COASTAL SMALL-SCALE VARIABILITY OF THE SURFACE TURBULENT MOMENTUM FLUX

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1 INTRODUCTION

Efficient downscaling methods are constantly being sought as a substitute for high-resolution applications with consequent high computational cost. A great part of the venture can be achieved by the careful choice of the input and output variables for particular application. The problem here is that the simplifications, necessary to reduce the complexity of the system one wants to describe, can render the resulting theory useless for some conditions. For example in the case of atmosphere-ocean interaction, the subject of this paper, among the important assumptions are homogeneity of the surface roughness, and a straight coastline.

The main motivation for this paper is to find and explain the small-scale variations of the surface turbulent fluxes of momentum within a coastal zone. It is believed that a reasonably useful picture of the small-scale features can be obtained from the larger-scale conditions. An example of possible application would be in oceanographic models, where the stress of the wind to the sea surface generates currents and can promote upwelling. Another example is the input of nutrients from the atmosphere, where the dry deposition is governed in part by the atmospheric turbulence. In this paper a diagnostic method is proposed, which when applied on known large-scale conditions results in a high-resolution spatial distribution of the surface turbulent momentum fluxes. The diagnostic is constructed using a high-resolution numerical model, which by its high resolution is able to explicitly describe the small-scale variability of the marine ABL near the coast. Traditionally, the parameters of a certain scheme are extracted from the results of an experimental campaign, set up for the particular problem. However, small-scale coast-induced features expressed in the form of turbulent stress to the water surface are sensitive to many large scale variables and their combinations and exhibits a high spatial variability. A huge and very expensive field experiment would be needed in order to obtain a sufficient amount of real data from a wide enough range of conditions for an accurate description of those features under all possible conditions. An efficient and less costly method is to employ a numerical model. This also allow control of the background conditions, makes it possible to exclude unwanted and irrelevant influences and to widen the range of conditions for the study. Thus, in this research a numerical model is used as a proxy for the real atmosphere. We firmly believe that the benefits of this approach overweight the negative considerations of not using real data for determining the diagnostic parameters. Of course, the output of the diagnostics can then only be considered valid down to the resolution of the model used for constructing the method.

2 THE NUMERICAL MODEL

The meso-scale meteorological model, developed at the Meteorological Institute, Uppsala University (MIUU) used in the present study is a hydrostatic, primitive equations model, using an Eulerian advection scheme in grid-point space. A terrain-following, η coordinate is used for vertical discretization and the horizontal grid is stretched. The model has been successfully used for various applications by e.g. Enger (1990), Brooks et al. (2001).

In the model the turbulent exchange coefficient K_M is computed through a local, 1.5-th order turbulent closure. This means that an additional prognostic equation for the turbulent kinetic energy ($\mathrm{TKE} = 0.5(\overline{u'^2}+\overline{v'^2}+\overline{w'^2})$) is introduced, while the remaining second-order moments can be calculated from steady-state equations. This technique enables a good estimation of the K_M, which is a diagnostic function of the TKE, stability, wind shear and a turbulent length scale.

The simulations, which serve as the reference frame for this study, were performed with a two-dimensional configuration of the MIUU model. The full width of the domain was 320 km and the center was situated exactly at the coastline, $x = 0$, with sea to the west and land to the east. Horizontal grid spacing, Δx, was 4 km at the coast and 16 km at the border of model domain. Tests with higher resolution showed no significant change in the simulated horizontal variations in surface stress and other variables close to the coastline. Surface roughness (z_0) was set to 1 mm over sea and 10 cm over land. 30 levels were distributed between ground and 5000 meters, the first few as following: z_0, 1, 3, 7, 15m, etc., and the top two levels at 4725 and 5000m. Specific domain features, limiting the generalizability of our study, are straight the coastline, an infinite ocean at the western side and infinite, flat land at the eastern side of the coast. Sensitivity tests, evaluating the effects of relaxing each of these features, demand much wider investigation than the framework of this study.

We performed numerous simulations with the model for different background conditions. The parameters that are varied are: geostrophic wind velocity and direction, temperature contrast between land and sea, and static stability of the undisturbed background state. On-shore and off-shore wind situations are simulated while varying other parameters in all combinations: 5 wind velocities, 8 land-sea temperature contrasts and 7 stability classes. The differential heating of the surface is introduced by specifying a sinusoidal evolution of the land surface temperature, from equal surface temperature (14^{o}C) of land and sea at 6:00 local time and increase the land surface temperature to reach the maximum difference (ΔT_s) compared to the sea at 13:30 local time. Integrations start at midnight.

3 THE RESULTS

Using the model database we tried to find relations that could be used for detailed description of the processes of interest. This was done for each factor separately, and for the combined influence of all factors.

3.1 Influence of land-sea temperature contrast

The air approaching the coastline from the ocean starts experiencing the effects of the coastline well before it reaches it. The surface stress for various ΔT_s and its horizontal variations are best approximated by the following relation:

$$\begin{aligned} \overline{u'w'_0} &= \overline{u'w'_{NC}}\,\mathrm{SF}_{\Delta T_s} \cdot \\ &\cdot \left(1 - B\sqrt{\frac{\Delta T_s}{T_{ss}}}\left(1 - \sqrt{\frac{x}{x_0}}\right)\right), \qquad (1) \end{aligned}$$

where $\overline{u'w'_{NC}}$ is the surface stress without the presence of the coast, T_{ss} the sea surface temperature and x_0 the distance from the coast (x) where the effects of the coast are not observed any more. The constant B is empirically defined so that the curves for different temperature contrasts come as close as possible to each other. We follow Garratt's (1987) reasoning of an inverse square-root dependence of the internal boundary layer (IBL) height on the temperature difference between land and sea, and square-root dependence of the same parameter on the distance, x, from the coast. $\mathrm{SF}_{\Delta T_s}$ is the similarity function, which in fact includes the remaining variability, not explained by the scaling, based on the temperature contrast. It mostly accounts for the sole presence of the coast, which even without any temperature contrast introduces variations in the surface turbulent flux of momentum. The scaling does not work in areas where local thermal circulations develop.

If the off-shore wind is strong enough to prevent the sea breeze from occurring, the surface fluxes at sea, scaled by the square-root of the temperature contrast, exhibit obvious similarity and the scaling relation is:

$$\begin{aligned} \overline{u'w'_0} &= \overline{u'w'_{NC}}\,\mathrm{SF}_{\Delta T_s} \cdot \\ &\cdot \left(1 - B\sqrt{\frac{\Delta T_s}{\Delta T_{NF}}}\left(1 - \sqrt{\frac{x}{x_0}}\right)\right), \qquad (2) \end{aligned}$$

The new parameter ΔT_{NF} is the value of the temperature contrast for which, together with the parameter B, the surface turbulent momentum flux at the sea immediately after the coastline approaches 0. For the investigated cases 15 K is the value of ΔT_{NF}, which results in best scaling. Such a formulation (2) is equivalent to (1). However the two parameters B and ΔT_{NF} are obviously not independent; scaling relation for off-shore wind can be also written with only one parameter, but such a choice would suffer from aesthetic point of view. For off-shore wind conditions we did not find it necessary to introduce any dependence of x_0 on stability. In fact, a slightly more natural parameter in off-shore wind conditions would be the travel time of the air parcel above sea surface, compared to some time scale τ, dependent on the wind speed, stability and temperature contrast. It is immediately seen that some parameters would enter here in a nonlinear fashion, which is not suitable for this simplified diagnostic method. Therefore we keep x_0 as a scaling parameter also in the off-shore cases.

It is worth noticing that the amount of unexplained variability is smaller in the case of off-shore wind, than in that of on-shore wind. This indicates that the magnitude of the temperature contrast at the coast is the main reason for the variations in the surface stress in off-shore situations, whereas in on-shore situations the important feature is the sole fact that the boundary layer over land is well mixed.

3.2 Influence of the stability

In this study, we are considering the background stability ranging from the near neutral stratification ($\partial\Theta/\partial z = 1$ K km^{-1}) until the isothermal state ($\partial\Theta/\partial z = 10$ K km^{-1}). Using the Rossby radius of deformation as the reference, we can expect that the distance from the coast where the effects are already felt is larger if the air-mass is stable. This is confirmed by numerical simulations.

In the very stable atmosphere and with stronger winds, the fluxes near the coast are reduced by 10 %, comparing to the open sea values. Elsewhere, the differences are in general smaller. If the land is warmer, two modes are apparent. Under on-shore wind regime, the values of the fluxes mainly depend on the stability while under off-shore wind regime they depend mainly on the distance from the coast, owing to the fact that the ABL gradually adjusts to the new lower boundary.

By scaling the surface turbulent momentum flu-xes by the background atmospheric stability most of the influence of this parameter can be explained, at least in this study where only the stable boundary layer is concerned. The following expression is found to provide the best scaling properties for all on-shore cases, and for off-shore cases with weak to moderate wind:

$$\overline{u'w'_0} = \overline{u'w'_{NC}}(U, \gamma_i)\mathrm{SF}_N \left(\frac{\partial\Theta/\partial z}{\gamma_i}\right)^{-\alpha}, \quad (3)$$

$$\alpha = \begin{cases} 1/3 & ; |\vec{u}| > 10\mathrm{ms}^{-1} \\ 1/2 & ; |\vec{u}| \leq 10\mathrm{ms}^{-1} \end{cases}$$

where γ_i is the vertical gradient of the potential temperature in an isothermal atmosphere ($\gamma_i \sim 10$ K km^{-1}). The parameter $\overline{u'w'_{NC}}(U, \gamma_i)$ is the value of the surface stress at the open sea in conditions with wind speed U and atmospheric background stability γ_i.

3.3 Influence of the wind speed

The power of the relation between the geostrophic wind speed and the surface turbulent momentum flux was also investigated. Namely, it is found that the exponent in the power law varies for different values of the atmospheric background stability. The more general relation can be written as:

$$\overline{u'w'_0} = \mathrm{SF_U} \left(\frac{U}{U_R}\right)^{\beta}. \quad (4)$$

Here, the similarity function $\mathrm{SF_U}$ is the value of the surface momentum flux for the reference geostrophic wind speed U_R (10 m s^{-1} in our case), and can be computed separately; for example the undisturbed, open-sea value ($\overline{u'w'_{NC}}(U_R)$) can be taken. The exponent β is an empirically defined function of the atmospheric stability and the geostrophic wind speed and its value is somewhat less than 2:

$$\beta = 2 - \mu \left(1 - \frac{\partial\Theta/\partial z}{\gamma_i}\right) U, \quad (5)$$

where U is the geostrophic wind speed and γ_i is again the isothermal vertical gradient of potential temperature. Parameter μ has an empirically defi-ned value of 0.03, which provides the best fit to the model data used. Equation (5) is saying that the relation between the background, geostrophic wind speed and the surface turbulent momentum flux is quadratic in the isothermal atmosphere but is reduced to around power of 1.4 in the neutral atmosphere and strong geostrophic wind. This does not mean that for certain surface wind speed the turbulent fluxes depend on stability as shown. The turbulent fluxes in stably stratified atmosphere, due to the shear generation of turbulence, is a function of the vertical wind shear and the turbulent exchange coefficient. The wind shear depends on, among other, the atmospheric stability and the geostrophic wind speed, and

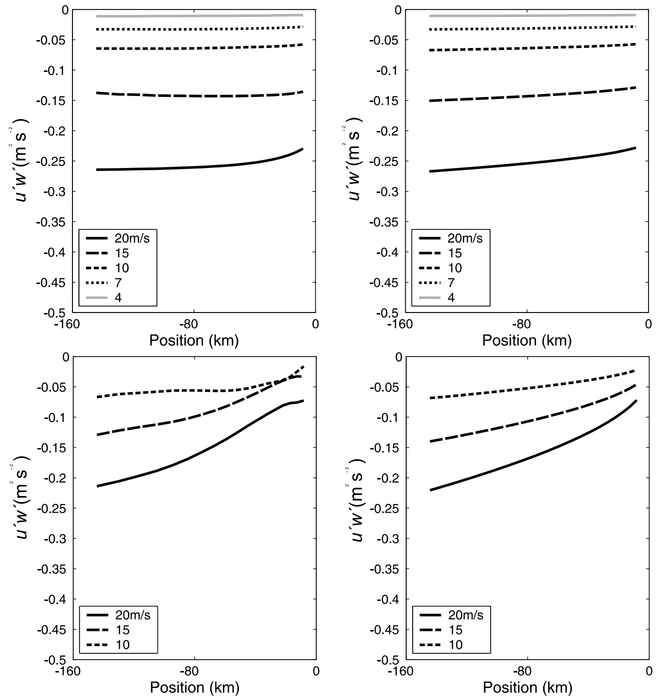

Figure 1: Surface turbulent fluxes as predicted by the MIUU model (left) and computed from the geostrophic wind speed, background stability and the temperature contrast (right). Above - on-shore wind, $\partial\Theta/\partial z = 9.9$ K km^{-1}, and $\Delta T_s = 4$ K. Below - off-shore wind, $\partial\Theta/\partial z = 4.8$ K km^{-1}, and $\Delta T_s = 12$ K.

the exchange coefficient is a complicated function of the TKE, which is again depending, among other, on stability and the geostrophic wind speed. Eqs. (4) and (5) only express the most statistically significant dependence of the surface turbulent momentum flux at the sea near the coast on geostrophic wind speed.

3.4 Generalization

From the relations in previous subsections, we can construct a combined parameterization of the small-scale coastal effects on the surface turbulent momentum fluxes based on the prescribed large scale parameters. Applied on a set of geostrophic wind speed and direction, temperature contrast between the land and the sea, and the background atmospheric stability, it results in a distribution of small scale fluxes, depending on the distance from the coast. The results, compared to the fluxes obtained from the high resolution model, are presented in Fig. 1. We can identify a certain level of agreement between the curves representing the modeled and the diagnosed values.

4 CONCLUSIONS

We constructed a diagnostic method for downscaling the influences of the coast with its properties, to the turbulent flux of the momentum to the sea surface. In short, the method consists of: taking the large scale values obtained either from the measurements or from a large-scale forecast model, applying the appropriate diagnostics obtained on the basis of a higher-order turbulent closure high-resolution model, and assuming that the results represent the small-scale variable distribution of surface fluxes. The results are mostly applicable in situations where a surface picture of the variable in interest is sought but only one or few observations are available. We have had in mind the estimation of dry deposition to the coastal waters for the purposes of environmental engineering, or the estimation of the surface wind stress to the sea for oceanographic studies.

Generally it is found that the surface momentum fluxes above the sea, near the coast, are almost always smaller than at open sea for the same large-scale atmospheric conditions. The only exceptions are the sea breeze and an off-shore wind from the land, which is colder than the sea (the nocturnal land breeze or its equivalent). This second example is additionally different because a convective boundary layer develops above the sea.

Assumptions we have used in constructing the diagnostic procedure in this study make the results applicable only under conditions contained in the assumptions. The main limiting factors are the shape of the coastline, which is rarely straight as in this study, and homogeneity of the surface of the ocean and the land side of the coastline.

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