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Original article

Features of Parameterizing Turbulent Interaction with Underlying Surface in the Regional Thermodynamic Model of Sea Ice

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Abstract
Purpose. The study is aimed at assessing the impact of choice of parameterizing the turbulent heat transfer at the ocean – atmosphere boundary upon the basic characteristics of ice regime in the Taganrog Bay apex.
Methods and Results. Thermal seasonal dynamics of the snow-ice cover thickness was studied using the non-stationary thermodynamic model of sea ice. The algorithm for defining the turbulent fluxes of momentum, sensible and latent heat in the sea ice regional model is based on the Monin – Obukhov similarity theory. The numerical experiments were performed for the winter seasons of 2007–2008 and 2017–2018, the meteorological conditions of which differed significantly. The transfer coefficients were determined both as the constant values, and as those depending on the atmosphere stratification and the aerodynamic roughness of underlying surface. Implementing stable stratification implied application of three different expressions for determining the stability functions. To avoid the iteration process required for numerical solving of the equations of the Monin – Obukhov similarity theory, the transfer coefficient parameterizations based on the approach relating these coefficients to the bulk Richardson number, were used in the model. Having been analyzed, the results of simulating the evolution of seasonal snow-ice thickness permitted to reveal the features in applying the parameterization of turbulent fluxes.
Conclusions. It is shown that the type of parameterizing the turbulent fluxes for the winters characterized by stable frosty weather and ice cover, does not impact significantly the basic elements of ice regime in the Taganrog Bay apex. However, in case the snow-ice cover is highly unstable during a season, the simulation results significantly depend on the method of determining the turbulent transfer coefficients. The best results in reconstructing the seasonal changes in ice cover thickness were obtained when using both the constant coefficients of turbulent transfer $C_H = C_E$ equal to $\approx 1.7 \cdot 10^{-3}$ and those depending on the atmosphere stratification at the ice geometric roughness equal to 8–10 cm.

Keywords: Monin – Obukhov theory, parameterization, turbulent fluxes, sea ice

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Introduction

Vertical turbulent fluxes of sensible, latent heat and momentum are the key quantitative characteristics of thermal and dynamic interaction in the ocean – atmosphere system. The heat and moisture transfer at the interface has a significant impact on processes and phenomena of different scales in the system of interacting environments. The duration of freeze-up, the rate of snow-ice cover formation and melting largely depend on the turbulent exchange of heat and moisture at the boundary of geo-environments. The difficulty in determining turbulent fluxes over the snow-ice surface is due to the complexity of field experiments and the lack of satisfactory information on the thermal state of the atmosphere and sea ice.

In thermodynamic models of sea snow-ice cover, the values of turbulent fluxes are determined using an algorithm based on the similarity Monin – Obukhov theory [1]. Integral (bulk) formulas relate the surface fluxes of momentum, sensible heat and latent heat to the measured meteorological parameters of the atmospheric surface layer. The heat and moisture exchange coefficients in the formulas can be fixed values or variables. The turbulent interaction parameterizations differ in the way in which the sea ice roughness parameters and stability functions, describing the influence of the atmosphere surface layer stratification, are determined. The temperature at the upper boundary of the snow-ice cover, not being an input parameter of the thermodynamic model, is found using an iterative process from the heat balance equation on the surface. Consequently, errors in model calculations appear already at the stage of solving the heat conduction equations for snow and ice and when finding the values of surface temperature, heat and moisture fluxes.

Rather thin unstable snow-ice cover of the seasonally freezing seas of the European territory of Russia is characterized by high spatial and temporal variability. In winter, both permanent ice cover and repeated appearance and disappearance of ice with alternating melting and freezing processes are possible in the water area. For example, the extreme instability of the Sea of Azov snow-ice cover is mainly due to contrasting weather in winter.

In the present paper, an attempt was made to evaluate the influence of the algorithm choice for calculating turbulent sensible and latent heat fluxes on the main characteristics of the Sea of Azov ice regime in the regional thermodynamic model of sea ice [2, 3]. Numerical experiments to reproduce the thermal dynamics of the snow-ice cover at the top of the Taganrog Bay were carried out for the winter seasons of 2007/08 and 2017/18, the meteorological conditions of which differed significantly.

Parameterization the fluxes of momentum, sensible heat and latent heat

The turbulent fluxes of momentum τ, sensible H and latent LE heat on the underlying surface are determined using the the bulk flux algorithm:

$$\tau = \rho C_{\mathrm{D}} U_{\mathrm{z}}^2,$$

$$H = \rho c_p C_{\mathrm{H}} U_{\mathrm{z}} (\theta_{\mathrm{surf}} - \theta_z),$$

$$LE = \rho L_v C_{\mathrm{E}} U_{\mathrm{z}} (q_{\mathrm{surf}} - q_z),$$

where U is the wind speed; θ, q is the potential temperature and specific air humidity at a given height (z index) and underlying surface (*surf* index); c_p, ρ and L_v is the heat capacity at constant pressure, density and latent heat of air vaporization; C_{D}, C_{H}, C_{E} are the coefficients of resistance, heat and moisture exchange, respectively. These resistance coefficients are related to the integral momentum, heat and moisture transfer coefficients C_m, C_h, C_q by the relations $C_{\mathrm{D}} = = C^2_m$, $C_{\mathrm{H}} = C_m C_h$, $C_{\mathrm{E}} = C_m C_q$.

The C_i ($i = m, h, q$) values in accordance with the Monin – Obukhov similarity theory are determined by the following relations:

$$C_i = \frac{\kappa}{\alpha_i - \Psi_i(\zeta) + \Psi_i(\zeta_i)}.$$

Here κ is the Karman constant; $\alpha_i = \ln\frac{z}{z_{0i}}$; the stability parameter $\zeta = \frac{z}{L}$, $\zeta_i = \frac{z_{0i}}{L}$ (L is the Monin – Obukhov length scale); z_{0m}, z_{0h}, z_{0q} are the parameters of the wind, potential air temperature and humidity roughness, respectively; Ψ_i are the integral universal functions. The dynamic roughness z_{0m} of the underlying surface is determined using semi-empirical expressions for the drag coefficient in the case of neutral stratification ($C_D = C_{DN}$) in the absence of ice and in its presence, as functions of wind speed $C_{DN} = (0.61 + 0.063U_z)\cdot 10^{-3}$ [4] and the geometric roughness of the snow-ice surface ξ (cm) $C_{DN} = (1.10 + 0.072\xi)\cdot 10^{-3}$ [5]. The z_{0h} and z_{0q} parameters for sea water were calculated according to the expressions [6]:

$$\frac{1}{\kappa}\ln\left(\frac{z_{0m}}{z_{0h}}\right) = \begin{cases} -2, & \mathrm{Re} < 0.1, \\ 4\,\mathrm{Re}^{1/2} - 3.2, & \mathrm{Re} \geq 0.1, \end{cases} \qquad z_{0h} = z_{0q},$$

for snow-ice surface – according to [7]:

$$\ln\left(\frac{z_{0m}}{z_{0h}}\right) = \begin{cases} -1.25, & \mathrm{Re} \leq 0.135, \\ -0.149 + 0.55(\ln \mathrm{Re}), & 0.135 < \mathrm{Re} < 2.5, \\ -0.317 + 0.565(\ln \mathrm{Re}) + 0.183(\ln \mathrm{Re})^2, & \mathrm{Re} \geq 2.5, \end{cases}$$

$$\ln\left(\frac{z_{0m}}{z_{0q}}\right) = \begin{cases} -1.61, & \mathrm{Re} \leq 0.135, \\ -0.351 + 0.628(\ln \mathrm{Re}), & 0.135 < \mathrm{Re} < 2.5, \\ -0.396 + 0.512(\ln \mathrm{Re}) + 0.18(\ln \mathrm{Re})^2, & \mathrm{Re} \geq 2.5. \end{cases}$$

In the formulas above $\mathrm{Re} = \frac{z_{0m} C_{DN}^{1/2} U_z}{\gamma}$ is the Reynolds number; $\gamma = (0.9065(T_z) - 112.7)\cdot 10^{-7}$ is the kinematic viscosity of air. The argument of the Ψ_i functions is the stability parameter ζ depending on the Monin –Obukhov scale, which in turn is a function of fluxes of the momentum and heat. The dependences between the stability parameter ζ and the Richardson number $\mathrm{Ri} = \frac{2zg(\theta_z - \theta_{surf})}{(\theta_z + \theta_{surf})U_z^2}$ proposed in [8, 9] make it possible to avoid the iterative solution of the similarity theory equations. Thus, according to [9], expressions ζ (Ri) for different types of atmospheric stratification have the following form:

– for unstable (Ri < 0, type I) stratification

$$\zeta = A\mathrm{Ri}^2 + B\mathrm{Ri}, \; A = a_{11}\alpha_m,$$
$$B = (b_{11}\beta + b_{12})\alpha_m^2 + (b_{21}\beta + b_{22})\alpha_m + (b_{31}\beta^2 + b_{32}\beta + b_{33}),$$
$$a_{11} = 0.0450,\; b_{11} = 0.0030,\; b_{12} = 0.0059,\; b_{21} = -0.0828,\; b_{22} = 0.8845,$$
$$b_{31} = 0.1739,\; b_{32} = -0.9213,\; b_{33} = -0.1057;$$

– for close to neutral ($0 \le \mathrm{Ri} < 0.08$, type II) and weakly stable ($0.08 \le \mathrm{Ri} < 0.2$, type III) stratification

$$\zeta = A\mathrm{Ri}^2 + B\mathrm{Ri}, \; A = (a_{11}\beta + a_{12})\alpha_m + (a_{21}\beta + a_{22}),$$
$$B = (b_{11}\beta + b_{12})\alpha_m + (b_{21}\beta + b_{22}),$$
$$a_{11} = 0.5738,\; a_{12} = -0.4399,\; a_{21} = -4.901,\; a_{22} = 52.50,$$
$$b_{11} = -0.0539,\; b_{12} = 1.540,\; b_{21} = -0.6690,\; b_{22} = -3.283;$$

– for stable ($\mathrm{Ri} \ge 0.2$, type IV) stratification

$$\zeta = A\mathrm{R}i + B, \; A = a_{11}\alpha_m + a_{21}, \; B = b_{11}\alpha_m + b_{21}\beta + b_{22},$$
$$a_{11} = 0.7529,\; a_{21} = 14.94,\; b_{11} = 0.1569,\; b_{21} = -0.3091,\; b_{22} = -1.303.$$

Here $\alpha_m = \ln\frac{z}{z_{0m}}$, $\beta = \ln\frac{z_{0m}}{z_{0h}}$.

Numerical simulation results

The numerical experiments that reproduce seasonal thermal dynamics of the snow-ice cover at the top of the Taganrog Bay for the winter seasons of 2007/08 and 2017/18 were carried out using a locally one-dimensional thermodynamic model of sea ice, taking into account accumulation and melting of snow on its upper boundary. Vertical temperature profiles in snow and ice layers are found by solving non-stationary heat conduction equations with a radiation source. The equations for the balances of heat and mass fluxes are the boundary conditions on the upper and lower surfaces of the snow-ice cover. At the upper boundary, the heat flux consists of turbulent sensible and latent heat fluxes, long-wave and short-wave radiation balances of the surface and heat fluxes associated with cooling processes and subsequent possible crystallization of liquid precipitation. The weakening of the solar radiation intensity in the ice layer is described by a two-layer scheme, in the snow thickness – by the Bouguer – Lambert law. The long-wave radiation balance is determined taking into account total cloudiness. At the snow-ice boundary, the conditions for the heat and temperature continuity are satisfied. In the absence of ice, the change in sea water temperature is determined by the balance of heat fluxes on the sea surface. The scheme for describing the snow layer dynamics includes taking into account the phase nature of precipitation, changes in the freshly fallen snow density depending on air temperature and wind speed, changes in thermal conductivity and albedo of snow, and snow transformation into ice. The heat capacity, thermal conductivity, latent heat of snow and ice melting are determined using the empirical dependences of these quantities on temperature and salinity. The mathematical formulation of the problem, calculation methods and numerical scheme for its solution, the form of semiempirical dependencies, the values of parameters and coefficients used in calculations of the thermal dynamics of snow-ice cover are given in [2, 3].

The data of eight-term observations of the main meteorological parameters (surface temperature, atmospheric pressure, humidity, total cloudiness, wind speed and total precipitation) of RIHMI-WDC [1] at the Taganrog meteorological station were used as an external forcing. A comparison of the simulation results of the thermal evolution of the snow-ice cover using various parameterizations of the turbulent heat and moisture exchange processes was carried out for winters with significantly different meteorological conditions.

Thus, the 2007/08 winter can be classified as relatively severe and one of the most ice-covered in recent years. The air temperature dropped to –19 °C, the sum of average daily negative temperatures was –311 °C. Due to the long exposure of the eastern part of the Taganrog Bay under the Siberian anticyclone influence, the ice thickness here significantly exceeded the norm, and the duration of the ice period was more than 70 days. In addition, this winter can be considered dry and snowless. The total amount of atmospheric precipitation from October 2007 to April 2008 was half the norm and amounted to only 170 mm.

The 2017/18 winter, in accordance with the accepted typification, can be classified as warm and humid. The sum of average daily negative temperatures was –162 °C, the total amount of atmospheric precipitation for the season was 339 mm. December 2017 was very warm, the cooling came only in mid-January 2018. However, the established ice cover came apart in the second decade of February, and frosty weather in late February – early March contributed to its reappearance.

Two series of calculations were carried out. In the first series, constant drag coefficients were used in the evaluation of turbulent sensible and latent heat fluxes ($C_\mathrm{H} = C_\mathrm{E} = 1.2 \cdot 10^{-3} - 2 \cdot 10^{-3}$); in the second series, the exchange coefficients obtained taking into account the type of stratification in the surface atmospheric layer and the underlying surface roughness were used. For unstable stratification conditions, the stability functions proposed by Högström [10] were used. For a stable atmosphere, three options for calculating turbulent fluxes were considered using the Beljaars – Holtslag (hereinafter BH) [11], Cheng – Brutsaert (hereinafter CB) [12], and SHEBA [13] stability functions. When choosing the roughness coefficient value, its dependence was assumed only on the underlying surface shape (geometrical roughness of sea ice ξ). In calculations, ξ varied in the range of 1–10 cm.

[1] Bulygina, O.N., Veselov, V.M., Razuvaev, V.N. and Aleksandrova, T.M. [*Description of the Urgent Data Array on the Main Meteorological Parameters at the Stations of Russia*]. 2014. [online] Available at: http://meteo.ru/data/163-basic-parameters#описание-массива-данных [Accessed: 06 July 2023] (in Russian).

Table 1 shows the number of cases of implementation (in %) of unstable (type I), close to neutral (type II), weakly stable (type III) and strongly stable (type IV) atmospheric stratification, as well as average values of C_H and C_E coefficients for the presented calculation options. It can be seen that, despite the differences in meteorological conditions, the ratios of stratification types for the winters of 2007/08 and 2017/18 are similar to each other. According to the calculations, both for cold and warm winter seasons, in more than 85% of cases, an unstable or close to neutral type of stratification is realized.

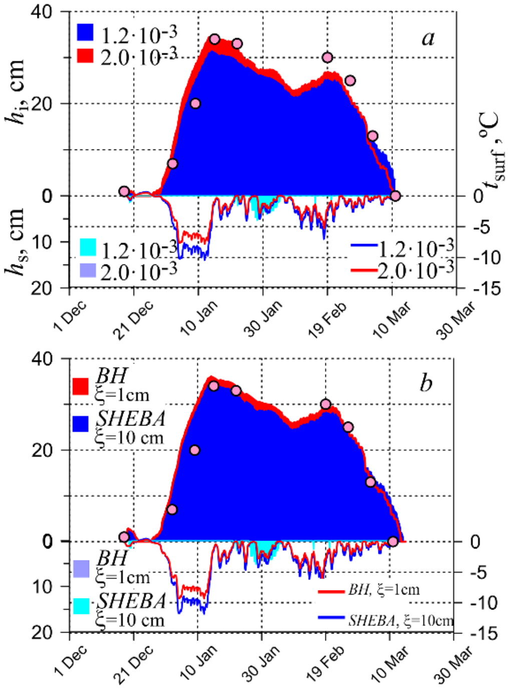
T a b l e 1

Proportion of the cases of implementing the types of atmosphere stratification, and average values of the C_H and C_E coefficients for the presented calculation variants

Calculation	Conditions of calculation		$C_H \cdot 10^3$	$C_E \cdot 10^3$	Proportion of the cases of implementing the stratification types			
					I	II	III	IV
2007/08								
The first series	$C_H = C_E$ = const		1.2		47	41	7	5
			2.0		46	43	6	5
The second series	ξ, cm	1	1.51	1.56	46	40	8	6
		5	1.79	1.83	42	46	7	5
		10	2.03	2.07	42	46	7	5
2017/18								
The first series	$C_H = C_E$ = const		1.2		48	45	6	1
			2.0		48	46	5	1
The second series	ξ, cm	1	1.56	1.60	45	49	5	1
		5	1.85	1.87	45	49	5	1
		10	2.07	2.11	45	49	5	1

N o t e. Designations: type I – unstable stratification (Ri < 0); type II – close to neutral one ($0 \leq$ Ri < 0.08); type III – weakly stable one ($0.08 \leq$ Ri < 0.2) and type IV – strongly stable one (Ri ≥ 0.2).

The conditions for stable (types III, IV) stratification appear much less frequently. The share of such cases (of the total number of calculated time series) is ~ 13% for the 2007/08 season and ~ 6% for the 2017/18 season. Note that in the warm winter of 2017/18, the stable stratification was noted mainly during the period of intense melting, and in the winter of 2007/08, the number of such occurrences was approximately equally distributed between the periods of melting and cooling of the snow-ice cover.

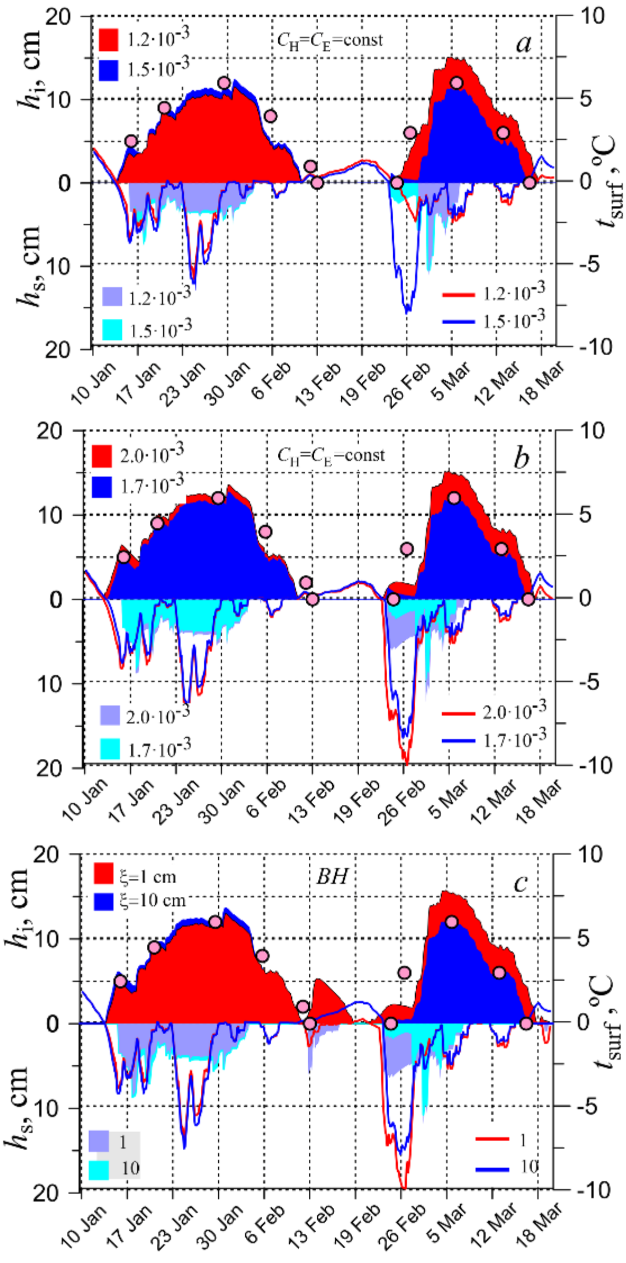


F i g. 1. Results of modeling thermal evolution of snow-ice cover in winter 2007–2008 using different parameterizations of the processes of turbulent heat and moisture transfer. Shaded areas above and below the zero mark illustrate seasonal variation of ice thickness h_i and snow depth h_s, respectively; solid lines show the diurnal average surface temperature t_{surf} of snow-ice cover

Fig. 1 and 2 show the seasonal course of the ice thickness h_i, the snow depth h_s and the surface temperature t_{surf} of the snow-ice cover for the winters of 2007/08 and 2017/18. Fig. 1, a illustrates the calculation results of the first series of the experiments (constant coefficients), Fig. 1, *b* – the second series (variable coefficients), obtained considering stratification of the atmosphere and the geometric roughness of sea ice. The figures for each of the series show cases corresponding to the calculation options in which for a given season the largest and smallest development of a snow-ice cover is realized. The solid circles show sea ice thickness taken from ESIMO [2] maps. Table 2 shows an average forecast error $ME = \frac{\sum_l (h_{obs} - h_i)}{l}$ and standard deviation h_i from h_{obs} $RMSE = \sqrt{\sum_l \frac{(h_{obs} - h_i)^2}{l}}$ for the considered calculation options, where l is the length of the time series.

2018

[2] ESIMO. *Unified State System of Information on the Situation in the World Ocean (ESIMO)*. 2023. [online] Available at: http://193.7.160.230/web/esimo/azov/ice/ [Accessed: 06 July 2023].



F i g. 2. Results of modeling thermal evolution of snow-ice cover in winter 2017–2018 using constant coefficients of turbulent exchange (*a*, *b*) and stability functions *BH* (*c*) at ξ taking on the values 1 and 10 cm. See designations in Fig. 1

It can be seen that for the conditions of the 2007/08 winter, when the ice cover formed as a result of prolonged cold weather remained for almost the entire season, and the snow cover was insignificant and short-lived, the calculation results are close to each other both within the series and between them. Thus, during the period of maximum ice cover development, the difference in ice thickness within the first and the second series of experiments was ~ 10 and 5%, respectively. It should be noted that the results of calculations of the second series at 5 cm ≤ ξ ≤ 8 cm were closest to the field data of ice thickness. Somewhat greater differences were manifested in the calculated surface temperature; for both series, they reached 20–25% during the period of intense ice formation.

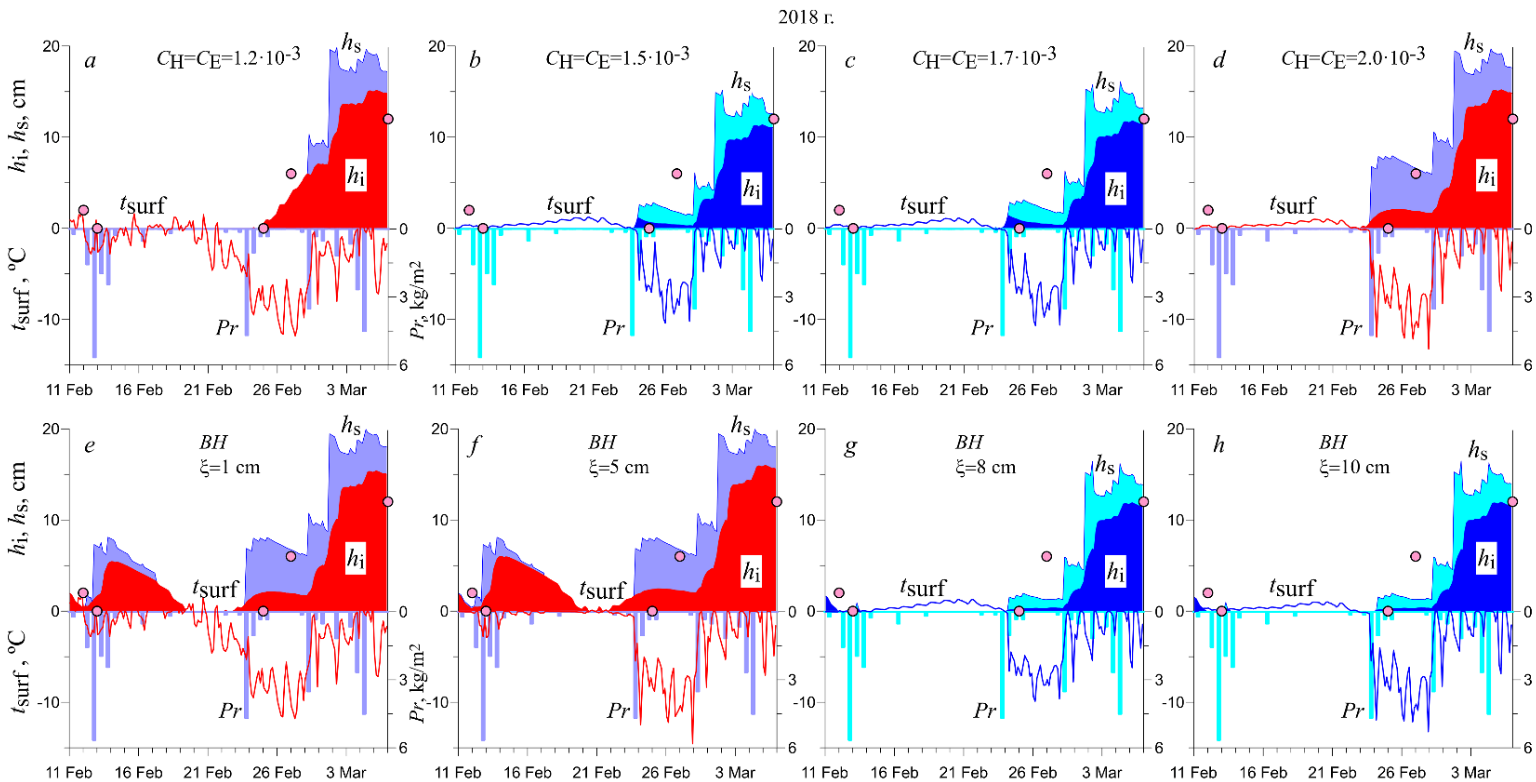
Table 2

Quantitatively estimated quality of forecasting seasonal variation of ice thickness using different parameterizations of heat and moisture transfer processes

Calculation	Conditions of calculation		Forecast errors, cm	
			ME	*RMSE*
2007/08				
The first series	$C_H = C_E$ = const	$1.2 \cdot 10^3$	1.3	2.9
		$1.5 \cdot 10^3$	0.5	2.6
		$1.7 \cdot 10^3$	0.1	2.7
		$2.0 \cdot 10^3$	–0.3	3.2
The second series*/TT	ξ, cm	1	–1.8	3.2
		10	–2.1	3.5
2017/18				
The first series	$C_H = C_E$ = const	$1.2 \cdot 10^3$	–0.9	2.7
		$1.5 \cdot 10^3$	0.3	2.1
		$1.7 \cdot 10^3$	0.1	2.1
		$2.0 \cdot 10^3$	–1.3	2.9
The second series*	ξ, cm	1	–2.5	3.7
		10	–0.1	2.1

* In the second series, for ξ = 1 cm and ξ = 10 cm, the stability functions *BH* and *SHEBA* were applied, respectively.

The winter of 2007/08 was relatively severe and icy, which led to the presence of an ice cover almost from the beginning of ice formation to its complete destruction. However, in most moderate and warm winters, the snow-ice cover is extremely unstable and the sea water area can be cleared of ice and covered with it again several times during the season. The numerical experiments have shown that for such winters, in particular for the winter of 2017/18, the choice of the method for parameterization of turbulent exchange processes can significantly affect the simulation results and lead to significant differences in determining the characteristics of the ice cover during clearing and refreezing periods.



F i g. 3. Illustration of influence of a type of parameterizing the processes of turbulent heat and moisture transfer upon reconstructing the thermal evolution of snow-ice cover in winter 2017–2018 during the melting and refreezing period. Shaded areas illustrate seasonal variation of ice thickness h_i and snow depth h_s, solid lines – surface temperature t_{surf} of snow-ice cover, and bar graph *Pr* – total for every 3 hours water equivalent of precipitation

According to Fig. 2, for the first half of the winter season (from the first ice formation to the first complete clearing), the calculated ice thickness values obtained using constant (Fig. 2, *a*, *b*) and variable (Fig. 2, *c*) C_H and C_E coefficients are close to each other. A noticeable dependence of the simulation results on the C_H (C_E) values appears during the open water period and then during refreezing. Both at small (less than $1.5 \cdot 10^{-3}$) and large (over $1.8 \cdot 10^{-3}$) values of the variable coefficients, a noticeable (up to 35%) overestimation of the ice thickness is observed compared to the observed values (Fig. 2, *a*, *b*). Moreover, in both cases, atmospheric precipitation has a significant impact on the snow-ice cover formation.

In the period from the evening of February 23rd to the morning of February 25th, snowfall was observed (slightly more than 6 kg/m^2 of *Pr* water equivalent), and in calculations with low exchange coefficients, most of it fell on open water, since ice formation began only in the evening of February 24th (Fig. 3, *a*). The absence of any significant precipitation during the next three days at low (~ –10 ºC) air temperatures contributed to a fairly rapid ice growth. On the contrary, when modeling very intense heat exchange with the atmosphere (Fig. 3, *d*), a primary ice cover was formed almost a day before the start of precipitation and snow fell on the already formed ice, partially turning into snow ice. The ice cover characteristics closest to field data are reproduced at $C_H(C_E) \approx 1.7 \cdot 10^{-3}$. As for the options with the definition of turbulent flows calculated using the three stability functions under consideration, for some selected values of geometric roughness ξ, overestimated values of h_i were observed, and the appearance of "model ice" was also noted during the open water period (Fig. 3, *e*, *f*). A similar effect in the seasonal ice evolution modelling was manifested for the stability functions *BH* at $\xi < 3$ cm, *CB* – at $\xi < 7$ cm, *SHEBA* – at $\xi < 4$ cm. Calculations of h_i at $8 \le \xi \le 10$ cm were closest to the measured values, and for any of the three types of stability functions.

Conclusion

Based on the numerical experiments to simulate seasonal variation of ice thickness at the top of the Taganrog Bay for the 2007/08 and 2017/18 winter seasons, the prevailing types of atmospheric stratification are determined. It is shown that in more than 85% of cases, an unstable or close to neutral type of stratification is realized. A series of calculations were carried out using constant coefficients of turbulent exchange and the coefficients obtained considering stratification of the near-ice atmosphere. It is shown that for conditions of prolonged cold weather and in the presence of an ice cover for almost the entire season, the choice of any of the considered types of turbulent interaction parameterization with the underlying surface does not make significant differences in determining the maximum seasonal ice thickness, as well as the dates of freezing and clearing. However, for conditions of extreme instability of the snow-ice cover, the dependence of the simulation results on the method of determining the turbulent exchange coefficients can be quite noticeable. At small (less than $1.5 \cdot 10^{-3}$) and large (more than $1.8 \cdot 10^{-3}$) values of the coefficients, a noticeable (up to 35%) overestimation of the ice thickness is monitored compared to the observed values.

The calculations using both constant coefficients of $C_H = C_E$ turbulent exchange, which take values of $1.5–1.7 \cdot 10^{-3}$, and coefficients obtained taking into account the atmosphere stratification at geometric ice roughness values of 8–10 cm, give quite satisfactory results. However, for a more accurate simulation of values of the heat fluxes at the ice cover upper boundary, more careful selection of model parameters and comparison of the results of model calculations with the corresponding data of field measurements are required.

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