The response of bora-type flow to sea surface temperature

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SUMMARY

A non-linear, two-dimensional, hydrostatic, incompressible numerical model with a higher-order turbulence closure scheme is used to study the effect of sea surface temperature on the severe downslope wind called bora at the Adriatic coast. A non-linear large-amplitude mountain wave is generated and is broken beneath and within its critical layer, due to resonant tuning between the initially single-layer atmosphere and the terrain. The tuning is governed by the Froude number. A qualitative and sometimes quantitative analogy exists between the wave-breaking (unsteady, stratified) flow and the hydraulic jump (steady, two-layer flow). It is also known that the strongest Adriatic bora appears during the winter season, when the sea surface temperature is typically larger than the ground surface temperature.

Firstly, a relatively higher (lower) sea surface temperature means an additional distortion (moderation) of the mountain wave and consequently a larger (smaller) area with bora wind maxima. For a relatively higher sea surface temperature a propagating hydraulic jump occurs. Typically bora maxima are about three to four times larger than the related geostrophic wind (8 m s\(^{-1}\)). Secondly, the presence and importance of the inertial oscillation are indicated. Since the wave-breaking is the vital component of the strongest bora cases, there is a relatively large, elevated area—i.e. the critical layer—with substantial flow decelerations and generally low wind speeds. The wave-breaking area has a Rossby number \(\sim O(1)\). Hence, the earth’s rotation appears to be an important part of bora evolution. The simulations presented consider generalized bora cases which may pertain to other similar orographic flows.

KEYWORDS: Downslope wind  Mountain wave  Numerical model

1. INTRODUCTION

(a) Observations and theory

The so-called bora is a gusty, downslope wind which appears at the eastern Adriatic coast. Strong bora develops in certain situations with strong synoptic pressure gradients (e.g. Poje 1992). The hourly average surface wind speed exceeds 17 m s\(^{-1}\) and gusts reach values over 50 m s\(^{-1}\) during situations with a very strong bora. Dynamically similar winds appear at other places, e.g. at the eastern side of the Rocky Mountains. The literature reveals a few dozen observational studies dealing with bora, some of which are described in e.g. Lukšić (1975), Jurčec (1981), Smith (1987), and Tutiš and Ivančan-Picek (1991). Makjanić (1970) addresses the occurrence of diurnal variations of the bora wind speed; Smith (1979, 1985) reveals the bora dynamical nature to be similar to the Boulder (Rocky Mountains) downslope wind storm. Moreover, Smith (1985, 1987) and Smith and Sun (1987) modify the hydraulic theory and were the first to successfully relate bora to this theory (Long 1954). It appears that the hydrostatic assumption captures the essentials of the bora-type flows, although moderate non-hydrostatic effects may occur at the lee side (Smith 1991; Scinocca and Peltier 1993). Since linear wave theory does not hold for bora, Bajić (1991), Vučetić (1993), and others have followed Smith’s hydraulic approach. However, there is also evidence that the hydraulic theory cannot explain all situations with bora, e.g., some bora cases appearing in the southern Adriatic coast cannot be resolved as a hydraulic flow (Ivančan-Picek and Tutiš 1996). Besides three-dimensional (3-D) effects, the real bora exhibits unsteady, sheared, stratification-varying effects, which are not accommodated by the hydraulic theory†.

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† Smith and Sun (1987) extended the theory to nonlinear steady-state solutions for stratified two-layer flows; however, wind shear was not accommodated.
Froude number, $Fr$, is the principal parameter describing the nonlinear airflow over terrain (Long 1954; Miles and Huppert 1969). Its background value, $Fr_0$, is defined as

$$Fr_0 = U_0/(h_0N_0)$$

(1)

where $U_0$, $N_0$ and $h_0$ are the unperturbed background wind speed perpendicular to the obstacle, buoyancy frequency, and the maximum obstacle height, respectively (all values positive). Its 3-D consideration is given in e.g. Miranda and James (1992), Castro and Snyder (1993) and Reisner and Smolarkiewicz (1994), while its profound meaning for the nonlinear two-dimensional (2-D) flow is revealed in e.g. Durran (1986) and Scinocca and Peltier (1993). Grisogono (1995; henceforth G95) uses the Froude number to distinguish linear and non-linear flow regimes, later exhibiting wave-breaking, and their subsequent effects on the atmospheric boundary layer (ABL). In some respects, the present study reverses the problem by analyzing the effect of an already horizontally-inhomogeneous ABL on the mountain wave-breaking. It will be shown that the state of the ABL significantly suppresses or enhances the area occupied by wave-breaking and hydraulic jump. These findings are in accordance with Reisner and Smolarkiewicz (1994) who show, for the $0 < Fr < 1$ regime, that thermal effects are substantially more pronounced on the lee than over the windward side of the mountain.

The only successful analytical approach explaining a considerable part of bora realizations is the hydraulic theory; it relates here to shallow water dynamics. If the far-upstream bora flow is presumably subcritical, $Fr < 1$, then around the top of the ridge $Fr \to 1$; furthermore, on the lee side the flow becomes supercritical, $Fr > 1$, reaching the maximum speed after being continually accelerated. Finally, the flow returns through a hydraulic jump back to the subcritical regime (e.g. Holton 1992). The bora dynamics depend on the state of the lee-side environment, which contains a substantial distortion of the mountain gravity wave. If the supercriticality for bora parcels is short and very localized, the flow recovery via the hydraulic jump moves the parcel up before reaching the end of the lee side. In contrast, if the parcels are allowed to travel all the way down the lee side and offshore—more thorough potential energy conversion into kinetic energy—the bora front will propagate further out over the sea as the hydraulic jump is postponed.

One way of keeping $Fr > 1$ as the parcel approaches the bottom of the lee side is by lowering the local buoyancy frequency. Bora will then presumably be intensified, but the details are not well-known. This hints at the main idea of this study.

(b) Numerical simulations

Numerical simulations are mainly concerned with the models’ ability to reproduce the bora-type flows (Hoinka 1985, Bougeault and Lacarrère 1989, G95), or to assess and interpret the dynamic essence of the bora flows (Clark and Peltier 1977, 1984; Durran 1986; Klemp and Durran 1987). The latter explain that the bora shooting flow and its mountain wave-breaking aloft resemble the transition to supercritical flow and the eventual hydraulic jump. There are more processes involved in the mountain wave dynamics and bora flows than can be explained by the simple hydraulic theory. However, this theory still serves as a very good conceptual model. Klemp and Durran (1987) show with their numerical simulations of bora flows, that the essence of the dynamics is the wave overturning, not the inversion layer or the critical layer, i.e. the layer where wind speed equals the wave phase speed—both equal to zero here (following the definition of Drazin and Reid (1981), and extending it from the level to the layer). Shooting flow and the highest wind speeds on the lee side are produced by the overturning aloft. Their results, as well as some others, have been qualitatively repeated before starting this investigation. Bougeault and Lacarrère (1989)
employ a higher-order closure for turbulence parametrization when studying orographic airflows on meso-β scale (ours is on meso-γ). They indicate the importance of this type of parametrization because of: (1) improvement of the mean flow calculations as the waves are modified in turbulent regions, (2) advancing our theoretical knowledge about orographic flows, and (3) benefits for applied meteorology. Also within the higher-order closure, Richard et al. (1989) study the effect of surface friction on the Boulder downslope wind and reveal a moderate delay in the windstorm onset with increased friction. Trüb and Davies (1995) indicate our lack of knowledge about the ABL effects on nonlinear orographic waves in the f-plane. Those points add to the justification of the goal assigned and the tool used in this study.

The present working hypothesis is to control the sea surface temperature (SST), altering the coastal ABL buoyancy frequency, which in turn may modify the nonlinear mountain wave dynamics, wave overturning, and the offshore structure of the bora-type flow. One particular goal is to study the effect of the temperature difference (ΔST) between SST and land surface temperature on the maximum horizontal extension of the bora wind. For this purpose a meso-scale numerical modelling procedure is adopted. The conditions assigned in the present study are idealized situations, but still based on simplified real ones. In a larger context, the bora’s intensity, duration and internal structure are not well-predicted today. Besides insufficient resolution in operational forecasting models, there is also a lack of data over the Adriatic sea which exhibits a complicated response to bora and sirocco forcing (Orlić et al. 1994). Moreover, our theoretical understanding of stratified nonlinear flows is still under development. Needless to say, bora greatly affects local traffic, tourism and air pollution distributions. Also, the scientific emphasis is on severe bora cases which occur more often in winter (e.g. Yoshino 1976; Poje 1995).

2. THE NUMERICAL MODEL

(a) Overview

A nonlinear, two-dimensional, hydrostatic, incompressible numerical model with a higher-order turbulence closure scheme is employed. The model, sometimes called the MIUU model, has been well-documented in the literature (Tjernström 1988; Tjernström et al. 1988; Enger 1990; Enger et al. 1993), tested against measurements (e.g. Svensson 1996; Enger et al. 1993) and against analytical solutions (Yang 1993; G95). In particular, the model has been employed for studies of orographic and coastal flows; these studies include theoretical analyses (Yang 1993; G95; Holmgren 1995; Grisogono and Tjernström 1996) and actual 3-D studies (Enger et al. 1993; Svensson 1996; Tjernström and Grisogono 1996). It solves the prognostic equations for the horizontal wind components (U and V), potential temperature (Θ), specific humidity and turbulent kinetic energy (TKE); the other meteorological parameters involved are solved diagnostically every time-step. The full description of the model is not repeated here, but the main 3-D equations are listed in appendix A. The best characteristics of the model are nonlinearity and the treatment of turbulence. The latter involves a “level 2.5” closure (Mellor and Yamada 1974) which is described and improved by Andrén (1990). Since the Adriatic bora is a relatively shallow flow, it is believed that compressibility effects are of minor importance. In the present study a new advection scheme, which is of third order both in time and space, has been applied, see appendix B.

(b) The model setup and initialization

The total model domain is 460 × 10 km² resolved with 141 × 61 main grid points with the finest resolution in the centre near the surface, 2 km × 4 m, and the coarsest resolution
at the upper lateral boundaries, 5.5 km × 180 m. Monotonic decreases in resolution away from the area of interest attenuate spurious reflections from the model boundaries. A sponge layer occupies the uppermost 4 km. The time-step is 10 s. Dynamic initialization (e.g. Pielke 1984; Enger 1990) with one hour of one-dimensional (1-D) and 24 hours of 2-D pre-integration introduces 50 hours of total simulation time. In order to distinguish a more 'pure' bora evolution unrelated to the diurnal cycle, the surface temperature is kept constant in time—lowering the set of time-scales involved. The present study deals with bora wind development over a south–north ridge with a sea west of the ridge. The idealized terrain used qualitatively corresponds to the coastal mountain range at the eastern Adriatic coast. This is a half Gaussian-like ridge with $h_0 = 800$ m and half-width equal to 10.1 km that is steeply$^*$ cut off on its lee side. Sea resides immediately west of the asymmetric ridge, $x$ less than −8 km in the Figures. A so-called control run and several sensitivity tests are performed.

In the control run an easterly geostrophic wind, constant both in space and time, with $U_g$ of 8 m s$^{-1}$ and a background stratification of 5 K (km)$^{-1}$, are used. This yields a background vertical wave length

$$\lambda_z = 2\pi U_g / N_0$$

which is equal to 3.75 km. The zero-level air temperature is 285 K. Initially, a low specific humidity, monotonically decreasing with height from its surface values of $4 \times 10^{-3}$ kg (kg)$^{-1}$, is used. The 2-D ridge is situated at latitude 45°N. Hence, the runs loosely relate to the wintertime, anti-cyclonic clear-sky bora (e.g. Jurčec 1981; Poje 1995). The surface roughness, $z_0$, is set to 0.01 m for all land and to 0.00025 m for sea. This yields a background $Fr = 0.747$, as in G95, which is about 4% larger $Fr$ value than in Durran (1986); see his Figs. 15 to 18.

Types of sensitivity tests—differences from the control run:

- Different geostrophic wind, one case from the south–east 11.3 m s$^{-1}$, and another case from the north–east 11.3 m s$^{-1}$, giving a geostrophic wind component perpendicular to the ridge of the same magnitude as in the control run, i.e. 8 m s$^{-1}$.
- The SST higher than the land surface temperature, i.e. $\Delta ST = 0.5, 1.5, 2.5, 5, 7.5$, and 10 K.
- The SST lower than the land surface temperature, i.e. $\Delta ST = -5$ and $-10$ K.
- Two other values of $Fr$, around 0.5 and 1 (varying $U_g$ and $N_0$, independently).
- Roughness length over the land equal to 0.1 m and 1.0 m.
- Different resolution of the vertical grid, i.e. coarser vertical grid resolution by using 31, 46 and 53 grid points, respectively, and finer vertical grid resolution by using 71 grid points.
- Several variations of latitude, thus providing different $f$-values.
- Somewhat different shape of the lee side of the ridge.

These sensitivity runs will be discussed at various lengths below.

3. RESULTS AND DISCUSSION

(a) The control run

A few characteristic results from the control run will be shown first. They are in overall agreement with those simulations mentioned in subsection 1(a), although they depart in details due to variations among the model assumptions, initial and boundary conditions.

$^*$ The steepness never exceeds the slope of 0.1 required by the hydrostatic approximation.
Figure 1 shows (a) the potential temperature, $\Theta$, (b) the wind speed, (c) the $U$-component, (d) the $V$-component, (e) the vertical velocity and (f) the TKE after 30 hours of simulation following the pre-integration. A nonlinear large-amplitude mountain wave is generated beneath, and broken within, its critical layer. The dynamics of the layer and the wave are inter-related. The induced critical layer starts roughly at $\lambda_c/2 \approx 1.8$ km above surface—the $U$-component (Fig. 1(c)) within this layer is changing sign and the magnitude of the total wind speed (Fig. 1(b)) is less than 2 m s$^{-1}$. Such an association between Eq. (2) and the height of the wave-breaking region is also found in other studies (e.g. G95). The wave-breaking within the critical layer appears in the area with the steepest isentropes, giving near neutral or slightly unstable stratification (Fig. 1(a)).

The generated wave causes the isentropes near the surface to merge closer to each other. This is most pronounced on the lee side of the ridge (Fig. 1(a)). The wind speed at the lee side of the mountain increases by roughly a factor proportional to the ratio between the isentropes' separation in the undisturbed upstream area and the isentropes' separation over the lee side of the ridge. In this control run the maximum wind speed is around three times the geostrophic wind speed. The consequent wind decrease is due to the hydraulic jump which also causes divergence and produces large positive vertical velocities (Fig. 1(g)). Note that the isentropes become steeper at the downslope edge of the wave, where the hydraulic jump occurs. The highest wind speed occurs as shooting flow with low turbulence (Figs. 1(b) and 1(f)). In contrast, the highest TKE is obtained in the area where the stratification is neutral or unstable and where the largest shear occurs. This area is situated just above the high-speed, stable layer on the lee side. The highest TKE is situated approximately 1.3 km above the surface.

The lee-side TKE within and above the critical layer shows a quasi-periodic variation (Fig. 2). The period of 15 to 16 h nearly corresponds to the inertial oscillation period for this latitude (45°N). The oscillation is not present during the first few hours of the 2-D initialization period (first five hours in Fig. 2), but starts when the the wave breaking sets in. All other runs also show this quasi-periodicity which begins only after wave breaking is activated. This means that it is not an artifact of the initialization. No runs show a longer significant period than the inertial period. However, somewhat shorter periods may occur, as is the case in Fig. 2, because unsteady nonlinear processes may release TKE in the breaking layer earlier. At higher elevations, where wave breaking and turbulence are less vigorous, the TKE shows, more systematically, a period that is consistent with the inertial period. Executing the model at other latitudes confirms these findings. In the lowest layer beneath the breaking layer all the way down to surface, no periodicity is found. Turbulence there is almost constant with time.

Summary of the control run:

- A wave steepening, overturning, and breaking occur at the lee side of the ridge within a critical layer that starts at a level equal to $\lambda_c/2$ from the surface.
- The shooting flow close to the surface is around three times faster than the geostrophic wind and concentrated at the lee-side slope; TKE there shows its minimum at $\lambda_c/8$.
- The TKE is largest above the shooting layer with values of O (10 m$^2$ s$^{-2}$), roughly at a height above surface corresponding to $\lambda_c/3$.
- A quasi-periodic signal close to the inertial period is present after onset of the wave breaking.
- Large vertical velocities, of O (1 m s$^{-1}$), are produced at the downstream edge of the hydraulic jump.
Figure 1. The control run fields ($\Delta S T = 0$) after 30 hours of simulations: (a) potential temperature $\Theta$ (K), (b) the total wind speed (m s$^{-1}$), (c) the across-shore wind component $U$ (m s$^{-1}$), (d) the along-shore wind component $V$ (m s$^{-1}$), (e) vertical velocity $W$ (cm s$^{-1}$), and (f) the turbulent kinetic energy, TKE (m$^2$ s$^{-2}$).
Figure 2. A lee side time-height cross-section of the TKE for the control run (2-D initialization period included). A few layers occur and these are associated with the near-surface maximum, the shooting flow (low-level TKE minimum), the wave-breaking layer (quasi-periodic TKE maximum), and the ‘wavy region’ above the main breaking layer.

(b) Sensitivity tests

As mentioned in sub-section 2(b), 61 grid points are used in the vertical direction. For this model setup and flow regime, tests show that at least a total of 53 vertical grid points are needed to correctly resolve the dynamics of the breaking layer. Very small differences occur when using more than 53 grid points. To assure proper resolution in all runs 61 grid points are used.

Figure 3 shows (a) the potential temperature, $\Theta$, (b) the wind speed, (c) the $U$-component, (d) the $V$-component, (e) the vertical wind speed, and (f) the TKE for the geostrophic wind south-easterly and 11.3 m s$^{-1}$. This corresponds to a geostrophic wind component perpendicular to the ridge of the same magnitude as in the control run, i.e. $-8$ m s$^{-1}$, but also with a positive along-ridge component. In the calculation of the vertical wave length, Eq. (2), the involved wind speed is the background wind speed perpendicular to the ridge. By comparing Fig. 3 with Fig. 1 one notices that an orographic wave is produced at the same levels as in the control run. However, the wave recovery in the form of a hydraulic jump propagates further downstream compared to the control run. This effect is caused by the presence of the assigned synoptic lower pressure, as expressed by the south-easterly geostrophic wind, at the lee-side of the ridge. This pressure gradient perpendicular to the ridge favours the orographic wave propagation further downstream. Compared to the control run, the shooting flow expands further out over the sea, following the hydraulic jump. The bora front related to the hydraulic jump stretches about 20 km offshore and generates another vertically propagating gravity wave. This wave is clearly seen in the vertical velocity (Fig. 3(e)), and is roughly of the same wave length as the orographic wave. The $V$-component displays two distinct shear zones over the sea (Fig. 3(d)) where
Figure 3. Same as Fig. 1 but a positive $V_y$-component is added. Thus, $(U_x, V_y) = (-8, 8)$ m s$^{-1}$ giving a total geostrophic wind of 11.3 m s$^{-1}$. The imposed large-scale low pressure over the sea induces offshore propagation of the bora front.
the elevated zone, related to the propagated hydraulic jump, is more intense here than in the control run (compare with Fig. 1(d)). Due to the offshore-extended hydraulic jump, the TKE here occupies a much larger area (Fig. 3(f)) compared to the control run. Moreover, the maximum TKE is positioned further downstream. Consequently, the marine atmospheric boundary layer (MABL) and the whole lower troposphere show pronounced variabilities both in time and space.

If a north-easterly geostrophic wind is used instead, the synoptic pressure will be higher over the sea. This higher pressure causes the orographic wave to be horizontally narrower compared to the control run. Furthermore, the wave is positioned closer to the top of the ridge (not shown). The horizontal suppression of the orographic wave in this case is an effect which contrasts with the offshore wave extension in the south-easterly case.

Using SST higher than the land surface temperature will, as in the case with south-easterly wind, produce a lower pressure over the sea. However, this is a mesoscale phenomenon. The pressure is calculated as

$$\frac{\partial \Pi}{\partial \eta} = -\frac{g}{\Theta} \left( \frac{s - z_g}{s} \right),$$

which is integrated from the model top, using constant Exner function $\Pi$ at the top, downward to the surface. Recall that the background potential temperature gradient is $5 \, \text{K km}^{-1}$. Higher SST heats the MABL; thus, the larger $\Delta ST$ the deeper the near-neutral MABL. Figure 4 shows (a) the potential temperature, $\Theta$, (b) the wind speed, (c) the $U$-component, (d) the $V$-component, (e) the vertical wind speed, and (f) the TKE for $\Delta ST = 5 \, \text{K}$. The structure of the meteorological parameters are similar to the south-easterly case (Fig. 3). The main differences are: the bora front propagates somewhat further out over the sea and the MABL is more convective, giving higher TKE and less shear. Above the MABL, around 2 km, the shear in the $V$-component is comparable to that in the south-easterly case. However, the $V$-component here shows a northerly jet. The propagating hydraulic jump transfers the $U$-momentum into the other two directions which causes the second vertically propagating gravity wave (Figs. 3(e) and 4(e)) and the jet-type effect in the $V$-component. The shear in both cases, south-easterly and $\Delta ST = 5 \, \text{K}$ (Figs. 3(d) and 4(d), respectively), significantly changes the $V$-component; towards zero in the former case (‘anti-jet’), and towards large negative values in the latter case (a northerly jet). These $V$-extrema are similarly situated around 2 km above the sea in both flows. The wave breaking TKE occurs somewhat further downstream compared to the control run, but the TKE extrema are similar. However, in contrast to the control run, the periodicity of the TKE above the shooting layer is almost equal to the inertial period.

For the cases with $\Delta ST$ equal to 7.5 K and 10 K, the bora front is propagating still further offshore. With $\Delta ST = 10 \, \text{K}$, shown in Fig. 5, the bora front weakens, namely the maximum of $|\partial U/\partial x|$ decreases. Increasing $\Delta ST$ causes an increase of $Fr$ through a relative destabilization compared to the control run. This $Fr$ increase makes the hydraulic jump susceptible to offshore propagation as suggested in Houghton and Kasahara (1968; Fig. 3). Growth of the MABL is a function of time and $\Delta ST$; relatively small $\Delta ST$ requires a long time for the MABL to grow to the layer containing the hydraulic jump. Larger $\Delta ST$ yields earlier offshore propagation of the hydraulic jump. Very small $\Delta ST$ requires an unrealistically long time to increase the local $Fr$ sufficiently to allow the propagation of the hydraulic jump. This modification of the flow for $\Delta ST = 10 \, \text{K}$, as shown in Fig. 5, depends on the state of the MABL—its $\Theta$ and TKE fields in particular. The hydraulic jump becomes smoothed out over the sea because of the intensified, convective MABL.
Figure 4. Same as Fig. 1 but the temperature difference ($\Delta ST$) between SST and zero-height land surface temperature is $\Delta ST = 5$ K. Note the effects similar to those with imposed large-scale low pressure over the sea (Fig. 3).
Figure 5. Same as Fig. 1 but $\Delta ST = 10$ K.
The weakening of the bora front implies a degradation of the vertically propagating gravity wave which is associated with the hydraulic jump (e.g. compare Figs. 3(e), 4(e) and 5(e)).

For negative $\Delta ST$, that is the SST colder than the land surface temperature, the orographic wave narrows and is positioned somewhat closer to the top of the ridge (not shown here), as in the case with north-easterly geostrophic wind.

Figure 6 summarizes some of the main findings of the present study with regard to the $\Delta ST$ effect. The bora front propagation is plotted as function of $\Delta ST$. For negative $\Delta ST$, the off-shore propagation never occurs (dotted). For small positive $\Delta ST$ (<3 K here), two states are possible depending on the time from the onset of the hydraulic jump. First, the smaller $\Delta ST$ the longer the time it takes to initiate the propagation of the hydraulic jump; the non-propagating state is shown dotted. Second, once this propagation is initiated it swiftly approaches the propagating state shown in Fig. 6 (solid and dashed lines). Large $\Delta ST$ (>3 K here) implies an almost immediate development of a propagating hydraulic jump. The sharpness of the bora front corresponds to the agreement between solid and dashed lines, the speeds of $1.2U_g$ and $2U_g$, respectively, corresponding to two different thresholds for the bora (by using this definition a measure of the sharpness of the bora front is addressed). For small positive $\Delta ST < 3$, having two possible states, it takes one to three days to evolve into the propagating state.

All runs show quasi-periodic behaviour of the TKE within and above the breaking layer; the periodicity is between 15 and 17 hours as in the control run (Fig. 2). The bora wind speed maxima are hardly influenced by $\Delta ST$, but the area occupied by this shooting flow differs in accordance with the position of the bora front. The control run is executed at various latitudes, a test somewhat similar to those in Arritt (1989) and Xian and Pielke (1991) for sea breezes. All runs, ranging from 30°N to 75°N (thus, $U_g$ from $-11.3$ m s$^{-1}$ to $-5.9$ m s$^{-1}$, and background $Fr$ from 1.06 to 0.55, respectively) exhibit bora regimes (wave-breaking and shooting flow beneath) at some stage of the flow evolution, although
varying in subtleties. That is in overall agreement with linear expectations, with worldwide findings of bora-type flows, and other nonlinear simulations (e.g. Smith 1979; Jurcée 1981; Ólafsson and Bougeault 1996). Trüb and Davies (1995) analyze pathways to regime transitions for flows over a mesoscale ridge; they mark differences in nonlinear wave dynamics depending on the inverse \( Fr \) and the Rossby number. Their study also suggests the importance of geostrophic imbalance on the evolution of orographic flows. Moreover, the decrease of \( U_g \) does not only lower the related \( Fr \) but also \( \lambda_z \). As the decreased \( Fr \) value here approaches the onset of a blocked flow (e.g. Miranda and James 1992; Trüb and Davies 1995), and there is an increased level of the upslope turbulence, the orographic waves become relatively more immersed in the ABL. This yields sufficient variations in the orographic flow to exhibit subtleties within bora regimes. Moreover, Ólafsson and Bougeault (1996) show that a very low \( Fr \) regime requires a 3-D treatment.

The nonlinear nature of the bora flows does not allow us to ignore the Coriolis effect, as one might, based on a scale analysis of the background flow characteristics (i.e. from a relatively small half-width of the ridge and relatively large background Rossby number). Mountain wave-breaking causes substantial flow decelerations and wind speed reductions so that the actual Rossby number approaches unity for layers of considerable thicknesses; thus, rotation eventually becomes important in this case. Furthermore, significant ageostrophic winds once induced, continue to play an important role in turbulent interactions and coastal jet formations.

In accordance with Richard et al. (1989), a radical uniform increase in \( z_0 \) (from 0.01 m to 1 m) did not produce drastic changes in the bora flow evolution. The wind speed maxima decreases slightly (1–3 m s\(^{-1}\)), shifts upward (for 0–300 m), and is delayed somewhat compared to the control run. Thus, it is concluded that even substantial variations in \( z_0 \) may not set back the evolution of bora-type flows. Finally, moderate variations of the lee-side slope, which always remains steep, did not give any significant departures from the control run (also see in Miller and Durrant 1991). At this point it is difficult to say which of the results presented here will have to be modified once the 3-D, non-hydrostatic simulations with time-varying surface forcing are considered.

A question about the bora classification as a local wind arises from Fig. 7. The wind profiles from \( \Delta ST = 7.5 \) K are shown at 100 km (Figs. 7(a) and 7(b)) and at 180 km offshore (Figs. 7(c) and 7(d)). Both the wind speed and the direction are far from being geostrophic, or even nearly monotonic with height at far-offshore distance. This means that soundings at the Italian coast may be strongly influenced by bora at the Croatian coast, especially for the upper \( \lambda_z/2 \). The lower parts of the model profiles correspond to a well-mixed MABL, up to 1 and 1.9 km high at 100 and 180 km offshore, respectively. This convective MABL is capped by a few distinct stable shallow layers containing the information about the wave breaking and the bora front. These shallow elevated layers and their structure can be seen in previous figures with cross-sections, e.g. Fig. 4.

Summary of the sensitivity tests:

- Appropriate spatial resolution is crucial for resolving the nonlinear dynamics considered here.
- With \( \Delta ST > 0 \), a propagating hydraulic jump occurs. This moves the bora front and the shooting flow out over the sea and gives a more vigorous bora flow. The offshore distance of the propagation is a function of \( \Delta ST \). However, for small positive \( \Delta ST \), the onset of the propagation is delayed.
- With \( \Delta ST < 0 \), the hydraulic jump narrows and is constrained toward the top of the ridge (no propagation).
Figure 7. The vertical wind profiles of the bora flow for $\Delta ST = 7.5$ K: (a) wind speed and (b) wind direction 100 km offshore; (c) wind speed and (d) wind direction 180 km offshore. Even at far offshore distances the airflow remains highly perturbed.

- Geostrophic wind with a positive along-shore component (from the south-east here) gives an effect similar to $\Delta ST > 0$.
- Geostrophic wind with a negative along-shore component (from the north-east here) gives an effect similar to $\Delta ST < 0$.

4. Conclusions

This numerical study addresses the bora-type flow response to SST forcing and geostrophic wind direction; the main independent parameters are the difference between the SST and the zero-level land surface temperature, namely $\Delta ST$, and the geostrophic wind direction. The meso-scale model employed is nonlinear, with a higher-order turbulence closure scheme, hydrostatic and with constant Coriolis parameter. It is used here as a 2-D prognostic tool with a fine spatio-temporal resolution.
It is known that the essence of bora-type flows is orographic wave-breaking related to $Fr \sim 1$ (e.g. Klemp and Durran 1987). A conceptual analogy with the wave-breaking is nonlinear hydraulic flow (e.g. Smith 1987). Relatively little is known about interactions between the ‘basic’ bora flows and the ABL, the Coriolis effect is commonly neglected, etc. This study attempts to answer some of the questions pertaining to those additional, ubiquitous interactions in the real atmosphere.

The main findings in this study are:

The offshore propagating hydraulic jump, associated with the bora front and the shooting flow behind, may be induced by a relatively lower pressure on the lee side. This lower pressure is due either to $\Delta ST > 0$ (a mesoscale phenomenon), or synoptically induced (geostrophic wind direction). The hydraulic jump remains on the lee side of the ridge (not propagating) for a relatively higher pressure on the lee side. This high pressure is caused by $\Delta ST < 0$, or by synoptic forcing. Hence, not only the synoptic situation, but also the state of the MABL is crucial for the bora development. For the given background conditions, the offshore propagation of the bora front is a function of $\Delta ST$ (see Fig. 6). Here it also becomes clear that the wintertime bora ($\Delta ST > 0$) must be more persistent and occupy a relatively larger area than its summertime counterpart. The presence and importance of the inertial oscillation near the wave-breaking layer is indicated. Since the wave-breaking is the vital component of the strongest bora cases, there is always an elevated, relatively broad, area with substantial flow decelerations and very low wind speeds (below and at the wave critical layer). This implies a relatively small local Rossby number regardless of its large background value dictated by $U_g$, $f$ and relatively small half-width of the ridge. It is shown that bora is not necessarily a local phenomenon. For example, the bora with a propagating hydraulic jump can cause a jet-type flow at over 100 km offshore. This preliminary study calls for further 3-D simulations and observational studies.

**Appendix A**

The governing equations

The 3-D model employs the following equation set in a terrain-influenced coordinate system (this study employs a 2-D version of the model):

\[
\frac{dU}{dt} = \left(\frac{s}{s - z_g}\right)^2 \frac{\partial}{\partial \eta} K_M \frac{\partial U}{\partial \eta} - \Theta \frac{\partial \Pi}{\partial x} - \Theta \frac{\eta - s}{s - z_g} \frac{\partial z_g}{\partial x} \frac{\partial \Pi}{\partial \eta} - fV_g + fV \tag{A.1}
\]

\[
\frac{dV}{dt} = \left(\frac{s}{s - z_g}\right)^2 \frac{\partial}{\partial \eta} K_M \frac{\partial V}{\partial \eta} - \Theta \frac{\partial \Pi}{\partial y} - \Theta \frac{\eta - s}{s - z_g} \frac{\partial z_g}{\partial y} \frac{\partial \Pi}{\partial \eta} + fU_g - fU \tag{A.2}
\]

\[
\frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} + \frac{\partial W}{\partial \eta} = \frac{1}{s - z_g} \left( U \frac{\partial z_g}{\partial x} + V \frac{\partial z_g}{\partial y} \right) \tag{A.3}
\]

\[
\frac{d\Theta}{dt} = \left(\frac{s}{s - z_g}\right)^2 \frac{\partial}{\partial \eta} K_H \frac{\partial \Theta}{\partial \eta} + \sigma_t \tag{A.4}
\]

\[
\frac{dR}{dt} = \left(\frac{s}{s - z_g}\right)^2 \frac{\partial}{\partial \eta} K_R \frac{\partial R}{\partial \eta} \tag{A.5}
\]
\[
\frac{dq^2}{dt} = \left( \frac{s}{s - z_g} \right)^2 \frac{\partial}{\partial \eta} \left( \frac{5}{\alpha_1 \rho^2} \right) \frac{\partial q^2}{\partial \eta} + 2 \left( \frac{s}{s - z_g} \right)^2 K_M \left[ \left( \frac{\partial U}{\partial \eta} \right)^2 + \left( \frac{\partial V}{\partial \eta} \right)^2 \right] \\
-2 \frac{s}{s - z_g} \beta g K_H \frac{\partial \Theta}{\partial \eta} - 2 \frac{q^3}{B_1 \lambda} 
\]

(A.6)

\[
\frac{d}{dt} = \frac{\partial}{\partial t} + U \frac{\partial}{\partial x} + V \frac{\partial}{\partial y} + W \frac{\partial}{\partial \eta} \\
\eta = \frac{z - z_g}{s - z_g}
\]

where \(U\) and \(V\) are the new quasi-horizontal wind components, \(W\) is the vertical wind component in the terrain-influenced coordinate system, \(\Theta\) is the potential temperature, \(f\) is the Coriolis parameter, \(g\) is the acceleration due to gravity, \(R\) is the mixing ratio, \(q^2\) is double the turbulent kinetic energy, \(z_g\) is the terrain height, \(s\) is the height of the model top, \(\lambda\) is the master length scale, \(\beta\) is the coefficient of thermal expansions, \(\alpha_1\) and \(B_1\) are closure constants, \(\Pi\) is the scaled pressure (Exner function), \(K_M, K_H,\) and \(K_R\) are functions of \(q^2, \lambda, \partial U/\partial \eta, \partial V/\partial \eta, \partial \Theta/\partial \eta\). For more information see e.g. Enger (1990) and Andrén (1990).

**APPENDIX B**

**Employed advection scheme**

The advection scheme used in the model is an upstream scheme that has been corrected for numerical diffusion. For a positive wind in the x-direction, the Taylor expansions around \(t + \Delta t, x + \Delta x, x - \Delta x, x - 2\Delta x\) are used. (The numerical diffusion is mostly caused by the second-order derivatives.) Assuming a constant wind speed, \(U\), and using the advection equation

\[
\frac{\partial f}{\partial t} = -U \frac{\partial f}{\partial x} \tag{B.1}
\]

the second- and third-order derivatives can be written as:

\[
\frac{\partial^2 f}{\partial t^2} = U^2 \frac{\partial^2 f}{\partial x^2}
\]

\[
\frac{\partial^3 f}{\partial t^3} = -U^3 \frac{\partial^3 f}{\partial x^3}
\]

These expressions are inserted into the Taylor expansion of Eq. (B.1). The second derivative, \(\partial^2 f/\partial x^2\), is calculated from the Taylor expansion around \(x + \Delta x, x - \Delta x, x - 2\Delta x\) giving:

\[
\frac{\partial^2 f}{\partial x^2} = a \left[ \frac{f(x + \Delta x) - 2f(x) + f(x - \Delta x)}{(\Delta x)^2} \right] + b \left[ \frac{f(x) - 2f(x - \Delta x) + f(x - 2\Delta x)}{(\Delta x)^2} \right] + \Delta x f_x^{(3)}
\]

where \(a + b = 1\). This expression is also inserted into the Taylor expansion of Eq. (B.1). By choosing \(b = (1 + C)/3\), where \(C\) is the Courant number, the scheme will be of third
order in both time and space for a constant wind speed. The new scheme reads:

\[
\begin{align*}
  f(t + \Delta t) &= f(t) - C(f(x) - f(x - \Delta x)) + 0.5(C^2 - C)[a(f(x + \Delta x) - 2f(x) + f(x - \Delta x)) + b(f(x) - 2f(x - \Delta x) + f(x - 2\Delta x))] \quad (U \geq 0) \\
  f(t + \Delta t) &= f(t) - C(f(x + \Delta x) - f(x)) + 0.5(C^2 + C)[a(f(x + \Delta x) - 2f(x) + f(x - \Delta x)) + b(f(x) - 2f(x + \Delta x) + f(x + 2\Delta x))] \quad (U < 0)
\end{align*}
\]

where \(a + b = 1\) and \(b = (1 + |C|)/3\).

The scheme is numerically stable for \(|C| \leq 1\). When the wind speed is not constant, \(\bar{U}\) built up by the nearest grid points is used: \(\bar{U} = (U(x + \Delta x) + U(x) + U(x - \Delta x))/3\).

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