season/Region Hydrometeorology,* Banff, AB, Canada, Canadian Geophysical Union, 209–212.

——, E. L. Genikhovich, and P.-M. Zhai, 1996a: "Overall" cloud and snow cover effects on internal climate variables: The use of clear sky climatology. *Bull Amer. Meteor. Soc.,* **77,** 2055–2065.

——, R. S. Bradley, and E. L. Genikhovich, 1996b: Climate sensitivity to cloud and snow cover variations: Comparison of global climate models with empirical data. Preprints, *Second Int. Scientific Conf. on the Global Energy and Water Cycle,* Washington, DC, GEWEX Project of the World Climate Research Programme, 523–524.

Hahn, C. J., S. G. Warren, and J. London, 1994: Edited synoptic cloud reports from ship and land stations over the globe, 1982–1991. Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, TN, 47 pp. and appendixes. [Available from National Technical Information Service, 5285 Port Royal Rd., Springfield, VA 22161.]

Hall, F. G, K. E. Huemmrich, S. J. Goetz, P. J. Sellers, and J. E. Nickeson, 1992: Satellite remote sensing of surface energy balance: Success, failures, and unresolved issues in FIFE. *J. Geophys. Res.,* **97** (D17), 19 061–19 089.

Harrison, E. F., P. Minnis, B. R. Barkstrom, V. Ramanathan, R. D. Cess, and G. G. Gibson, 1990: Seasonal variation of cloud radiative forcing derived from the Earth Radiation Budget Experiment. *J. Geophys. Res.,* **95,** 18 687–18 703.

Henderson-Sellers, A. A., Z.-L. Yang, and R. E. Dickinson, 1993: The Project for Intercomparison of Land-Surface Parameterization Schemes (PILPS). *Bull. Amer. Meteor. Soc.,* **74,** 1335–1349.

IGBP, 1991: The International Geosphere–Biosphere Programme, Plant–water interactions in large-scale hydrological modeling. Report of a Workshop 17, 44 pp.

IPCC, 1990: *Climate Change. The IPCC Scientific Assessment.* Cambridge University Press, 362 pp.

——, 1996: *Climate Change 1995: The Science of Climate Change. The Second IPCC Scientific Assessment.* Cambridge University Press, 572 pp.

Kagan, R. L., 1979: *Averaging of Meteorological Fields* (in Russian). Gidrometeoizdat, 212 pp.

Kuchment, L. S., V. N. Demidov, and Yu. G. Motovilov, 1983: *Streamflow Formation* (in Russian). Nauka, 216 pp.

Lamb, H. H., 1955: Two-way relationship between the snow or ice limit and 1,000–500 mb thickness in the overlying atmosphere. *Quart. J. Roy. Meteor. Soc.,* **81,** 172–189.

Leathers, D. J., A. W. Ellis, and D. A. Robinson, 1995: Characteristics of temperature depressions associated with snow cover across the northeast United States. *J. Appl. Meteor.,* **34,** 381–390.

Mihailovic, D. T., B. Rajkovic, B. Lalic, and L. Decic, 1995: Schemes for parameterizing evaporation from non-plant-covered surface and their impact on partitioning the surface energy in land-air exchange parameterization. *J. Appl. Meteor.,* **34,** 2462–2475.

Namias, J., 1985: Some empirical evidence for the influence of snow cover on temperature and precipitation. *Mon. Wea. Rev.,* **113,** 1542–1553.

Peck, E. L., and J. C. Schaake, 1990: Network design for water supply forecasting in the west. *Water Resour. Bull.,* **26,** 87–99.

Phillips, T. J. 1994: A summary documentation of the AMIP models. PCMDI Rep. 18, 343 pp. [Available from Program for Climate Model Diagnosis and Intercomparison, Lawrence Livermore National Laboratory, Livermore, CA 94550; and National Technical Information Service, 5285 Port Royal Rd., Springfield, VA 22161.]

Pokrovskaya, T. V., and L. P. Spirina, 1965: Estimates of the snow cover impact on spring air temperature over the European part of the USSR (in Russian). *Trans. Main Geophys. Observatory,* **181,** 110–113.

Potter, G. L., J. M. Slingo, J.-J. Morcrette, and L. Corsetti, 1992: Modeling perspective on cloud radiative forcing. *J. Geophys. Res.,* **97** (D18), 20 507–20 518.

Razuvaev, V. N., E. B. Apasova, R. A. Martuganov, D. P. Kaiser, 1995: Six- and three-hourly meteorological observations from 223 USSR stations. ORNL/CDIAC-66, NDP-048, 68 pp. plus appendixes. [Available from Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, TN 37831-6335.]

Robock, A., K. Ya. Vinnikov, C. A. Schlosser, N. A. Speranskaya, and Y. Xue, 1995: Use of midlatitudinal soil moisture and meteorological observations to validate soil moisture simulations with biosphere and bucket models. *J. Climate,* **8,** 15–35.

Rosenberg, N. J., B. L. Blad, and S. B. Verma, 1983: *Microclimate. The Biological Environment.* John Wiley & Sons, 495 pp.

Sellers, P. J., and F. G. Hall, 1992: FIFE in 1992: Results, scientific gains, and future research directions. *J. Geophys. Res.,* **97** (D17), 19 091–19 109.

Sohn, B.-J., and F. R. Robertson, 1993: Intercomparison of observed cloud radiative forcing: A zonal and global perspective. *Bull. Amer. Meteor. Soc.,* **74,** 997–1006.

State Committee of the USSR on Hydrometeorology and Environment Control, 1985: *Handbook of Hydro-Meteorological Observations at Primary and Secondary Hydro-Meteorological Networks* (in Russian). Vol. 3, Part 1, Gidrometeoizdat, 300 pp.

Stouffer, R. J., S. Manabe, and K. Ya. Vinnikov, 1994: Model assessment of the role of natural variability in recent global warming. *Nature,* **367,** 634–636.

Voeikov, A. I., 1889: Snow cover, its effects on soil, climate, and weather and methods of investigation (in Russian). *Notes Russian Geograph. Soc. on General Geography,* **18** (2).

Wagner, A. J., 1973: The influence of average snow depth on monthly mean temperature anomaly. *Mon. Wea. Rev.,* **101,** 624–626.

Warren, S. G., C. J. Hahn, J. London, R. M. Chervin, and R. L. Jenne, 1986: Global distribution of total cloud cover and cloud type amounts over land. NCAR Tech. Note TN-273+STR, 229 pp. [Available from National Technical Information Service, 5285 Port Royal Rd., Springfield, VA 22161.]

Weare, B. C., and I. I. Mokhov, 1995: Evaluation of total cloudiness and its variability in the Atmospheric Model Intercomparison Project. *J. Climate,* **8,** 2224–2238.

Zubenok, L. I., 1947: On the question of measuring the soil surface temperature. *Trans. Main Geophys. Observatory,* **6,** 71–74.