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Modeling the Response of Microwave Self-Radiation of the Ocean–Atmosphere System to Horizontal Heat Transfer in the Atmospheric Boundary Layer

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Abstract—Carried out is the study of the response of microwave radiation of the ocean–atmosphere system to the horizontal heat transfer in the atmospheric boundary layer (ABL). Model estimates are obtained for the radiation on the wavelengths of 0.6, 0.8, 1, 1.35, and 1.6 cm. It is demonstrated that the value and sign of brightness temperature contrasts induced by the horizontal transfer depend on the ABL density stratification and transfer direction relative to the orientation of horizontal gradients of air temperature and air humidity. Variations of brightness temperature in the ABL at the wavelength of 1.35 cm reach 30–40 K. Observed is the high correlation between the variations of brightness temperature in the ABL at the wavelength of 1.35 cm and the vertical fluxes of sensible and latent heat for different types of the ABL stratification and for different conditions of advective transport.

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1. INTRODUCTION

Satellite microwave radiometric methods of analyzing the characteristics of the heat interaction between the ocean and atmosphere have been more and more widely used as traditional methods of studying the ocean and have become more and more attractive for oceanologists, meteorologists, and climatologists. The study of mechanisms of the formation of the relationship between the satellite-measured characteristics of the microwave radiation of the ocean–atmosphere system and such characteristics of the contact layer (interface) of the system as vertical turbulent fluxes at the ocean–atmosphere interface is of great importance for working out these methods. The urgency of this problem is associated with the fact that the field of the microwave self-radiation contains the information not only about the lower atmospheric layers (directly participating in the energy exchange with the ocean surface) but also about upper layers.

Papers [4, 9] theoretically studied the relationship between the intensity of the microwave self-radiation (brightness temperature) of the ocean–atmosphere system and the surface heat fluxes taking account of the vertical redistribution of heat and moisture in the atmospheric boundary layer. However, later on, the joint analysis of the data of oceanographic and meteorological observations made it clear that the effect of the vertical transport of heat and moisture in the atmosphere cannot explain some features of brightness temperature variations in the ocean–atmosphere system. For example, during the NEWFOUEX-88 and ATLANTEX-90 shipborne experiments in the Newfoundland energy-active zone of the North Atlantic noted for its intensive cyclonic activity [5], the brightness temperature contrasts observed from the satellite exceeded by an order of magnitude the estimates of brightness temperature contrasts obtained with account of only the effects of the vertical transfer of heat and moisture in the atmosphere.

Such discrepancies can be explained by the influence of horizontal advective motions in the atmosphere on the temperature and humidity characteristics of the ocean–atmosphere system in the zones of cyclonic activity; such advective motions are not taken into account in these papers. The evidences of this fact can be found in the above-mentioned observational data in ATLANTEX-90 experiment. In particular, these data include the cases, when the decrease in the mean values of air temperature and enthalpy of the atmospheric boundary layer (ABL) by 5–8 C and 2–3 MJ/m^2, respectively, and the increase in surface fluxes of sensible and latent heat by 500–800 W/m² were observed on April 9–11, 1990, when the powerful cyclone approached the areas of the location of *Viktor Bugaev* (43° N, 43° W) and *Musson* (42° N, 46° W) research vessels. Such dramatic changes would be impossible without taking account of horizontal advective processes accompanying the movement of cyclones.

The present paper tries to assess the dependence of the microwave self-radiation of the ocean–atmosphere system on the variations of the structural boundary layer caused by horizontal heat transfer.

2. ESTIMATION OF DEPENDENCE OF THE ABL STRUCTURE ON HORIZONTAL HEAT TRANSFER

2.1. Initial Model

The structure of meteorological fields in the ABL that define the energy exchange between the ocean and atmosphere and the fluxes of electromagnetic radiation registered on the upper boundary of the atmosphere, is formed first of all under the influence of the processes developed along the vertical and is described within the framework of one-dimensional models [6, 10]. The processes developed along the vertical are the key ones for the formation of major features of the ABL structure whereas the role of the horizontal inhomogeneity cannot always be considered negligible. The influence of horizontal advection becomes considerable, in particular, near the frontal zones associated with intensive cyclones or in the vicinity of the land–sea boundary, where the increase in horizontal gradients of meteorological fields is observed.

To describe the structure of the ABL and associated turbulence characteristics, let us use a model constructed within the framework of traditional approximations with respect to the methods of parameterization of small-scale turbulence but taking account of the horizontal advection of heat and moisture. Basic equations of the model and boundary conditions are written in the following way:

$$f(v - v_{\mathrm{g}}) + \frac{1}{\rho}\frac{\partial \tau_x}{\partial z} = 0, \quad -f(u - u_{\mathrm{g}}) + \frac{1}{\rho}\frac{\partial \tau_y}{\partial z} = 0; \tag{1}$$

$$-\frac{1}{\rho c_p}\frac{\partial F_T}{\partial z} - u\nabla_x T - v\nabla_y T = 0; \tag{2}$$

$$-\frac{1}{\rho}\frac{\partial F_q}{\partial z} - u\nabla_x q - v\nabla_y q = 0; \tag{3}$$

$$u = u_h = u_{\mathrm{g}};\ v = v_h = v_{\mathrm{g}},\ T = T_h,\ q = q_h \ \text{ if } z = h = 1400 \text{ m}; \tag{4}$$

$$u = 0,\ v = 0 \ \text{ if } z = z_0; \tag{5}$$

$$T = T_0,\ q = q_0 \ \text{ if } z = z_{0H}. \tag{6}$$

Here, $\tau_x = -\rho\overline{u'w'}$ and $\tau_y = -\rho\overline{v'w'}$ are horizontal components of the vertical turbulent momentum flux (tangential stress); $F_T = \rho c_p\overline{T'w'}$ and $F_q = \rho\overline{q'w'}$ are the vertical turbulent fluxes of heat and moisture, respectively; $\mathbf{u} = (u, v)$ is the horizontal velocity vector of the mean motion with the components u and v along the x- and y-axes of the Cartesian coordinate system; w is the vertical velocity along the z-axis; T is the air temperature; q is specific humidity; ρ is the air density; c_p is its specific heat at constant pressure; $\nabla_x T$ and $\nabla_y T$ are the components of the vector of horizontal air temperature gradient within the ABL along the x- and y-axes, respectively; $\nabla_x q$ and $\nabla_y q$ are the analogous components of the vector for its specific humidity; $u_{\mathrm{g}} = -(1/(\rho f))\,\partial p/\partial y$ and $v_{\mathrm{g}} = (1/(\rho f))\,\partial p/\partial x$ are the horizontal components of the vector of geostrophic wind speed considered height-independent (independence of pressure gradient from z is the usual assumption of the boundary layer theory [7]); f is the Coriolis parameter; p is the air pressure; h is the height of the ABL; z_0 and z_{0H} are the roughness parameters for the momentum and heat (moisture) fluxes. Strokes are turbulent pulsations and the overline is a sign of statistical averaging.

The model is based on the numerical solution of equations (1)–(3) with the first-order closure scheme for vertical turbulent fluxes

$$\tau_x/\rho = -\overline{u'w'} = k_{\mathrm{m}}\,\partial u/\partial z, \quad \tau_y/\rho = -\overline{u'w'} = k_{\mathrm{m}}\,\partial v/\partial z; \tag{7}$$

$$F_T/(\rho c_p) = \overline{T'w'} = -k_H(\partial T/\partial z + g/c_p); \tag{8}$$

$$F_q/\rho = -\overline{q'w'} = k_H \partial q/\partial z, \tag{9}$$

including the nonlinear dependence of coefficients of the turbulent mixing

$$k_{\mathrm{m}} = l_m^2 f_m(\mathrm{Ri})\left|\partial \mathbf{u}/\partial z\right|, \quad k_H = l_m l_H f_H(\mathrm{Ri})\left|\partial \mathbf{u}/\partial z\right| \tag{10}$$

on the local wind speed shear $\left|\partial \mathbf{u}/\partial z\right|$ and density stratification characterized by the Richardson number

$$\mathrm{Ri} = \frac{g[(1/T)(\partial T/\partial z + g/c_p) + 0.61\,\partial q/\partial z]}{\left|\partial \mathbf{u}/\partial z\right|^2}. \tag{11}$$

In (8), (10), and (11), g/c_p is the adiabatic temperature gradient; g is the acceleration of gravity; l_m and l_H are the displacement lengths at the neutral classification specified with the Blackadar formulae [8]; f_m and f_H are the stability functions depending on Ri and specified with empirical formulae.

Turbulent fluxes on the sea–air interface are defined using the Monin–Obukhov similarity theory for the surface atmospheric layer [7] which upper boundary coincides with the lower computation level of the model (1)–(3).

The influence of the horizontal inhomogeneity at large (i.e., exceeding the size of small-scale turbulent vortices) scales is taken into account in the models by means of adding corresponding terms into (2) and (3), where the horizontal gradients of the fields of the air temperature $\nabla T = (\nabla_x T, \nabla_y T)$ and humidity $\nabla q = (\nabla_x q, \nabla_y q)$ not depending on height are considered known and are specified depending on the simulated situation. The height-dependent wind speed $\mathbf{u} = (u, v)$ is the desired function determined during the solution of the problem.

Let us use this model when assessing the impact of horizontal transfer on the structure of meteorological fields in the atmospheric boundary layer and on the intensity of its energy exchange with the ocean.

2.2. A Scheme of the Numerical Experiment

To obtain the quantitative estimates of the impact of horizontal transfer, three series of numerical experiments were carried out. Three types of density stratification in the ABL were considered as base background conditions: quasineutral, unstable, and stable stratification. Background conditions are the cases of horizontal homogeneity, where the horizontal advection is absent. The variations of the background stratification were specified by means of varying the air temperature on the upper boundary of the ABL within the range of T_h = 280–310 K at the fixed temperature on the water–air interface T_0 = 300 K. The height of the upper boundary of the ABL for computations is taken to be equal to 1400 m.

As to the seasonal frequency of the separated types of the ABL stratification as applied to the Newfoundland energy-active zone of the North Atlantic, according to the data presented in [5], the neutral or stable stratification (when the drops of temperature of water and the air are small or slightly negative) is typical of summer conditions, and the unstable stratification (water is warmer than the air) is mostly typical of winter. According to [5], here the climatic frequency of meridional atmospheric circulation makes up 40% and that of the zonal circulation, 60%.

After that, the horizontal advective heat transfer with the varying orientation of the transfer vector relative to the background wind field on the upper boundary of the ABL was introduced for each type of background conditions. The vertical distribution of the wind speed providing the horizontal advection when horizontal gradients of air temperature and air humidity are present, was computed in the process of the problem solution. Horizontal gradients of temperature $\nabla T = (\nabla_x T, \nabla_y T)$ and specific humidity $\nabla q = (\nabla_x q, \nabla_y q)$ were supposed to be specified. The vector of the geostrophic wind speed on the upper boundary of the ABL was oriented along one of the coordinate axes: u_{g} = 10 m/s and v_{g} = 0. The choice of the direction of these axes can be arbitrary due to the invariance of the problem of the coordinate system rotation relative to the vertical axis. Only the mutual orientation is significant of wind speed vectors and gradients of transported substances defining the rate of advective changes $\mathbf{u} \cdot \nabla T$ and $\mathbf{u} \cdot \nabla q$ expressed through the scalar product of wind speed vectors $\mathbf{u}$ and gradients ∇T and ∇q.

To single out more distinctly the effects of horizontal advection, rather large values of horizontal temperature gradients of about 5×10^{-5} K/m = 5 K/100 km were selected. Such increase in horizontal gradients as compared with usual values being smaller by an order of magnitude, is observed near the frontal zones associated with intensive cyclones or in the vicinity of the boundaries between different types of the underlying surface such as the land–sea interface.

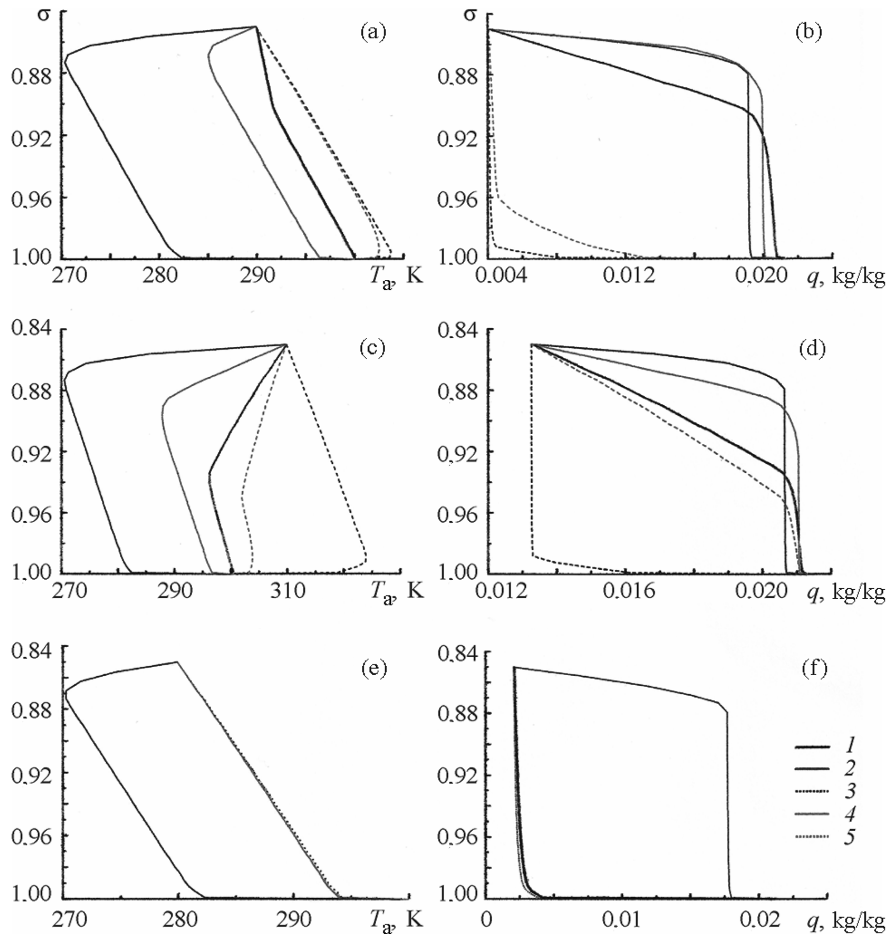

Fig. 1. Vertical profiles of (a, c, e) air temperature and (b, d, f) specific humidity for the cases of (a, b) neutral, (c, d) stable, and (e, f) unstable background stratification under different conditions of horizontal heat transfer within the ABL. (*1–5*) Numbers of cases given in the table.

The modeling of different types of the background stratification was carried out by means of varying the temperature T_h on the upper boundary of the ABL at the fixed values of the sea surface temperature $T_0 =$ $= 300$ K and relative humidity of 30% on the upper boundary of the ABL: $T_h = 290$ K for neutral stratification, $T_h = 280$ K for unstable stratification, and $T_h = 310$ K for stable stratification.

For all cases presented below, the vector of the geostrophic wind speed on the upper boundary of the ABL is $\mathbf{u}_g = (u_g, v_g) = (10, 0)$ m/s, the sea surface temperature is $T_0 = 300$ K, the saturating moisture content at the surface temperature is $q_0 = 2.12 \times 10^{-2}$ kg/kg, $\partial_x q = \partial_y q = 0$.

2.3. Results of the ABL Structure Computation for Different Background Conditions

The results of computation of vertical distributions of air temperature and specific humidity forming the filed of the ABL microwave self-radiation for different background conditions of density stratification and for different contributions of horizontal advection are presented in Fig. 1. The fluxes of sensible $F_{T0} = F_T\big|_{z = z_{0H}}$ and latent $F_{L0} = L_e F_q\big|_{z = z_{0H}}$ heat on the water–air interface ($L_e = 2.5 \times 10^6$ J/kg is the specific heat of evaporation, F_q is the moisture flow) associated with these distributions are given in the table.

The comparative analysis of these characteristics enables solving the major problem formulated in the present paper, namely, obtaining the estimates of the sensitivity of microwave radiation characteristics of the ABL to the variations of its thermal regime caused by horizontal heat transport.

The first case in the table corresponds to background conditions with the advection absence; in the second case, the vector of the geostrophic wind speed coincides with the positive temperature gradient (cold

Characteristics of heat exchange (W/m^2) on the water–air interface for the cases of neutral, unstable, and stable background stratification at different orientations of horizontal temperature gradients relative to the vector of the geostrophic wind speed on the upper boundary of the ABL

Number of the case	${}_xT$	${}_yT$	Stratification					
			neutral		unstable		stable	
	K/m		F_{T0}	F_{L0}	F_{T0}	F_{L0}	F_{T0}	F_{L0}
1	0	0	–1.4	15.7	136.8	998.3	–5.5	4.5
2	$5\ 10^{-5}$	0	610.6	173.3	616.4	288.9	600.5	52.7
3	$-5\ 10^{-5}$	0	–39.6	345.2	131.1	1021.8	–66.8	45.1
4	0	$5\ 10^{-5}$	61.3	53.9	136.9	997.0	54.8	10.1
5	0	$-5\ 10^{-5}$	–22.0	182.4	136.6	999.6	–26.2	3.8

Note: Explanations are given in the text.

advection); in the third case, the air masses are transported towards the temperature drop (warm advection). The fourth and fifth cases are intermediate: the geostrophic wind vector is oriented in the normal direction to the vector of the gradient T. Advective transport within the ABL is also observed in this case because the rotation of the wind speed vector is registered within the limits of the boundary layer and, according to the model formulation, the horizontal temperature gradient does not vary with height. The vertical structure of the wind field in each case depends in a complicated way on the whole set of external (regarding the ABL) conditions and on the dynamics of the ABL. On the whole, the effects of horizontal advection in this case turn out to be smaller but not negligible as compared with the above situations with the coincidence of the direction of vectors $\mathbf{u}_g = (u_g, v_g)$ and $T = ({}_xT, {}_yT)$.

In Fig. 1, the dimensionless coordinate representing the ratio of the air pressure at the height z to the sea-level pressure is used as a vertical coordinate. The geometric height of about 1400 m corresponds to the upper boundary of the ABL $= 0.85$.

As clear from the data of the table and Fig. 1, the impact of advection in the case of neutral background stratification is quite considerable. As could be expected, the temperature background within the limits of the ABL either decreases (in the case of cold advection) or increases (in the case of warm advection) and the values of temperature on the upper and lower boundaries of the ABL remain invariable according to the conditions of numerical experiments. If these values are constant, the heat fluxes on the ocean surface change cardinally: instead of the close to zero value of the sensible heat flux registered at advection absence, the flux F_{T0} in the case of cold advection reaches 600 W/m^2.

It should be noted that this value is close to the values of heat fluxes observed in the Newfoundland energy-active zone of the North Atlantic in the zones of activity of mid-latitude cyclones during the NEWFOUEX-88 and ATLANTEX-90 experiments [5] as well as to the values of heat fluxes typical of the conditions of the genesis and development of tropical cyclones given in [2].

In the case of warm advection in the ABL, the excess heat is transmitted to the ocean and the negative values of F_{T0} corresponding to this direction of the transfer reach –39.6 W/m^2. The latent heat flux F_{L0} increases regardless of the advective transport direction although not as significantly as F_{T0}. However, the heat exchange due to the latent heat should depend considerably on the humidity advection specified to be equal to zero in the experiments.

In the case of the unstable background stratification, the disturbances caused by the horizontal advection, even in its most pronounced manifestations (i.e., for the cases 2 and 3 in the table), provoke the changes in F_{T0} and F_{L0} by several times although the sign of these fluxes in all these cases remains invariable. As typical of the unstable stratification of the ABL, the ocean transmits heat to the atmosphere in the sensible and latent forms.

In the case of cold advection, the monotonous and comparatively uniform temperature drop within the ABL that takes place at advection absence is replaced by its dramatic decrease in the surface layer changing into the temperature rise as approaching to the upper boundary of the ABL (Fig. 1). It should be noted that here, unlike other cases, the profiles of temperature and humidity are grouped into two significantly differing types of heat condition in the ABL.

The disturbances in the surface fluxes F_{T0} and F_{L0} in the case of stable background stratification caused by advection are generally similar to those in the case of neutral stratification. Noticeable differences take place only for the latent heat flux F_{L0} which turns out to be negative in this case for all directions of horizontal transport. However, the absolute values of F_{L0} remain rather small. At such stratification the vertical structure of the ABL fields is characterized by typical of such cases inversion of temperature near the upper boundary of the ABL which is accompanied by the weakening of the vertical mixing that corresponds to the well-known concept of the structure of the stably stratified ABL [6].

3. RESPONSE OF BRIGHTNESS TEMPERATURE OF THE OCEAN–ATMOSPHERE SYSTEM TO VARIATIONS OF TEMPERATURE AND HUMIDITY CHARACTERISTICS IN THE ABL

3.1. Radiation Model of the Ocean–Atmosphere System

To compute the brightness temperature in the ABL, let us use a standard model that enables taking account of characteristics of the vertical distribution of temperature, pressure, and humidity of the air in the atmosphere. According to this model, the brightness temperature T^{b} of the ocean–atmosphere system in the case of satellite measurements is made up of three components [1]:

$$T^{\mathrm{b}}\quad T_1^{\mathrm{b}}\quad T_2^{\mathrm{b}}\quad T_3^{\mathrm{b}}, \tag{12}$$

where $T_1^{\mathrm{b}}\quad T_{\mathrm{s}}^{\mathrm{b}}\exp(\qquad)$ is the brightness temperature of the sea surface radiation attenuated by the atmosphere, and value of $T_{\mathrm{s}}^{\mathrm{b}}$ is proportional to the emissivity of the water surface and to its thermodynamic temperature T_{s};

$T_2^{\mathrm{b}}\quad {}_{0}^{h}\, T(z)\quad (z)\exp[\quad (z)\qquad (h)]dz$ is the brightness temperature of the upward atmospheric radiation;

$T_3^{\mathrm{b}}\quad \exp[\qquad (h)]R\ {}_{0}^{h}\, T(z)\quad (z)\exp[h\quad (z)]dz$ is the brightness temperature of the downward atmospheric radiation re-reflected by the water surface; $T(z)$ is the thermodynamic temperature of the atmosphere at the level z;

$(z)\qquad {}_{0}^{z}\,(z\ )dz$ is the integral absorption of radiation in the atmosphere depending on the linear absorption and thickness of the layer z counted from the sea surface ($z = 0$); R is the coefficient of the reflection of downward atmospheric radiation from the water surface.

The brightness temperature of the water surface radiation $T_{\mathrm{s}}^{\mathrm{b}}$ and the reflection coefficient R depend on the thermodynamic temperature of the sea surface, on the degree of its roughness, and on the intensity of foam formation associated with the surface wind speed. The brightness temperature of direct and reflected atmospheric components at centimeter and millimeter waves is defined by the absorption of radio waves with water vapor and molecular oxygen of the atmosphere that depends on the air temperature, air humidity, and on the characteristics of their vertical distribution [1]. These are the characteristics forming the base for computing the sensitivity of the brightness temperature of the ocean–atmosphere system to the characteristics of the horizontal heat transfer in the ABL; this is the main objective of the present paper.

3.2. Results of Brightness Temperature Computation for Characteristic Types of the Thermal Regime of the ABL

The computation of the brightness temperature in the ABL is carried out using the radiation model (12) within the wavelength range of 0.6–1.6 cm for the cases of the vertical sounding from the satellite, i.e., for nadir observation. This range partly embraces the area of the resonance absorption (emission) of radio waves in atmospheric molecular oxygen (its right wing) and completely embraces the area of the emission of radio waves in the atmospheric water vapor. These areas of the microwave range contain the information about the air temperature and air humidity and are of great importance for studying the heat interaction between the ocean and atmosphere using the satellite microwave radiometric methods [3].

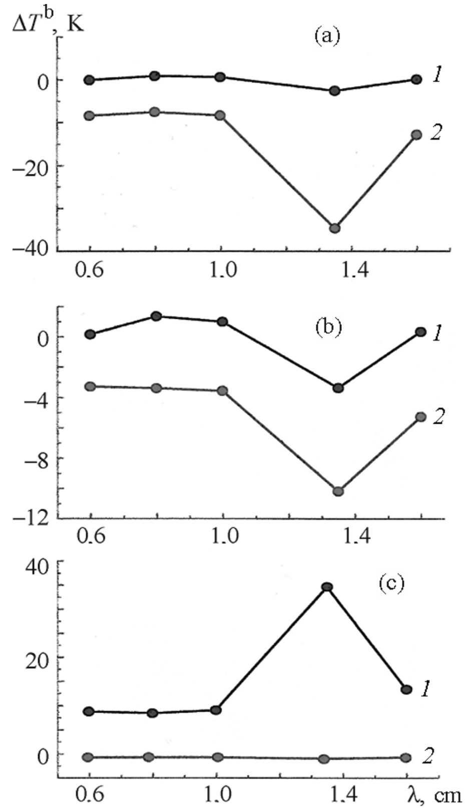

Fig. 2. The rises of brightness temperature T^b at different wavelengths at the passage from background conditions (the absence of horizontal advection in the ABL) to the cases of (*1*) cold advection and (*2*) warm advection: (a) neutral, (b) stable, and (c) unstable stratification of the ABL.

For different wavelengths of this range of radio waves the responses are analyzed of the brightness temperature of the ocean–atmosphere system to the changes in the vertical distribution of the air temperature and air humidity for different types of the ABL stratification and for different orientations of horizontal temperature gradients relative to the vector of the geostrophic wind speed at the upper boundary of the ABL (see the table).

In Fig. 2, the results are presented of computing the brightness temperature rise T^b for different wavelengths when passing from background conditions (absence of horizontal advection in the ABL, case 1 in the table) to the cases of cold (case 2) and warm (case 3) advection for neutral, stable, and unstable density stratification in the ABL. It is clear from Fig. 2 that the sensitivity of brightness temperature in the ABL to the changes in its thermal regime is maximal in the spectral absorption (emission) region of radio waves in the atmospheric water vapor centered relative to the line of 1.35 cm. This result corroborates the importance of this area of the microwave range for studying the processes of heat and moisture exchange between the ocean and atmosphere.

The main result is the fact that the variations of brightness temperature in the ABL on the line of 1.35 cm can reach 30–40 K that exceeds by an order of magnitude the value of its variations caused by the processes of vertical transfer of heat and moisture in the ABL equal to 3–5 K [4, 9].

It should be noted that microwave self-radiation of the ocean–atmosphere system on the line of 1.35 cm is formed in the atmospheric effective layer (~0–2 km). It should be expected that taking account of overlying atmospheric layers (2–10 km) which thermal characteristics are also subjected to the horizontal heat transport and which radiation also contributes to the total radiation of the ocean–atmosphere system, the revealed effect of intensification of T^b contrasts will be manifested more clearly.

Finally, let us consider the relationship between the variations of brightness temperature T^b calculated by the radiation model and heat fluxes on the water–air interface. This relationship revealed from the computation data (12) using the data on fluxes of the table, is shown in Fig. 3a for the flux of sensible heat F_{T0}

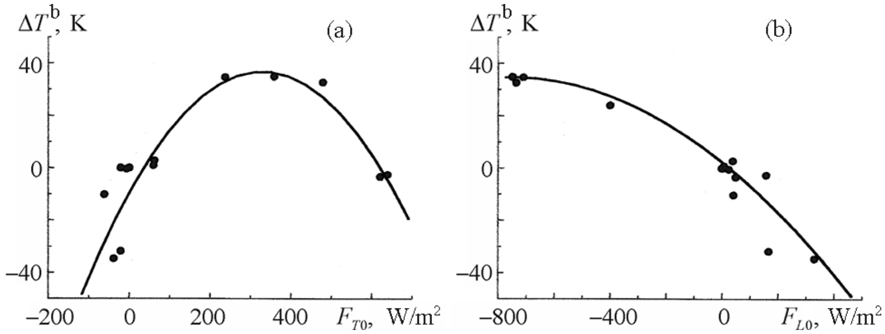

Fig. 3. The relationship between variations of brightness temperature T^{b} in the ABL at the wavelength of 1.35 cm and the vertical fluxes of (a) sensible heat F_{T0} and (b) latent heat F_{L0} on the surface of the water–air interface for different types of the ABL stratification and for different types of advection presented in the table.

and in Fig. 3b for the flux of latent heat F_{L0}. The relationship between T^{b}, on the one hand, and F_{T0} and F_{L0}, on the other hand, is clearly observed in the figure. It is in good agreement with experimental data showing the interdependence of the brightness temperature of the ocean–atmosphere system with surface heat fluxes [3]. These data have been obtained by comparing the results of measuring the brightness temperature of the ocean–atmosphere system from the DMSP F-08 American meteorological satellite using the SSM/I radiometer with the fluxes of heat and moisture based on the data of meteorological and upper-air observations in NEWFOUEX-88 and ATLANTEX-90 experiments [3].

4. CONCLUSIONS

The estimates of the impact of horizontal transfer on the structure of meteorological fields in the atmospheric boundary layer and on its energy exchange with the ocean presented in this paper indicate that characteristics of energy exchange between the ocean and atmosphere strongly depend on horizontal warm advection within the ABL. The value and sign of disturbances caused by horizontal transfer depend on the density stratification in the ABL and on the direction of the transport relative to the orientation of horizontal gradients of air temperature and air humidity.

Due to the horizontal heat transfer, the brightness temperature variations in the ABL on the line of 1.35 cm in the area of the resonance absorption of radio waves in atmospheric water vapor in the atmospheric boundary layer can reach 30–40 K that exceeds by an order of magnitude the value of its variations caused by the processes of vertical redistribution. This result explains brightness contrasts observed from meteorological satellites which cannot be explained only with taking account of local processes of the ABL structure formation.

Close correlation is registered between the variations of brightness temperature in the ABL at the wavelength of 1.35 cm and the vertical fluxes of sensible and latent heat for different types of the ABL stratification and for different conditions of the advective transport.

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