Observations and simulations of a non-stationary coastal atmospheric boundary layer

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SUMMARY

This paper studies the turbulent and mesoscale structures of a coastal marine atmospheric boundary layer in the southern Baltic Sea. Data from a field experiment are examined for a case during autumn, and this case is subsequently modelled with a mesoscale numerical model.

During this day, the marine atmospheric boundary layer develops from a well-mixed cloud-capped state into a stably stratified, sheared boundary layer. This development is due to the dissipation of the cloud layer by radiation processes and by synoptic-scale subsidence. The boundary layer becomes non-stationary, and spectral analysis shows that the turbulence structure deviates from the ideal steady-state homogeneous boundary-layer turbulence, which forms the basis for most turbulence closures in numerical models. It is shown that the subsiding boundary-layer top and the increasing stability alter the spatial scale of the turbulent eddies; they become vertically distorted and smaller than those expected during steady-state conditions. In spite of these deviations, the turbulence statistics are sufficiently well behaved for simplified higher-order closure modelling to be applicable.

When the mixed layer collapses due to the dissipating clouds, the transition triggers an inertial oscillation and a non-classical low-level jet forms. At the same time, the changes in depth and stability of the marine boundary layer are perturbed by the mesoscale interaction between the flow and the coastal terrain. Thus, the upstream terrain increasingly blocks the flow, so that the inertial jet forms only for the part of the flow that overcomes the terrain obstacle. A numerical simulation of the case reveals the primary mechanisms, and several sensitivity simulations, wherein the terrain and the time-evolution of the forcing are manipulated, confirm the hypotheses proposed as causing the observed flow structure.

KEYWORDS: Internal oscillation Low-level jet Mesoscale structure Turbulence

1. INTRODUCTION

Atmospheric motions occur on a wide range of spatial and temporal scales, all of which may be important for a particular event. This often makes data from field experiments difficult to interpret. Different measurements must be made at many locations and at different time-scales in order to account for this wide-ranging variability. This is particularly true in the coastal zone. Here the atmospheric boundary layer responds to a step-change in surface conditions induced by the coast. As a result, a variety of mesoscale phenomena occur in the coastal zone (National Research Council 1992). These alter the local interaction of the atmosphere with the surface, as the mesoscale wind field becomes highly variable, while the mesoscale circulations themselves are sensitive both to the heterogeneous environment and to the interaction with the evolving synoptic-scale motion (Cui et al. 1998). Study of the meteorology in such a variable environment can be greatly augmented by analysing data from field experiments with the aid of a numerical model (Tjernström and Grisogono 1996, 1999; Grisogono and Tjernström 1996; Cui et al. 1998). If the model simulations can adequately re-create the meteorological situation that was encountered during the experiment, they could be deemed sufficiently reliable. Then the model data can be used to evaluate variables or parameters that were not, or could not be, measured. Complete agreement of the model results with the measurements may not be expected, but there must at least be a qualitative agreement, e.g. the observed and simulated spatial and/or temporal structures or flow types must agree. If this can be achieved, then the large-scale variability of the flow can either be removed, or manipulated in other ways, so that the sensitivity of the

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system can be investigated. Also, forcing from the surface can be specified in some detail, and can also be varied.

Since atmospheric motions are multi-scaled and evolving, the parametrizations in the model must be adequate in dealing with this aspect. For example, in the coastal zone, on the mesoscale \(O(1-10 \text{ km})\), the turbulence fields (e.g. surface fluxes of heat, water and momentum) can be highly variable in time and space (Tjernström and Grisogono 1996). Still, most turbulence closures rely on a stationarity assumption in some way, (cf., for example, Pielke 1984). Even models that carry prognostic equations for one or more turbulence moments, so called higher-order closure models, often require specification of a length-scale. This is usually determined from steady-state considerations. Some of the schemes frequently in use assume that while the turbulent kinetic energy (TKE) can be modelled with a time-dependent equation, other second-order moments, like variances and fluxes, can be derived from their steady-state equations (Mellor and Yamada 1974; Yamada and Mellor 1979; Andrén 1990). It is of some importance to investigate the validity of such turbulence schemes when the synoptic-scale or mesoscale conditions are changing in time. This can only be done by a careful analysis of field experiment data, obtained from developing situations that are reasonably well documented. If this question can be resolved, a numerical model can then be used to falsify, prove or improve hypotheses on the cause of an observed development or phenomenon.

This paper deals with observations and simulations of a chain of events, observed during a field experiment at locations close to the south-eastern Swedish coast of the Baltic Sea. During the day in question, a marine stratocumulus layer gradually dissipated, partly due to solar radiative heating and partly (probably) due to synoptic-scale subsidence; the latter is almost impossible to verify directly. The cloud layer, while it lasted, maintained a relatively deep, well-mixed boundary layer. It will be argued that the observed gradual collapse of this marine atmospheric boundary layer (MABL) during the remainder of the day, following the dissipation of the cloud, was under the influence of the synoptic-scale subsidence. The gradual shrinking in depth and increasing stability of the boundary layer provide a simple form of a non-stationary boundary layer. This evolution was sufficiently slow for detailed turbulence measurements to be possible at different times, but fast enough for the turbulence structure to transit clearly from one state to another—from neutral and well mixed to stable and sheared during a period of \(O(\text{hours})\). Simultaneously, the reduced depth of the boundary layer over the sea, and the increase in static stability, continuously changed the interplay between the coastal orography and the atmosphere, and it will be hypothesized that this interplay is the cause of the spatio-temporal development of the low-level wind speed maxima that was also observed. The aim of this study is partly to ascertain if the model turbulence closure can adequately handle the transition from well mixed to stably stratified conditions, with the shrinking MABL depth, and partly to explain the observed development of the mesoscale wind structure.

In the following sections the structure of the MABL obtained from the observations will first be examined, and then numerical simulations of the observed event will be presented.

2. THE FIELD EXPERIMENT

A map of the experimental area, indicating the locations of the research flights, is shown in Fig. 1. The coastline in this region is oriented approximately due west–east for about 80 km, turning south at the western end and north at the eastern end. The turning
Figure 1. The model terrain showing southern Scandinavia and including a zoomed-in horizontal cross-section of the model domain. The terrain heights are indicated by shading (the interval is 30 m and thus the highest terrain is about 250 m) and the grid points by dots. The flight tracks are also indicated for the first (dashed) and second (dotted) research missions on 3 October 1990. The horizontal stacks for both flights are along the WNW-ESE line centred at $x = 0$ in model coordinates.
of the coastline at the eastern end is abrupt at almost 90° and the coastal sea forms an open bay—Bay of Hanö—west of this point. There is also a small archipelago along the eastern part of this coastline. During early summer the sea surface here is normally cold, typically around 8–12 °C, and often substantially colder than the land, frequently generating land–sea surface temperature differences of 5–15 degC. During autumn and winter, however, the sea surface is typically warmer than the land. To cover both types of conditions two experiments were performed in this location, one during spring (29 May to 15 June 1989) and another during autumn (17 September to 5 October 1990).

The dataset consists of airborne measurements and surface-based measurements; the latter from the island of Utlångan, located outside the south-eastern end of the coast (see Fig. 1). The surface-based measurements were made using a 25 m tower, instrumented for mean profile and turbulence measurements, and located close to the shoreline at the southern part of the island, with an open-sea fetch of ∼50 – 280°. The tower was equipped with wind, temperature, and humidity instrumentation at several levels. Turbulence sensors at three levels measured wind, temperature and humidity (Tjernström and Smedman 1993). The bulk of data analysed in this paper will be from the airborne measurements. These were made with an instrument package mounted on a Sabreliner 40A aircraft, operated by the Testing Directorate of the Swedish Defence Material Administration, and made available for atmospheric research on a rental basis. Wind measurements were made using the so-called ‘radome gust probe’ technique (Brown et al. 1983), where the local wind vector relative to the aircraft is inferred from measurements of the pressure distribution over the radome of the aircraft, ambient air temperature and static pressure.

The aircraft horizontal velocity vector, vertical acceleration and attitude angles relative to the earth were obtained from an Inertial Navigation System, a Litton LTN-72. Air temperature was measured by a Rosemount 102EAL total temperature probe, humidity by an EG&G dew-point hygrometer for a low-rate stable reference humidity and a Lyman-α instrument for fast-response humidity. In addition to the above, measurements of long-wave and short-wave radiation, cloud liquid water, surface temperature, and radio altitude were also made. The measurement system is described by Tjernström and Friese (1991) and Tjernström and Samuelsson (1995), including the calibration procedure and estimates of measurement uncertainty. In all, 13 and 14 missions were flown in the two experiments, respectively. In this paper, the last two missions of the second experiment, performed on 3 October 1990, will be considered. These two research flights were flown in the morning (1010–1204 LST) and in the afternoon (1323–1516 LST) respectively. (LST is Local Standard Time which is GMT +1 hour.)

3. The model

The MIUU* meso-γ-scale model used in this study is a three-dimensional (3D) hydrostatic model, with a consistent higher-order turbulence closure. The vertical coordinate is transformed into a terrain-influenced coordinate system (Pielke 1984). The turbulence closure is an improved, consistent version of the ‘Level 2.5’-closure (Yamada and Mellor 1979), in the hierarchy of closures introduced by Mellor and Yamada (1974). It carries an improved description for the pressure-redistribution terms (the ‘near-wall’ correction) and an algorithm to keep second-order moments realizable (Andrén 1990). The model also includes routines for subgrid-scale condensation and radiation, as well as for surface energy balance, but the latter is not used here for the sake of simplicity. The model has previously been applied to a variety of situations, including terrain-induced

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flows (Tjernström 1987a, 1988a, 1989; Enger 1990a; Enger and Tjernström 1991; Enger et al. 1993; Koracin and Enger 1994; Grisogono 1995; Enger and Grisogono 1998), coastal flows (Tjernström and Grisogono 1996, 1999; Grisogono and Tjernström 1996; Grisogono et al. 1998; Cui et al. 1998; Rogers et al. 1998; Tjernström 1999), dispersion calculations (Enger 1983, 1986, 1990b), marine stratocumulus (Tjernström 1988b; Tjernström and Koracin 1995; Svensson and Tjernström, personal communication) and air chemistry (Svensson 1996a, b; Svensson 1998; Svensson and Klemm 1998). It has thus been thoroughly examined for a variety of flows and is well documented. Detailed descriptions are given by Tjernström (1987a, b). Shorter descriptions are given by Tjernström (1988a) and Enger (1990a). Also, see Andrén (1990) for a detailed discussion of the turbulence closure.

(a) The grid

The terrain was extracted from a ~500 m resolution terrain database, and was averaged to the model grid. The horizontal grid expands towards the lateral boundaries, to achieve maximum resolution in the central parts of the domain while moving the boundaries away from the area of interest. A simple radiative boundary condition is applied at the lateral boundaries. Maximum resolution is provided at the domain centre, chosen to be at the south-eastmost corner of the coastline. The horizontal domain size of the model is roughly 400 × 300 km², and the finest resolution is 2 × 3 km². The vertical grid also expands log-linearly towards the model top at 2 km, with the finest resolution at the surface, 1 m, degrading to ~100 m close to the model top. Fig. 1 shows the distribution of the model grid points and the flight tracks for the two flights, along with the terrain.

(b) Initialization and boundary conditions

Initial and boundary conditions were chosen on the basis of the measurements and the prevailing synoptic conditions during the flights. The fact that this is an attempt to simulate an observed flow is an important factor in prescribing the initial conditions. The model simulations were initialized dynamically: the model is given horizontally homogeneous temperature, humidity and wind fields and run through a pre-integration period with realistic surface forcing, during which the model fields adjust gradually to a realistic quasi-balance. Cui et al. (1998) investigated the necessary length of the pre-integration period, and found that buoyancy and buoyancy-inertia waves generated by the initially unbalanced fields had propagated out of a model domain of an even larger size within ~12 hours. The simulations presented here were initialized at 1800 LST on the day preceding the events, and no data for the first 12 hours are used.

The pressure-gradient terms in the equation of motion in this model are decomposed into a resolved mesoscale part and one representing the background (synoptic scale) forcing (see Pielke (1984), p. 124). The model thus requires an assigned background synoptic-scale pressure field. For the sake of simplicity, the synoptic-scale flow was specified as a geostrophic wind, estimated from synoptic charts for the day in question, aided by flight-level wind data, and was kept constant in time. Initial profiles of potential temperature and specific humidity were taken as well mixed (constant values) in the MABL, capped by an inversion and stable stratification and constant (low) humidity aloft; these values were also estimated from flight-level data from the actual event. Since estimates of the synoptic-scale subsidence are very difficult to obtain from observations, it was estimated by trial and error as follows. Initially, a one-dimensional version of the model was run, with the appropriate marine boundary conditions (see below), and an
estimated subsidence, taken from the sinking of the MABL top in the observations. This subsidence was then varied slightly, until the bulk evolution of the MABL, the cloud dissipation, and the subsiding MABL top were qualitatively reproduced. This value of the subsidence (corresponding to a divergence of $2.95 \times 10^{-5} \text{ s}^{-1}$) was then prescribed and further refined in fully 3D runs until the simulated MABL on an average resembled the observed one. It must be noted, however, that this set of initial conditions may not necessarily be unique. Although unlikely, there is a possibility that a different set of initial conditions would cause a similar development.

The sea surface temperature (SST) was set constant in time. For the control simulation it was also constant in space at 12.5 °C; in one sensitivity test, a horizontal SST gradient was imposed. The diurnal temperature variation of the soil surface was prescribed using a sinusoidal-type function with an average value of 13.4 °C and a maximum of 15.4 °C at sea level; this was then allowed to decrease with terrain height by $6 \times 10^{-3} \text{ degC m}^{-1}$. Similarly, specific humidity was specified at the surface using potential evaporation over the sea, and a fraction of this value over land. These conditions were transferred onto the lower model boundary using matching surface-layer and roughness sub-layer similarity theories. Together, this allows coastal land surfaces to be influenced by local advection of moist and cold air from the sea, while allowing for a strong experimental control over the surface forcing. Although there appears to be a certain amount of arbitrariness in the total set-up of the initial and boundary conditions and the synoptic-scale forcing, and some assumptions made seem crude, it is felt that it is very unlikely that one could find a different set of conditions that, within the constraints of the observed general conditions, will reproduce the observed development. The assumptions involved also have the advantage of providing a great degree of control over the simulated flow; this facilitates the sensitivity tests performed.

4. **MEASUREMENT RESULTS**

Figure 2 shows the profiles of wind speed and direction, potential temperature and specific humidity taken at the beginning and end of each flight. Two features clearly stand out: one, the sinking of the MABL top, from $\sim 500$ m before the first flight (solid) to $< \sim 30$ m (below the minimum altitude of the aircraft) by the end of the second flight (dotted, see, for example, Fig. 2(c)) and two, the change in the MABL structure from well mixed (with essentially all variables in the first profile showing this) to a stratified state, with increasing wind shear and stability. The MABL is also heated and dried, presumably through entrainment of the subsiding air. Spurious variations in humidity aloft, between the first and the following three profiles remain unexplained. A solid cloud layer was present only during the first profile; see the moist adiabatic layer between 250–500 m (Fig. 2(d)). The flight notes reveal the presence of a haze layer over the sea, topped by a decreasing cloud amount of 4–6/8 during the first flight, and $<1/8$ during the second flight. Over land, the clouds remained for a longer duration.

Figures 3 and 4 show an attempt to present a composite picture of the MABL during this day. Each plot is made up of all the data obtained during one flight, from one stack of flight legs each. The stacks were oriented WNW–ESE roughly along the eastern section of the east–west coastline. Data from each flight leg were block-averaged into forty horizontally co-located segments, and the cross-sections were constructed from this matrix of averaged data. For a proper analysis this method would obviously require a stationarity assumption, which is certainly not fulfilled here. However, the variables along each flight leg that lies within the MABL may be assumed to be quasi-stationary during that flight leg, whose duration is $\sim 10$ minutes. Assuming that each variable is
Figure 2. Measured vertical profiles of (a) scalar wind speed, (b) wind direction, (c) potential temperature, and (d) vapour specific humidity. The different lines are for aircraft soundings at ~1030 (solid), ~1200 (dashed), ~1330 (dashed-dotted) and ~1500 (dotted) LST on 3 October 1990.

Conserved in a Lagrangian fashion in this duration, only the height axis of the composite plot would be affected by the non-stationarity. Thus, although strictly incorrect, these cross-sections can still provide a representation of the MABL structure. However, the individual cross-sections in Figs. 3 and 4 may not have existed exactly in reality.

In general, it is seen that during the first flight (Fig. 3), the MABL appears well mixed and without large horizontal gradients; this is particularly true for temperature and humidity. On closer examination, temperature and humidity show minor variations, with slightly higher temperature and lower humidity to the east. The wind field appears to be well mixed in the western half, with more vertical wind shear to the east, while the wind direction is almost constant in the MABL. Here, the winds are from almost due south, more or less perpendicular to the flight legs, while the direction veers to southwest above the MABL. This should be contrasted to the structure in Fig. 4, representative of the second flight. Here the wind speed is clearly inhomogeneous, with a low-level jet in the eastern half, with winds increasing to 13 m s⁻¹. Also, temperature and humidity have significant horizontal and vertical gradients. There seems to be a clear separation of the domain into a western and an eastern half. It is interesting to note that the line
Figure 3. Composite vertical cross-sections of measured (a) scalar wind speed (m s⁻¹), (b) wind direction (°), (c) potential temperature (°C), and (d) vapour specific humidity (g kg⁻¹) from the first flight on 3 October 1990. The cross-sections are taken along the flight track, shown in Fig. 1. See the text for a discussion on the composite technique.

approximately separating the two domains more or less coincides with the downstream location of the bend in the coastline, see Fig. 1.

To summarize the observations of the mean variables, two main features are evident. First, the transition of the MABL with a decreasing depth and increasing wind shear and static stability and, second, the formation of a low-level jet, which is present during the second flight, but not the first. The following chain of events is proposed to have taken place. As long as the marine clouds were present, processes in the cloud provided sufficient turbulence to keep the entire MABL well mixed. Consequently, a relatively deep and low static stability MABL formed, and the impact of the shallow coastal terrain was minor, at least only weakly visible at the location of the flight paths. When these clouds dissipated over the sea, certainly by absorption of solar radiation and possibly also by subsidence, a sinking of the MABL top followed. Assuming the presence of synoptic-scale subsidence accounts for the gradual reduction in MABL depth, this reduction was not possible before, since the strong additional mixing in the cloud layer acted in the opposite direction. It is thus suggested that synoptic-scale subsidence was responsible for the observed gradual transition of the bulk MABL character after the clouds dissipated.
The following explanation seems plausible for the observed situation, in the light of the above analysis. As the stability increases, the momentum flux should decrease rapidly, at least on a time-scale shorter than $f^{-1}$, where $f$ is the Coriolis parameter. Consequently, an inertial oscillation starts and a low-level jet is formed. The formation of the low-level jet through this mechanism was first described by Blackadar (1957). In his description of the ‘nocturnal jet’, the turbulent momentum flux was decoupled at sunset and the instantaneously unbalanced wind then starts an inertial oscillation. Later, a spatial analogue to this phenomenon was found, i.e. when the planetary boundary layer (PBL) is advected out over colder water (Smedman et al. 1993); this phenomenon is quite frequent in the Baltic sea (Källström 1998). Following Blackadar (1957), using the first wind-speed profile (solid line in Figs. 2(a) and (b)), the wind speed at the time of the last profile would be $\sim 18 \text{ m s}^{-1}$ at $210^\circ$. This is much stronger than observed, although the wind direction is roughly correct. However, here the decoupling is neither complete nor instantaneous, resulting in a fractionally damped oscillation (Thorpe and Guymer 1977).

At the same time, the heterogeneity of the MABL also increases, and it is notable that the most significant ‘signal’ in the cross-section of wind speed (Fig. 4(a)) coincides with the upstream kink in the coastline. The coastal terrain may be considered steep if the non-dimensional slope $\Delta = (h_m/l_m)(N/f) \rightarrow O(1)$, where $h_m/l_m$ is the slope of the terrain, $(h_m$ is the height increase over the distance $l_m$) and $N$ is the Brunt–Väisälä
frequency. Initially, when the MABL is close to neutrally stratified, $\Delta \to 0$, since $N \to 0$. However, in the last profile $N \sim 2.4 \times 10^{-2} \text{ s}^{-1}$ (Fig. 2(c)), and then $\Delta \sim 0.6$. On the other hand, the Froude number ($Fr$), which within ‘shallow water’ theory is defined as the ratio of the flow velocity and the phase speed of external gravity waves, is subcritical; $Fr \sim 1$ initially increasing to $Fr \sim 2.5$ towards the end of the second flight. This will impede the upward propagation of the flow adjustment to any downstream blocking. The corresponding Rossby radius of deformation also decreases from $\sim 80$ km to $<40$ km, indicating a narrowing zone of coastal influence. In summary, one would expect an increasing influence of the downstream terrain in a narrowing zone of influence as time continues. This, however, is confined to the region west of $x \sim 0$ km (see Fig. 1). West of this line, a south-south-easterly flow may get blocked by the upstream coast, which here is oriented almost perpendicular, while east of the line much less blocking is present (see Fig. 1). The presence and location of the low-level jet in the wind speed is thus consistent with the MABL development and with absence of upstream blocking, east of $x \sim 0$ km. The jet to the east thus appears to be caused by an inertial oscillation due to the temporal development of the MABL, while the flow west of the kink in the coastline is partially blocked and the jet development seems therefore hindered there.

The above hypotheses will be tested with the help of model simulations, to be presented below. Before that, however, the MABL structure will be examined in more detail, so as to ascertain if the model turbulence closure can describe the details in this time-varying turbulent flow adequately. The developing MABL turbulence structure is illustrated in Fig. 5. Here ‘snap-shot’ second-order moments are extracted from the four aircraft profiles, using a technique described by Lenschow et al. (1988), and discussed in detail by Tjernström and Smedman (1993). As this technique involves filtering, the magnitudes of the moments may be questionable. The profiles do not represent an ensemble average due to the short sampling time, but the structure is deemed accurate. From Fig. 5 it is clear that the cloud layer (250–500 m) greatly influences the MABL structure in the first profile; both TKE and all fluxes have maxima in the cloud layer. Immediately after the dissipation of the cloud layer, the depth of the turbulent layer decreases from about 500 m to 300 m, but the structure of the sub-cloud MABL remains similar. From the two subsequent profiles, it is clear that the reduction in depth of the MABL continues, while turbulence is decaying, and in the final profile the surface heat flux has reversed sign.

Figure 6 shows the time variation of some MABL parameters, used for scaling purposes; MABL depth ($z_i$), friction velocity ($u_*$) and the temperature scale ($\theta_s$). The surface parameters were estimated by a combination of data from the first flight leg (at 30 m) in each flight, assuming this to be in the surface layer, and data from extrapolating turbulence data from each of the four profiles to the surface. The inversion height was obtained from the four profiles only. The markers in Fig. 6 indicate the estimates, while the solid lines are best-fit polynomials. The latter are used for scaling the variance and flux profiles as well as the spectra and cospectra. It is seen from Fig. 6 that during the course of the observations, the MABL depth decreases from about 550 m to below 50 m. The friction velocity decreases from about 0.35 m s$^{-1}$ to 0.2 m s$^{-1}$, and the temperature scale from about 0.04 K to about $-0.04$ K. In Figs. 7 and 8, the second-order moments from all the 14 flight legs are scaled in two ways: with the average values of the above parameters for each flight, or using ‘time-local’ values, taken for the mean time of each flight leg using the polynomials. The fluxes are based on the raw unfiltered data, while the variances are high-pass filtered to remove low-frequency noise (see below).
Figure 5. Measured profiles of (a) turbulent kinetic energy, (b) momentum flux, (c) sensible-heat flux, and (d) latent-heat flux. See the text for a discussion of the analysis technique and Fig. 2 for explanation of the four lines.

The fluxes shown in Fig. 7 appear to follow the linear dependency with height, which is expected for a stationary well-mixed PBL. During the first flight, five flight legs lie within the MABL, while only three lie in the MABL for the second flight. There are differences in the scaled momentum flux, obtained using the two methods of scaling. However, within the scatter of the results, it is impossible to say which method gives the best result. This is also true for the heat flux during the first flight, but not during the second flight, when $\theta_e$ decreases to very small values and eventually changes sign. The heat flux is linear and consistent with a stationary PBL profile with an entrainment of 20% of the surface flux. From these results alone, it appears that a turbulence closure based in part on steady-state considerations will indeed be appropriate.

In Fig. 8, the scaled velocity variances are compared with analytical expressions from Brost et al. (1982) as well as with laboratory data from Townsend (1976). Here there are clear signs of departure from the stationary PBL. The streamwise component is about 20% too small, while the transverse component is somewhat too large. This indicates different decay time-scales for the different turbulent moments, but also reflects the different effects on variances and fluxes by the increasing stability and the subsiding inversion. The ratio between the horizontal components is consistent with a larger than normal directional wind shear (see Fig. 2). The vertical component has the
expected magnitude below $z/z_i < 0.5$, but is significantly smaller in the upper half of the MABL. This is entirely consistent with the damping effect of the inversion which is gradually subsiding, thus more and more restricting the vertical size of the turbulent eddies from above. The typical turnover time for this PBL, judging from the friction velocity, is of the order of 20–30 minutes. During that time, the inversion height lowers by $>50$ m. Even though the subsidence by itself is typically an order of magnitude smaller than the friction velocity, the former will affect the local turbulence directly, rather than via the surface. Compared with Brost et al. (1982), the normalized TKE is 10% too small, and decreases with height to about 30% too small closer to the inversion. At the same time, these values are in reasonable agreement with Townsend (1976). The leading idea behind the so-called ‘level hierarchy’ of closures used in the modelling part of this study (Mellor and Yamada 1974) is that while some moments (here the variances or the TKE) are modelled with a time-dependent equation (i.e. allowed to deviate from their steady-state values), other moments (the turbulent fluxes) are modelled to be in steady-state. This hypothesis appears to be supported by this analysis.

More information on the turbulence structure can be extracted from the power spectra of the velocity components shown in Fig. 9. In order to compare spectra from different heights, conditions and times, they must first be normalized. However, of the different scalings suggested in the literature for different conditions, none are entirely applicable here, since the PBL here develops from a well mixed to a stably stratified
Figure 7. Vertical profiles of scaled (a) momentum flux and (b) sensible-heat flux, as functions of scaled height. Open symbols (boxes and circles) are for the first flight on 3 October 1990 while crosses and pluses are for the second flight. Boxes and pluses are values scaled with average parameters, constant for each flight, and circles and crosses are scaled using time-local parameters. The solid lines indicate the linear dependency, assuming no effects of entrainment, while the dashed line in (b) is the linear profile, assuming an entrainment flux equal to 20% of the surface flux. See the text for a discussion.
Figure 8. As Fig. 7 but for the velocity variances (a) along-flow, (b) across-flow, and (c) vertical components. The solid lines indicate the profiles from Brost et al. (1983) while the dotted lines are taken from Townsend (1976).
condition. Thus, while the PBL appears well mixed at least initially, convective scaling does not really apply throughout, only possibly for the first two flight legs. Neither does stable scaling apply throughout, only for the very last flight leg. Also, surface-layer scaling will only apply close to the surface. As a compromise between all these different scalings, the spectra are here scaled so that their shape conforms to mixed-layer scaling in the inertial sub-range (Kaimal and Finnigan 1994, p. 47), but each individual spectrum is scaled using the 'time-local' parameters, i.e. MABL depth and surface heat flux that are functions of time. The aircraft true air speed is used as a proxy for the flow speed in the normalized frequency \( n = f z_i / U_a \), where \( f \) is frequency, \( z_i \) is the MABL depth, and \( U_a \) is the true air speed of the aircraft). Measurements from the two flights are plotted separately. Only spectra at levels within the MABL are scaled using the actual values obtained at those levels, while those above the MABL top are scaled using values from the lowest flight leg in each flight. The purpose is to illustrate the difference in variances at different frequencies, within and above the MABL.

For both the flights there is a suppression of the magnitude as well as the scale of the variances in the vicinity of the inversion. This is most clearly seen for the vertical component. Thus, even in the lower inversion, the power spectra still contain a well-established inertial sub-range, as indicated by the spectral slope, but with a suppressed energy containing range with smaller and vertically distorted eddies. For the mixed layer, the peak in the spectra for the vertical component should grow to about 1.5\( z_i \) in the upper half of the MABL and then decrease again towards the MABL top. For the first flight here, the maximum values in mid-MABL are lower, at about 2\( z_i \) (n ~ 1). The lowest and highest legs, at \( z_i / z = 0.1 \) and 0.8, have their peaks at \( z_i / z = 0.2 \), instead of at \( z_i / z = 0.6 \) and \( z_i / z = 1.4 \) times the MABL depth, respectively, as suggested by Caughey and Palmer (1979) for those heights. For the second flight the corresponding peaks are all between 0.5 and 1.0\( z_i \) for all four flight legs within the MABL. This is also smaller than that expected for a mixed layer, but the mixed-layer assumption is clearly violated here. For both flights the vertical spectra falls off at lower frequencies, although there are somewhat higher relative levels of energy at low frequency in the later flight than in the first. In contrast, all power spectra for the horizontal wind-speed components have significant low-frequency contributions, but most show a well pronounced spectral gap at a scale around 5–10 times the MABL depth (i.e. n ~ 0.1). Low-frequency energy above the MABL is comparable in magnitude with that within the MABL, which indicates that this energy is due to a larger-scale forcing that does not originate in the boundary layer. The longitudinal spectra for mid-MABL flight legs from the first flight show dominant scales in the expected range, \( z_i / z = 1.5 \) times the MABL depth. For the second flight, however, the wavelength for the maximum turbulent energy is shifted to scales that are about half the MABL depth. Here, there is also an enhanced level of energy at the maximum (Fig. 9(e)), compared with interal sub-range values. This could be due to 'fossil turbulence', i.e. the inertial sub-range of the spectra adjust more rapidly in time than does the energy-containing range. The corresponding cospectra (Fig. 10) also show that the fluxes occur at eddy sizes that are smaller than in the steady state well-mixed PBL. In particular, the heat flux during the second flight (Fig. 10(d)) indicates a strong distortion of the eddies. Also, in the inversion, where the flow is still turbulent, but strongly affected by the stability as evident in the power spectra, the fluxes of momentum and heat are strongly diminished or even zero. It is interesting to note from the flight legs above the MABL during the first flight that at the frequency band \( n = 0.2–2.0 \) the spectra fall off as \( n^{-2} \) (Figs. 9(a) to (c)). This is an indication of gravity waves (cf., for example, Stull 1988, p. 533), and these waves seem to have vanished in the second flight. The frequencies correspond well to the expected frequency of gravity waves at
Figure 9. Scaled power spectra for the (a) and (d) vertical, (b) and (e) along-flow, and (c) and (f) across-flow components from (a) to (c) the first flight and (d) to (f) the second flight on 3 October 1990. The thin solid line is used for those flight legs that appear to be within the marine atmospheric boundary layer (MABL) proper, the dashed line is used for the flight legs that appear to be in the vicinity of the inversion, and the dotted lines are those flight legs that are clearly above the MABL. Furthermore, the lines are marked from the surface and upwards with symbols. First flight: plus (z/z_l = 0.11); cross (z/z_l = 0.22); circle (z/z_l = 0.36); pentagram (z/z_l = 0.49); hexagram (z/z_l = 0.79). Second flight: plus (z/z_l = 0.20); cross (z/z_l = 0.39); pentagram (z/z_l = 0.62); hexagram (z/z_l = 0.89). The thick dashed-dotted lines represent the inertial sub-range from steady-state mixed-layer scaling (upper line) and the n^{-2} spectral slope (lower line). See the text for a discussion of the scaling.
this height, with $N \sim 3 \times 10^{-2}$, taken as an average of the inversion strength from the first two profiles, giving $n \sim 0.3$. This is also consistent with the significant fluxes above the MABL, of momentum and more so of sensible heat, around $n \sim 0.1–0.2$.

In summary, while the spectral analysis clearly shows a disturbed turbulence structure, with eddy sizes vertically distorted and significantly smaller than those in a well-mixed PBL, the total magnitude of the variances appear ‘normal’, at least in the lower half of the MABL. In the upper half of the MABL they appear somewhat lower, at least for the vertical component. Moreover, the average fluxes do not seem to bear any influence of the significant changes in bulk MABL structure, thus indicating that a turbulence closure with a time-dependent TKE, accounting for the changing PBL, and fluxes and variances derived from their steady-state equations, should be appropriate in a model intended to simulate the kind of MABL encountered here.

5. THE MODEL SIMULATIONS

Although the control run was set up to reproduce the event that was measured, it must be realized that with the degrees of freedom available in the model atmosphere, a complete match between the simulation results and the measurements is quite difficult to obtain. It is not even the aim here to achieve such a complete match. The aim is to reproduce, with a simplified forcing, the dynamic situation to an acceptable degree of agreement with measurements. Having achieved dynamic similarity, and also a
reasonable quantitative agreement between the measurements and the control run, a number of sensitivity simulations were performed. The applied forcing has been varied in a simple manner, so as to facilitate these sensitivity studies.

The sensitivity simulations include five runs:

(i) a 1D simulation, i.e. a homogeneous MABL without the coastal forcing;
(ii) a 3D simulation where the coastal terrain to the east has been extended, thus forming a continuous upstream coastline from east to west with no kink anywhere;
(iii) a 3D simulation with slightly less MABL humidity at the initial time, thus causing a cloud-free MABL;
(iv) a 3D simulation without the imposed subsidence; and
(v) a 3D simulation with an imposed homogeneous east–west SST gradient of 0.005 degC per km, set up in such a way that the SST at the location of the measurements is roughly unchanged, while the SST to the west is a little higher and that to the east, lower.

The extended coastline, (case (ii) above) was set up as a continuous straight coast, continuing eastward of the kink that is present in the unperturbed coast, sloping to 150 m over 50 km. The prescribed SST gradient, in case (v), has some support from the aircraft measurements of surface radiation temperature. Although each flight covered an area too small to be able to measure such a gradient, examining several flights from the whole field experiment indicated that such a gradient was plausible. However, changes in time during the entire experiment somewhat obscures this, and makes a definitive conclusion difficult.

(a) The control run

In Fig. 11, model profiles of wind speed and direction, potential temperature and specific humidity are shown. The location of these profiles roughly, but not exactly, coincides with that of the measured profiles, while their times approximately correspond to those of the four aircraft profiles, shown in Fig. 2. Even from a cursory examination of Fig. 11, it is clear that the bulk properties of the observed MABL are approximately reproduced by the simulation: the shrinking of the boundary-layer depth and the formation of a wind speed jet. From the first profile (the solid line), it is clear that well-mixed conditions prevail initially. The MABL, capped by a sharp inversion, is about 500 m deep, and within the mixed layer all the variables are nearly constant with height. The wind speed is 8–9 m s\(^{-1}\) and the temperature is \(\sim\)11 °C, agreeing well with the measurements. The wind direction, at 203°, is about 20° too westerly, and the specific humidity at \(\sim\)7.2 g kg\(^{-1}\) is about 0.5 g kg\(^{-1}\) too low. At this time in the simulation, the cloud (not shown) still prevails, and the mixing induced at the cloud top and in the cloud is responsible for the well-mixed layer.

As time proceeds, the MABL top subsides and the wind-speed profile first becomes more continuously sheared, and from the third profile and onward, a low-level jet is formed. Simultaneously, the shear in the wind direction increases and becomes quite similar to the measurements, from 180° at the surface to 220° at the MABL top. The MABL gradually becomes more stably stratified, and warmer. However, the complete collapse of the MABL, as seen in the final profile of the measurements, is apparently not captured in the simulation. The temporal variability in the measured humidity profiles is also not completely reproduced, although the trend with drying of the air above the inversion is captured. This development is further illustrated in the time–height cross-sections, see Fig. 12. Here again, the reduction of the MABL depth and the weakening of the inversion are illustrated. The dissipation of the cloud layer between 1030 and 1200 LST, as well as the presence of the cloud before that, are borne out by the humidity
and temperature cross-sections. The formation of the low-level wind-speed maximum at the MABL top is also evident. The time–height cross-sections indicate that towards the end of the period, the MABL, although shallow, has reached a nearly steady state, at least in the wind speed and temperature, unlike in the observations where it continues to subside all the way to the surface.

Thus, the wind profiles are quite adequately reproduced, while the thermodynamic structure, with the reduced depth and increasing stability, develops somewhat too slowly in the simulation. This development is brought about by the imposed synoptic-scale subsidence in the simulations; without this subsidence most of the temporal variability would vanish (see below). Increasing the subsidence would certainly increase the rate of shrinking of the MABL top. However, it also causes the clouds to dissolve too soon, thus altering the whole chain of events and causing an entirely different development than was observed.

West–east vertical cross-sections of the wind speed and potential temperature, taken roughly in alignment with the flight track, are shown in Fig. 13 for two instants corresponding to those before the first flight and after the second flight. In Fig. 13(a), it is seen that already at this time a jet has started to form east of the end of the flight track. However, in the part of the cross-section corresponding to the measurement location,
there is only an indication of the corresponding gradient. This is in quite good agreement with the composite structure obtained from the measurements, as shown in Fig. 3. The wind speed, further east of the measurement location, has already reached a maximum of \( \sim 14 \, \text{m s}^{-1} \). There is also a suggestion of a horizontal temperature gradient, although the bulk of the MABL remains well mixed; this also conforms to the measurements. West of \( x \sim 0 \, \text{km} \) (Fig. 1), the flow appears to be more blocked, with lower wind speeds around \( 8-9 \, \text{m s}^{-1} \), and a more well-mixed structure. This structure confirms the hypothesis, proposed earlier, of an inertial low-level jet, forming due to a frictional decoupling as the mixed layer starts to collapse, but being increasingly blocked by the upstream coastline to the west.

In Figs. 13(c) and (d), corresponding to the end of the second flight, the jet is seen to have spread into the eastern Hanô Bay, and is now seen also at the location corresponding to the measurements (cf. Fig. 4(a)). The maximum wind speed is \( 13.5 \, \text{m s}^{-1} \) at 250 m, quite in agreement with the measured values; further east the wind speed reaches magnitudes of \( 14.5 \, \text{m s}^{-1} \). However, the blocking of the flow in the bay appears stronger in the measurements than in the simulation, with winds as low as \( \sim 10 \, \text{m s}^{-1} \) in a deeper layer. The simulated jet is obviously broader and the horizontal wind-speed
Figure 13. Vertical cross-sections of simulated (a) and (c) scalar wind speed (m s\(^{-1}\)) and (b) and (d) potential temperature (°C) from the control run. The cross-sections are taken roughly along the flight track in Fig. 1 at times corresponding to (a) and (b) the start of the first and (c) and (d) the end of the second research flight on 3 October 1990.

The gradient at the jet height is weaker, as compared with the observations. Notwithstanding these discrepancies, the structure with a low-level wind-speed maximum, developing under a local minimum in wind speed, is quite adequately captured.

Figures 14(a) to (d) show west–east cross-sections at the upstream (southern) model boundary, taken at the same instants as those in Fig. 13. Figure 14(a), showing the total wind speed, indicates that a weak jet has already formed here by the time of the first flight. The temperature field (Fig. 14(b)) is shallower and more heterogeneous than that at the measurement location, indicating that the clouds have already started to dissipate here, even at this time. This is in agreement with the flight notes of visual observations from an altitude of ~2 km, before the first profile, which indicate that dissipation of the cloud field started offshore and propagated into Hanö Bay. Figures 14(c) and (d) reveal a more shallow and homogeneous MABL at the upstream boundary than closer to the coast (cf. Figs. 13(c) and (d)). There is a marked and relatively homogeneous low-level wind-speed jet across the entire domain, with wind speeds reaching 16 m s\(^{-1}\), in a shallow MABL that is 100–200 m deep.

Figures 15(a) and (b) show horizontal cross-sections of the wind speed at 1030 and 1500 LST respectively. These cross-sections are chosen at an altitude of about
50 m, so as to be able to be well within the MABL at both times. Also shown are the wind vectors, that clearly mark the wind speed and direction. It can be noted that close to the coastline, the flow is partially blocked and turns in response to the coast. A comparison of Figs. 15(a) and (b) would indicate that in the latter an increased blocking is evident, thus supporting the arguments given above. Figures 16(a) and (b) show similar horizontal cross-sections of the potential temperature, taken at the same height. They show that there is very little horizontal inhomogeneity in the potential-temperature field. The cold pool adjacent to the coast is an indication of a slight blocking already at this time, that does not propagate upstream due to the supercritical flow. Figure 17 shows the evolution of the cloud layer, at the same location as the profiles shown in Fig. 12, before it dissipates. It is seen here that by 1200 LST the clouds have dissipated totally. Also shown is the evolution of the TKE. There is a marked decrease in TKE after the dissipation of clouds, although the TKE never completely vanishes. One major difference between this TKE distribution and the measurements (Fig. 5(a)) is that the local maximum in TKE, observed inside the cloud layer in the first observed profile, is not obtained in the simulation. In the latter, the maximum is obtained rather close to the surface. It is not clear with the present investigations as to why this difference occurs. The reason why the maximum in the simulated TKE (when clouds are present) is obtained near the surface is that, in a cloud topped MABL, continuous net cooling at
Figure 15. Horizontal cross-sections of scalar wind (m s\(^{-1}\)) (thin solid lines) and wind direction (indicated by arrows) from the control run, taken at 50 m height and at the times corresponding to (a) the start of the first and (b) the start of the second research flights on 3 October 1990. The thick lines mark the coastline.
Figure 16. Same as Fig. 15, but for potential temperature (°C).
the cloud top will cool the entire MABL, and a constant temperature at the surface will ensure convective conditions that will generate more TKE near the surface than near the cooled cloud top.

In summary, the control run captures much of the dynamic situation that was observed, although details differ, when compared with the measurements. The simulated MABL structure at the location of the measurements conforms to the observations, including the formation of the low-level wind-speed maximum, the gradual collapse of the MABL, and the apparent blocking of the low-level flow to the west by the upstream coast. Together with the spatial structure in the entire model domain and its development in the simulation, all these features confirm the argument that the jet is an inertial oscillation, triggered by the collapse of the well-mixed layer as the clouds dissipate, and that the spatial structure of the jet is a consequence of the increased blocking by the upstream coast as the MABL becomes more shallow and stable.

(b) Sensitivity tests

The first four sensitivity simulations will be considered in two pairs, since they involve changes to two types of forcing: changes to the surface forcing (altering the terrain) and changing the internal structure (altering the clouds).

The profiles now shown are taken at the same location as those in the control run. Figure 18 shows a comparison of the sensitivity run with no terrain at all, to the
Figure 18. Some results from the sensitivity runs with altered terrain, showing profiles of (a) and (c) scalar wind speed and (b) and (d) potential temperature from the run without terrain ((a) and (b)) and the run with extended terrain ((c) and (d)). The vertical cross-sections show scalar wind speed (m s$^{-1}$) from the sensitivity run with extended terrain at (e) 1030 and (f) 1500 LST, roughly along the stack of flight tracks before the first and after the second research flights on 3 October 1990. The different lines correspond to the times of the aircraft soundings in Fig. 2, at 1030 (solid), 1200 (dashed), 1330 (dashed-dotted) and 1500 (dotted) LST.
run wherein the upstream coast was extended to form a continuous barrier. When the terrain is absent, the clouds dissipate earlier. Consequently, the MABL is shallower (compared with the control run) even at the time of the first flight, and a stronger low-level jet develops as a result (Fig. 18(a)). The MABL depth continues to be gradually reduced by the imposed subsidence, but here also, as in the control run, the observed complete collapse is never obtained (Fig. 18(b)). The wind speed at the time of the first measured profile is much higher than in the control run, due to the difference in temporal development (the jet develops earlier) as well as the absence of the blocking terrain. The difference in wind speed when compared with the control run, however, becomes smaller subsequently. In contrast to this situation, when the coast is extended the jet is weakened (Fig. 18(c)). However, the clouds remain longer in this case. Understandably, the perturbation to the vertical wind field induced by the coast is critical to the maintenance of the clouds. This feature is consistent with the earlier observation that the clouds dissipate faster at the southern end (where there is no terrain) in the control run, and to the visual observations from the aircraft that were mentioned earlier. The temporal development is similar to the control run (Figures 18(c) and (d)), although the jet develops somewhat more slowly (compare the second profiles in the two cases). The spatial structure is, however, much altered in this run. The structure of the wind field at the time of the first flight (Fig. 18(e)) is similar to the control run (cf. Fig. 13(a)), but the wind speed is lower. Both vertical and horizontal wind-speed gradients are much reduced, and while the jet has already formed towards the east in the control run, there is no indication of a jet here. As the MABL depth decreases in time, and the jet develops, the wind field becomes even more homogeneous (Fig. 18(d)). The jet is shallower and extends over the entire cross-section, at ~300 m in the vertical, while the maximum wind speed is ~3–4 m s$^{-1}$ lower than in the control run. Thus, it appears that the presence of the coastal terrain has a positive effect on the maintenance of the clouds, but weakens the jet by blocking it.

Figure 19 shows the results of two runs, one without subsidence (Figs. 19(a) to (d)) and one where the clouds did not form at the initial time (Figs. 19(e) and (f)). From the first run it is clear that removing the subsidence causes the clouds to linger during the entire simulation; the cloud layer is also somewhat thicker. The spatial and temporal variability observed in the control run, as well as in the measurements, are now more or less entirely removed. The MABL is in a quasi-steady state, governed by the presence of the clouds, and the small variability exhibited is due to the diurnal variation of the cloud field as it is subjected to varying radiative conditions. The MABL is thus continuously well mixed, with almost constant wind speed, in time and with height. The temperature profile reveals the marked local heating in the cloud layer due to the release of latent heat, but also a gradual overall cooling in time. The latter is due to the radiative deficit. During this time of the year, the absorption of solar radiation in the cloud layer is insufficient to off-set the negative long-wave radiation balance at the cloud top and, thus, the cloud layer is cooled. Since the lower-boundary condition is given by a fixed SST, this causes the MABL to become more and more unstably stratified; this supports the MABL mixing. (In reality, the SST would be reduced also, but only very slowly.) If clouds were never present at all (Figs. 19(e) and (f)), two things happen. First, the MABL is very shallow. In fact, the MABL depth in this run is the smallest of all the runs. However, since there is no presence and eventual dissipation of the clouds, the large temporal variability induced by these factors in the control run is absent here. There is only a very small variability. Second, there is a strong vertical shear in the wind speed, but no low-level jet. The jet therefore only forms when there is a transition from deep, well mixed and neutral state to a shallow and stable state.
Figure 19. Simulated vertical profiles of (a) and (e) scalar wind speed, (b) wind direction, (c) and (f) potential temperature, and (d) total specific humidity from sensitivity runs (a) to (d) without synoptic-scale subsidence and (e) and (f) without any clouds. The different lines correspond to the time of the aircraft soundings in Fig. 2, at 1030 (solid), 1200 (dashed), 1330 (dashed-dotted) and 1500 (dotted) LST on 3 October 1990.
Finally, on performing sensitivity test (v), it is found that the SST gradient gives rise to a similar gradient in the air temperature above, which results in a thermal wind whose direction is opposite to the mesoscale wind field. This weakens the jet marginally. Apart from this effect, which is to be expected, the SST gradient does not change the MABL evolution in any other way (Grisogono and Tjenström 1996).

6. Discussion and Conclusions

An analysis of airborne meteorological measurements taken in the Baltic Sea off the Swedish south-west coast is presented. The numerical simulations of these measurements, that were performed subsequently, are also presented and discussed. The measurements reveal the transition from a well-mixed cloud-capped MABL to a shallow stable boundary layer over a period of about 5 to 6 hours. Coinciding with the collapse of the MABL, a low-level jet develops in the eastern section of the measurement area. Downstream of the line dividing the two areas (with and without the jet), there is a kink in the coastline, and the jet develops where there is no terrain barrier. The terrain acts to block the flow and hinder the development of the jet in the initial stages, although subsequently the blocking is overcome and the jet forms near the coastal terrain as well, in the vicinity of the measurement area.

The analysis of the turbulence structure reveals a MABL that is strongly perturbed by its subsiding top and the increasing stability. Consequently, the scales of the turbulent eddies, as apparent in the power spectra and cospectra, are smaller than those expected for a neutral to slightly stable PBL. The normalized wind-speed variances also show a related, but much smaller, deviation from the expected structure. There is also relatively more energy in the transverse and vertical components in the lower MABL than in the longitudinal, which is interpreted here as evidence of a strongly sheared environment. The magnitude of the vertical-velocity component is also reduced with altitude, an indication of the subsiding inversion. These deviations are presumably caused by the forcing by the subsidence, causing a MABL that is not in internal steady-state balance. However, the normalized fluxes seem insensitive to this forcing, as they conform roughly to the ideal undisturbed MABL structure. It thus appears that a PBL parametrization that carries TKE as a prognostic variable, but estimates the remaining second-order moments from steady-state considerations, should be adequate to handle this type of a relatively slow transition.

In summary, the sensitivity simulations clearly confirm the hypotheses that are proposed on the basis of the observations: the initial well-mixed MABL is due to the presence of the stratocumulus clouds (its dissipation is probably caused by synoptic-scale subsidence), and the low-level jet forms as an inertial oscillation when the well-mixed MABL breaks down as the clouds dissipate. The absence of clouds altogether, caused the shallowest and most stratified MABL to form, but in this case no low-level jet is ever formed as there is no transition associated with the dissipation of clouds. Removing the subsidence caused the clouds to last throughout the simulation period, while being continuously cooled. Direct absorption of solar radiation by itself was insufficient to warm the cloud layer. Here also, no jet forms when the clouds linger. Thus, it appears certain that there has to be a reasonably abrupt transition, from the well-mixed MABL with high levels of momentum transfer, to a stable situation with small momentum flux, to enable jet formation. Consequently, in all runs where this transition occurred, a low-level jet formed. The timing of the formation, and the strength at any given time, appear to be related to the time elapsed since the clouds dissipated at any given location. Thus the jet forms first at the southern model boundary, where the clouds
dissipate first. As the jet forms, its spatial distribution is dependent on the downstream coast. In the run with an extended coastline a jet forms, but, as the MABL becomes shallower and more stably stratified, the continuous blocking by the upstream coast causes a very homogeneous distribution wind field with a much reduced jet strength. The formation of a low-level jet with the spatial distribution similar to the observations appears only in the run with a MABL transition and with a kink in the downstream coastline. The observed jet is thus caused by the inertial oscillation, but the absence of the jet in Hanoi Bay is caused by the local blocking of the flow; this blocking is not present east of the kink in the coastline and thus the jet lingers and develops to the east. Finally, these simulations show the potential of utilizing a numerical model while analysing experimental data.

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