Modeling of the Novorossiysk bora. Part 2. Energetics of the atmosphere at the Novorossiysk bora

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Important features of the Novorossiysk bora are being investigated using a nonlinear analytical stationary two-dimensional model. Vertical unboundedness of the atmosphere, characteristic features of the shape of the streamlined mountains, the effect of the upper layers on the disturbances in the troposphere are considered. For the first time disturbances are being studied in a wide range of changes in the features of an incident flow. It has been shown that disturbances quite often assume a rotor nature in the leeward zone behind the mountains, and the dependence of disturbances on the wave scale is decreasing only on average, and noticeable deviations from it are oscillatory in nature and depend on the level of reflection of wave energy from the upper layers. Jet stream velocity at the leeward slope estimates energetics of disturbances behind the mountains. It has been shown that its value lies in the range of 18–22 m/s under normal conditions. It has been testified that a sharp cooling in event of a bora is determined not by the process of flowing around the mountains, but by the arrival of a colder air mass. It has been concluded that an increase in speed indicator in the surface layer turbulent layer beyond the mountains is determined by the height and steepness of the leeward slope of the mountains, the effects of turbulence behind the mountains and is very sensitive to changes in the flow velocity at the leeward slope of the mountains. *KEYWORDS:* Atmosphere physics; hydrodynamics; modelling; mountain flow; wave frequency; Lira wave scale; Novorossiysk bora.

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1. Introduction

The research continues to study the features of atmospheric disturbances in event of Novorossiysk bora on the basis of a nonlinear analytical modeling (Part 1, [*Kozhevnikov et al.,* 2020]). The main attention is directed to the study of wind and temperature disturbances leading to catastrophic consequences. In recent years, the phenomenon of mountain flow has been investigated using numerical simulation methods. According to [*Bedanokov et al.,* 2018] in the works [*Efimov and Barabanov,* 2013; *Gavrikov and Ivanov,* 2015; *Shestakova et al.,* 2015; *Toropov and Shestakova,* 2014; *Toropov et al.,* 2013] it is shown that the Novorossiysk bora can be successfully predicted. When using analytical modeling, the solution of the problem can be narrowed down to obtaining specific formulas and, at the same time, to the control of the accuracy of the results obtained. This can be achieved by taking into account only the most important fac-

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tors of the phenomenon. When applying numerical modeling, it is possible to take into account many more factors of the phenomenon, but it is necessary to introduce a number of parametric simplifications and, as a result, it is very difficult to assess the true accuracy of the results. The studies have shown that this natural phenomenon is so difficult that it is impossible to do without the use of both modeling methods. In addition, it has been found that it is necessary to expand the spatial range of measurements of the characteristics of atmospheric disturbances significantly, especially at different altitudes.

In this paper we will use some results of both [*Berzegova and Bedanokov,* 2018; *Berzegova et al.,* 2017] and [*Efimov and Barabanov,* 2013; *Gavrikov and Ivanov,* 2015; *Shestakova et al.,* 2015; *Toropov and Shestakova,* 2014; *Toropov et al.,* 2013].

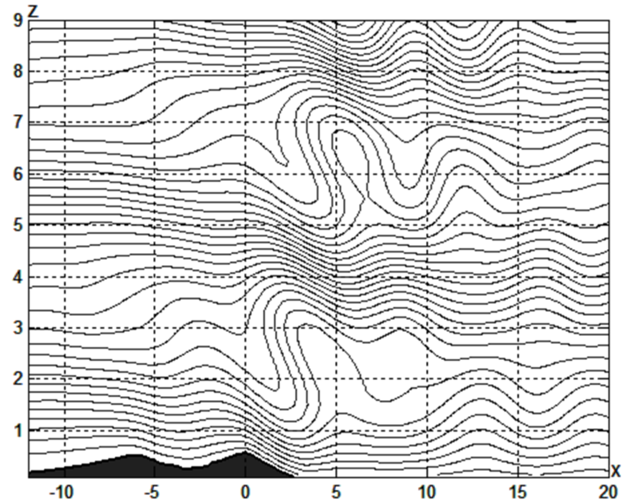

Figure 1. The trajectories of motion during a flow around an average relief at $\lambda_c = 4$ in the area of $0.1 < z < 9$, $-12 < x < 20$ km. The relief is blackened. Values of z_0 change with a step of 0.25 km.

2. The Model

The results were obtained when using a hydro thermodynamic analytical nonlinear two-dimensional modeling, described in detail in Part 1 [*Kozhevnikov et al.,* 2020] and [*Berzegova and Bedanokov,* 2018; *Berzegova et al.,* 2017; *Kozhevnikov,* 2019]. In the model the atmosphere was represented in the form of three layers, while the 10 km thick lower layer represented the troposphere, the next 8 km thick – the lower stratosphere, the latter – the rest of the atmosphere. The flow function fields of $\psi(x, z)$ (trajectories), velocity and temperature disturbances were calculated in $-30 < x < 40$ coordinate space and to heights of 30 km. Perturbations arising during the flow around mountains of four forms obtained in Part 1 [*Kozhevnikov et al.,* 2020] as a result of special processing of the height map of the region under consideration were analyzed. The *sr* relief, which characterizes the averaged two-dimensional features of the mountains was the main object. It included two main ridges of 511 m and 548 m and a hollow between them 5.5 km wide and almost 258 m deep. We call the highest ridge the coastal one. Its total length was about 25 km. The second was the *ch* relief, which was among the averaged forms and most noticeably differed from *sr* among the latter. The main features of this relief are shown in the lower part of Figure 1. The height of its coastal ridge was lower than the average relief by 127 m. Two more reliefs, $iskN$ and $iskV$, were created artificially. They had smooth windward and steep leeward slopes and one prevailing height: the first was of 350 m, the second was of 548 m. All reliefs had almost the same section areas and the steepness of the leeward slope. Therefore, they made it possible to investigate specifically the influence of the disturbances on the dominant ridges and the maximum mountain heights. The dependence of the disturbances on the properties of the incident flow in front of the mountains was investigated. Its speed was set the same, and γ vertical temperature gradients were different in the layers. As a result, characteristics of the incident flow were set in the following range:

$$\lambda_c = 3, 4, 5, 6, 6.66, 7, 7.5, 7.8, 9.5, 10, 12.2 \text{ km},$$

$$\gamma_j = 6, 0, 3 \text{ degrees/km} \tag{1}$$

where γ_j stands for bottom-up gradient values, λ_c – the values of the Lyra wave scale in the troposphere. The latter is defined by the ratio [*Lyra,* 1943]:

$$\lambda_c = 2\pi \frac{U}{N}$$

where N is Brent-Väisäl wave frequency, defined in a standard way through a temperature gradient.

The values given for λ_c variations correspond to U variations in the range from 6 to 24.4 m/s.

3. The Results

3.1. Figure 1 shows one of the results of modeling. Here, the solid lines show trajectories in the troposphere. The latter are identified by z_0 values of their heights in the incident flow, directed from left to right. The streamlined relief is painted over. The disturbances are great at all heights, but especially over the mountains. Here the trajectories with $z_0 = 2 - 2.75$ and $6 - 6.6$ have the character of developed rotor circulations [*Long,* 1955] with amplitudes of about 1 km. In the leeward region, disturbances are wavelike. At altitudes of 1–1.5 km above the mountains their amplitudes are about 250 m, they quickly decrease downstream, remaining significant in the surface layer: for trajectories with $z_0 = 0.35$ amplitudes above the sea make up about 50 m at distances of up to 20 km.

3.2. In our work, we will be interested in wind increase in the streams above the mountains [*Berzegova and Bedanokov,* 2018; *Berzegova et al.,* 2017; *Kozhevnikov,* 1999, 2019]. This increase is almost proportional to the density of the trajectories.

The flows above the mountains near the earth are of the greatest interest, as the particles go smoothly downwards when moving. There are two such flows, they appear above the leeward slopes of the main ridges, but the second downstream is much sharper. Similar flows can be seen also at altitudes of 3.5–5 and 7.2–8.5 km, but here the influence of the first ridge is no longer noticeable. The velocity module here is about 20 m/s. The intensity of the flows exacerbates especially with the appearance of rotors. Here flows appear where the particles rise abruptly upwards or even partially towards the main stream. The vertical speed here reaches 3 m/s. In the upper and lower parts of the flows, the convergence of the trajectories has different signs. This means that the speeds in the flows are maximum at a certain average height.

In the study of the bora, disturbances near the leeward slope of the coastal ridge are of the greatest interest. In this area, fields of ψ, speeds, and T' disturbances were calculated. Calculations were carried out for all reliefs and for all parameters (1).

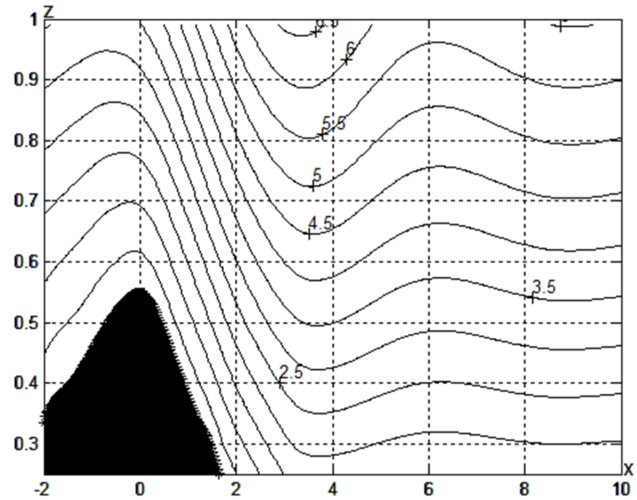

Figure 2. Temperature disturbances at the leeward slope at $\lambda_c = 5$, $\gamma = 6$, $U = 10$. Isolines are digitized with a step of 0.5 degrees.

It was found that the patterns of disturbances in all variants are qualitatively similar and this fact made it possible to illustrate the results in Figure 2. Here the isolines of T' disturbances are presented for the variant of $\lambda_c = 5$.

At the same time these isolines give a qualitative idea of the trajectories of motion. For example, the isoline for $T' = 2.5$ degrees reproduces the trajectory with $z_0 = 0.9$. The Figure 2 shows that the flow along the slope is almost homogeneous and has a thickness of about 1 km. At a sufficient distance from the slope, the movement of air particles acquires a wave character. This is clearly seen in Figure 1. The presented data show that there are practically no T' disturbances directly at the slope, and when moving away from it, they become positive. Here, in all variants in the air flow at a slope of about 1 km thick and in a surface flow behind the mountains of 300 m thick, the disturbances are especially positive and lie in the range of 0–1.5 degrees. This corresponds to the law of adiabatic dropping of air particles here. This means that the known sharp temperature drops during bora are determined mainly not by the laws of physics of the flow around the process, but by the fact that an air mass of a different temperature enters the area of a city Novorossiysk and the sea.

3.3. The results of model calculations of disturbances at altitudes above 2 km have been confirmed by direct measurements in [*Berzegova et al.,*

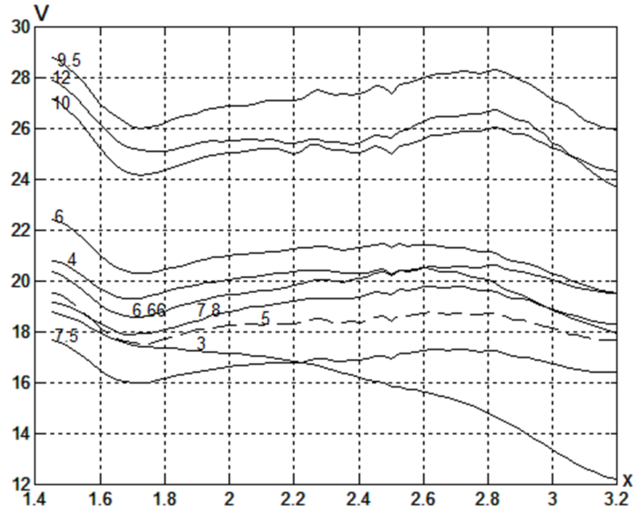

Figure 3. Changes in the modulus of flow velocity with distance from the slope (km). The numbers on the curves show the values of λ_c.

2017; *Kozhevnikov,* 1999, 2019; *Kozhevnikov et al.,* 1986]. It is evident that the used model cannot claim to reproduce the characteristics of disturbances in the surface layer behind the mountains, since it does not take into account the viscosity and the turbulent medium viscosity. It is also clear that the jet stream at the leeward slope of the mountains is not purely laminar. However, taking into account the hydrostatic stability of the air and high velocity of the particles, we will assume that the energetics of this flow does not alter the turbulent processes too much. This gives a basis to estimate the intensity of the bora in the city Novorossiysk and bay in the first approximation according to the calculated stream velocity in the jet at the leeward slope of the mountains.

The velocity field at the slope has been studied in detail for the average relief. First of all, changes in the values of $V(z,x)$ velocity modulus have been determined depending on x at the level of $z = 300$ m, which can be considered characteristic. Graphs $V(z,x)$ in m/s shown in Figure 3 show that the nature of the changes in speed is the same with distance from the mountain. Initially, the speed decreases monotonously (by about 1–2.5 m/s in the range of distances up to 200 m). Then, its values begin to change wavelike with amplitudes of about 1.5 m/s and a period of λ_c.

In this research the jet energy at the leeward slope has been estimated by the average value of $V(z,x)$ in the layer of 200 m thick. This value should obviously characterize the properties of the jet at a particularly chosen height. If the averaging is carried out at the level of maximum speeds, then the resulting value will depend on the specific value of λ_c only. The resulting value will be denoted as V_b, and will be called the characteristic speed of the jet. It will give an idea about both the energy of the flow at the slope, and the intensity of disturbances beyond the mountains.

3.4. It's worth noting that air particles do not reach the earth, as they descend along the slope from the uppermost layers of the jet and therefore, form a transition layer between the free atmosphere and the surface layer of interest to us behind the mountains. The air particles from the lowermost layers of the jet are obviously involved in the formation of the layer and its turbulation process. It becomes clear that the energy of turbulent gusts in the surface layer beyond the mountains is directly proportional to the kinetic energy of the jet, and, therefore, is determined by the value of V_b. Calculations of V_b have been carried out for all variants. Here, averaging of $V(z,x)$ along the horizontal coordinate has been carried out at such two height levels at which they had maximum values. According to two values obtained this way, the average value of V_b and dV scatter band have been determined. It's been established that dV value does not exceed 0.17 m/s. Figure 4 illustrates the obtained dependences on λ_c values in the incident flow. Here, asterisked curves represent smoothed data for the average relief. Figure 4a shows changes of a relative value of V_b/U. They allow to note the following:

1. For all λ_c, the value of V_b/U is greater than 1.
2. On average, a decrease of V_b/U with an increase of λ_c is observed, i.e. the previously formulated smoothing pattern is fulfilled (see Part 1[*Kozhevnikov et al.,* 2020]).
3. In some parts of λ_c range changes are not monotonous. This is especially noticeable in the neighborhood of values of $\lambda_c = 5$, 7.5, 9.5. When approaching extremely large values of λ_c, they smoothly approach 1.
4. The curve for the *ch* relief does not qualitatively differ from the curve for *sr*; quantita-

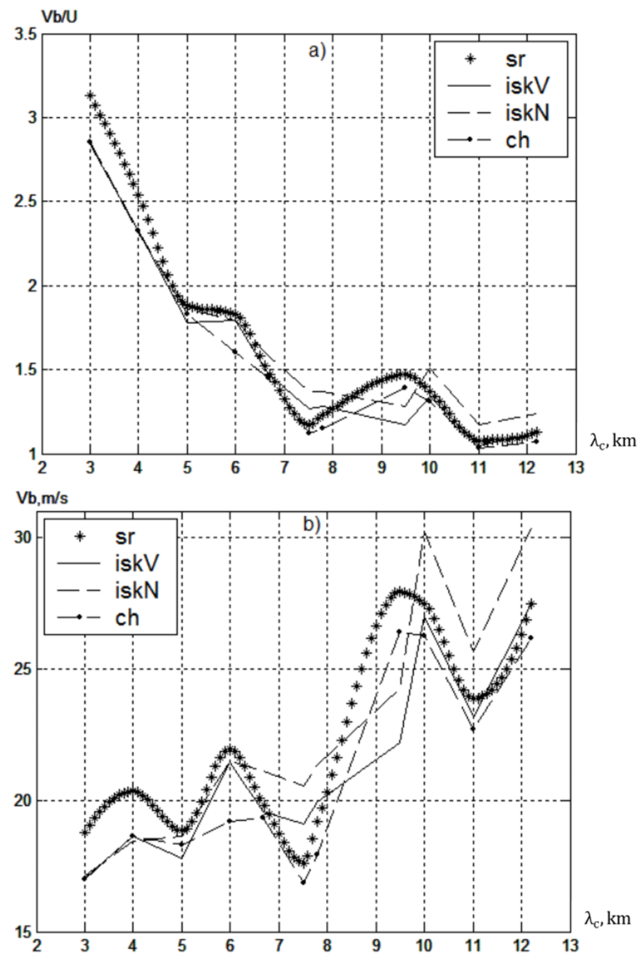

Figure 4. Dependence of the characteristic speed on λ_c for different reliefs: a) for V_b/U, b) for V_b.

tively, it differs only in that it does not reproduce an increase of V_b/U in the region of $\lambda_c = 6$. Hence, the results for the average relief are quite suitable for the study of the bora.

5. Reliability of the revealed dependences is confirmed by smallness of dV.

Figure 4b shows how values of V_b change when increases (or U speed). These changes have the form of noticeable fluctuations relative to a smooth and fairly rapid increase on average. This increase corresponds to the smoothing pattern discussed previously. In Part 1 [*Kozhevnikov et al.,* 2020] it has been obtained as a result of analyzing changes in the amplitudes of disturbances, higher – changes in relative speed of V_b/U. Since in the latter case the ratio of the speed to the speed of the incident flow is investigated, we can assume that the smoothing pattern characterizes a certain coefficient of increase of the characteristic speed under the influence of the effect of the flow around the mountains, and, therefore, the coefficient of amplification of disturbances. This coefficient increases when the Lyra scale decreases: when $\lambda_c = 3$ its value is of the order of 3, as λ_c increases, its values asymptotically approach 1. Taking into account this fact, it turns out that the curves in Figure 4b at the far end of the increase of λ_c should be asymptotically transformed into straight lines $V_b = \lambda_c + \text{const}$. Hence, Figure 4 illustrates the same pattern of smoothing disturbances from different sides when λ_c increases. This pattern, as in Part 1 [*Kozhevnikov et al.,* 2020], can be formulated as follows. Since the λ_c value is primarily determined by the speed of the incident flow, the time of interaction of the flow with irregularities decreases when λ_c increases, and, so does the disturbing influence of the latter.

3.5. In this model, the reflection of wave energy from the upper layers is determined by λ_c, vertical thickness of the atmospheric layers, the size and the shape of the relief. Figure 4 does not illustrate the fact that the form of the theoretical relief is always somewhat different from the given one in the model and this difference changes when λ_c increases. This change is controlled by dh constant of about 25 m, while ensuring that the greatest changes occur only in the region of principal heights. It turns out that each line in Figure 4 represents a certain set of reliefs, the differences in maximum heights in which do not exceed dh. The maximum heights of the *sr* and *ch* reliefs differ by 127 m, i.e. by about 5 dh values. This suggests that the scatter band of V_b values of the lines for *sr* and *ch* in Figure 4 is about 5 times less than the difference in the heights of the lines themselves. Hence, the amplitudes of the wave deviations of V_b from the average smoothing pattern are largely determined by the level of reflection of the wave energy from the upper layers.

3.6. Earlier it was established in [*Long,* 1955] that the details of the relief shape in its windward part affect less the resulting disturbances than the details of the shape in the leeward part, and the height and steepness of the leeward ridge is of great importance here. In the reliefs considered in the research, the steepness of the leeward slope was almost the same, and the heights were different. Besides, the *sr* and *ch* reliefs have two dominant ridges, while the others have only one. Now let's analyze the differences between the curves shown in Figure 4.

1. We draw attention to the fact that the curves for the *sr* and *ch*, *iskV* and *iskN* reliefs are pairwise close to each other. This can be viewed as the evidence that the nature of the V_b wave changes definitely depends on the indicated features of the relief shape.

2. For small values of the Lyre scale (for $\lambda < 6.2$), the *iskV* and *iskN* curves differ little. In almost the same range of λ_c values, the *sr* and *ch* curves differ noticeably, and the *ch* curve lies below. Obviously, this is determined by the fact that the height of the leeward ridge of *ch* is lower than that of *sr*.

3. Comparing the curves for *iskV* and *iskN*, we see that when $\lambda_c > 6.3$ the first curve (for a higher relief) lies below everywhere. This, apparently, is due to the fact that for a lower relief the pattern of smoothing disturbances begins to overcome the influence of the relief shape earlier than that of a high one. When comparing the results for the *sr* and *iskV* reliefs when $\lambda_c = 5$ in Part 1 [*Kozhevnikov et al.,* 2020], it was noted that the maximum range of vertical displacements in the rotor region in the second case decreased by 38%. Figure 4 shows that in the region of $\lambda_c < 6.7$ speed difference signs remain the same, but the significance of differences is about 5%, i.e. almost 7 times less. This indicates that the presence of a hollow in the relief has a much stronger effect on disturbances in the middle troposphere than on the leeward slope.

3.7. In Part 1 [*Kozhevnikov et al.,* 2020] the role of the dynamic interaction of layers was partially revealed when the results of trajectory calculations were compared at $\lambda_c = 5$ for three options for setting the gradient in the troposphere: $\gamma = 5$, 6 and 7. The analysis showed that when going from the option of $\gamma = 6$ to $\gamma = 5$, the wave energy reflection coefficient significantly decreased at the tropopause. This effect was noticed by an increase in the amplitudes of the waves in the upper layers. In this paper it was verified by analyzing changes of V_b at a point of $\lambda_c = 5$. Calculations showed that when $\gamma = 6$ then $V_b = 18.8$, when $\gamma = 7$ then V_b increases by 4, and when $\gamma - 5$ it decreases by 2.8. Therefore, the characteristic speed at the slope responds to changes in the energy exchange between the layers more clearly than the field of trajectories. In the model, the reflection coefficient on the interfaces at a specific value of λ_c is determined only by the discontinuity of the temperature gradients. The dependence of this coefficient on λ_c was not determined, but it could be assumed that it was not so great. Then we should expect that with a change of γ fiber values in (1) the values of V_b presented in Figure 4 will just slightly shift vertically. This means that the characteristics of the bora energy in the first approximation are not set specifically for the specified values of γ, but for a certain range of their values.

3.8. According to [*Efimov and Barabanov,* 2013; *Gavrikov and Ivanov,* 2015; *Shestakova et al.,* 2015; *Toropov and Shestakova,* 2014], devoted to the numerical simulation of the bora, according to measurements in the incident flow, wind speed values in most cases were in the range of 3–10 and rarely reached 15 m/s. Therefore, not the whole range (1) is of practical importance, but only the next part of it: $\lambda_c = 3 - 7.8$ km, or $U = 6 - 15$ m/s. The functions of $V_b(\lambda_c)$ presented in Figure 4, had relatively small oscillation amplitudes with a change in the Lyra scale in this range, so that the values of the characteristic speed here could be represented approximately as $V_b = (20 \pm 2)$ m/s.

3.9. The used model does not allow to calculate the velocity field in the surface layer beyond the mountains. However, the kinetic energy of disturbances here will be determined by V_b and, in particular, it can be assumed that the average wind speed should be close to it. The bora measuring data presented in [*Efimov and Barabanov,* 2013; *Gavrikov and Ivanov,* 2015; *Shestakova et al.,* 2015; *Toropov and Shestakova,* 2014] give some idea of this. It is reported in [*Efimov and Barabanov,* 2013] that the maximum speeds in the bay in event of bora were more than average ones by 5 m/s. It is recorded in [*Gavrikov and Ivanov,* 2015] that the wind speed in the bay increases by 5–10 m/s in event of gusts. There is no information on direct measurements of the characteristics of turbulence, but the presented data indicate the turbulent nature of disturbances in the surface layer. There is little data on the difference in the speeds on the leeward slope of the mountains from those in the bay. It is stated in [*Toropov and Shestakova,* 2014] that during the bora on 27 January 2012, the wind speed on the leeward slopes was 14 m/s, and at two stations in the surface layer beyond the mountains maximum values of average speed reached 35 and 25 m/s, i.e. speed values increased 2–2.5 times. This suggests that the obtained estimation of V_b has been made correctly and one can judge of it by wind increase in event of bora behind the mountains.

3.10. The conclusion that V_b practically does not depend on the properties of the incident flow during further research deserves special attention. The measurement results presented in [*Efimov and Barabanov,* 2013; *Gavrikov and Ivanov,* 2015; *Toropov and Shestakova,* 2014] indicate noticeable changes in the wind speed in the bay during the bora. All this suggests that the process of turbulation of movements in the surface layer behind the mountains opposite is very sensitive to changes in V_b and, therefore, changes in the properties of the incident flow.

One should bear in mind that the calculations considered only the characteristic two-dimensional features of the relief. A more detailed examination of the isolines of heights of the terrain presented in Part 1 [*Kozhevnikov et al.,* 2020] draws attention to the following. 1) There are two transverse gorges about 150 m deep near the coastal ridge of the mountains of the considered area in its northwestern and southeastern parts and opposite the lateral edges of the bay. Here, the average heights of the mountains in the northwestern part are almost 100 m lower than in southeast. 2) The lateral edges of the bay are framed by peninsulas with heights of about 150 m and a length of 11 and 6 km towards the sea. This suggests that: a) the obtained estimates of the jet stream energy on the leeward slopes underestimate true values; b) spatial heterogeneity of air inflow from the mountains to the surface layer behind the mountains contributes to a particularly strong turbulation of movements in Novorossiysk compared to neighboring areas. Further studies should take into account the specified features of the relief.

4. Significant Results

It has been shown that the wind in event of bora acquires a catastrophic force primarily due to the fact that the leeward slope of the streamlined mountains has a noticeable steepness and the air flow here has a jet character. It has been shown: a) the velocities in the jet theoretically depend on the Lyra wave scale in the incident flow in front of the mountains and, therefore, primarily on its speed; b) when the incident flow speed does not exceed 15 m/s, the speed in the jet is 20 m/s in the first approximation and weakly depends on the properties of the flow. According to calculations, the air temperature entering the city Novorossiysk and coastal areas after passing the mountains is heated by no more than 1.5 degrees due to adiabatic compression. Consequently, a serious temperature decrease in event of bora is determined not by the flow effect, but by the arrival of a cooler air mass in the investigated region.

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