Geophysica (1985), Vol. 21, No. 2, pp. 115–126
Printed in Finland

551.511.61

THE SEA BREEZE AND URBAN HEAT ISLAND CIRCULATION IN A NUMERICAL MODEL

by

HANNU SAVIJÄRVI

Department of Meteorology
University of Helsinki

Abstract

The sea-land breezes, the urban heat island circulation and their interaction are studied using a two-dimensional 10 level 4 km grid model forced by given (observed August) surface temperatures. The warm (ΔT = 8 °C) 12 km wide city produces a steady »city breeze», strongest at 8 p.m. with surface inflow ~0.9 m/s and maximum rising motion ~3 cm/s at 600 m height. The sea breeze remains weak with the (late summer) parameter values used but the inclusion of a warm city on the coastline enhances the sea breeze considerably and locks it above the coastline. The nocturnal land breeze is also enhanced and is lifted above the weak city breeze which is suppressed to the lowest 300 m layer.

The horizontal average of the simulated breeze is small except in the case of weak offshore prevailing wind. The area averaged flow over non-homogeneous lake and hill terrain is also studied in order to parameterize these subgrid (meso)scale features in a large grid model.

1. *Introduction*

Local mesoscale surface winds are often driven by local surface temperature differences, *e.g.* the sea and land breezes, the mountain-valley winds and the man-made urban heat island circulation. They affect the local weather and climate and air pollution in cities, yet are difficult to study needing a dense observation network. In this work the sea and land breeze, the »city breeze» and their interaction (a coastal city) are studied using a mesoscale numerical model where the circulations are forced by surface temperature and roughness differences between land, sea and city. The parameter values are tuned for Helsinki late summer but the results are presumably qualitatively valid for any coastal (city) area.

Both the sea-land breeze and the urban heat island circulation are well documented and investigated phenomena as reviewed *e.g.* by ATKINSON (1981). Their interaction is, however, less well known and this justifies the present study. The feedback to the large scale flow is also discussed, for the sea breeze and for flow over »lake breeze», area filled with small lakes and hills. This scale interaction has been a rather unexplored area until recently.

2. *The model and boundary conditions*

The numerical mesoscale model is in the present experiments a two-dimensional dry hydrostatic σ-coordinate model in a 4 km grid and 10 levels at the approximate heights of 10, 18, 33, 60, 110, 180, 330, 600, 1100 and 1800 m above the surface. The turbulent transfer of heat and momentum is based on mixing length theory with the eddy diffusivity K depending also on surface roughness at the lowest level and on hydrostatic stability at all levels. This ensures strong upward heat flux during free convection conditions even with weak winds. ALPERT *et al.* (1982) describes the model details and an application to sea breeze over a complex topography (Israel) with good fit to observations while ALESTALO and SAVIJÄRVI (1985) documented an improved model applied to the coastal convergence over a sea gulf. The only change in the present version concerns the pressure gradient force time integration, which has been changed from leapfrog to the Adams-Bashforth scheme in order to keep all terms in a two-level scheme.

In the present experiments the coast and/or the city lies in the middle of the 65 gridpoint area. The surface roughness length z_0 is 1 mm over sea, 10 cm over land and 1 m over the city, which is 12 km large. The coastal slope rises 80 m at 80 km, simulating Southern Finland relief. The slope effect (mountain and valley winds) was found to be very small and does not have any effect on the present results.

The (given) surface temperatures used for the city and land areas are based on observations in Helsinki during a nearly clear, calm warm period in August 15–16, 1973, shown in Fig. 1, after ALESTALO's (1975) screen measurements in the city centre and 10 km to the north of it within a large park area. The city heat island was, between 20–05, about 8 °C warmer than the surroundings; this was also found by spot observations throughout the city area. In the model, a two-term Fourier time series representation was used, shown in Fig. 1 as thin lines, *e.g.* to exclude the weak cloud effect at 10–12 o'clock.

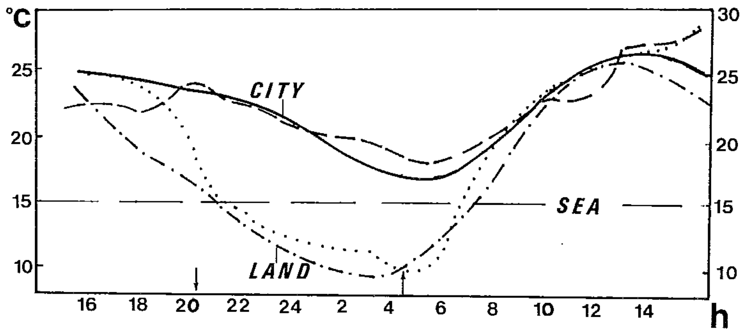

Fig. 1. The temperatures during 15–16 August 1973 on the top of Porthania building (city centre, – – –) and in Viikki (north side of Helsinki, ·····), and the temperatures used in the model for city (——), land (– · –) and sea (– –). Arrows designate sunrise and sunset.

The sea surface temperature is fixed to 15 °C, which was also the initial 10 m level temperature in the model at 04 o'clock. The initial lapse rate was 8.5 °C/km and winds = 0, from which 28 hour simulations were performed with an 8 s time step. The resulting 2nd day 08 fields were very close to the 1st day 08 fields.

3. *Results*

3.1 The city breeze

The urban heat island, *i.e.* a warm, rough zone in the middle of flat land produced a symmetric city breeze in the model (Fig. 2), in which the rising motion above the city at 600 m level was strongest (3 cm/s) in the evening at 20 o'clock while the surface inflow wind toward the city was 0.9 m/s. At 04 the surface flow was 0.4 m/s and rising motion, 0.3 cm/s. The return flow was 0.2–0.3 m/s in a deep 1–2 km layer. The city was warmest compared to the countryside during night hours but the strongest circulation developed as the consequence of unstable daytime conditions.

When the city size was increased from 12 to 20 km and larger, the circulation enhanced accordingly but the nighttime circulation then tended to show a more complex structure with side cells. If a background wind was applied to the city breeze, it tilted downstream during weak winds (geostrophic wind 1–2 m/s) and was swamped by frictional effects during stronger (> 3 m/s) winds.

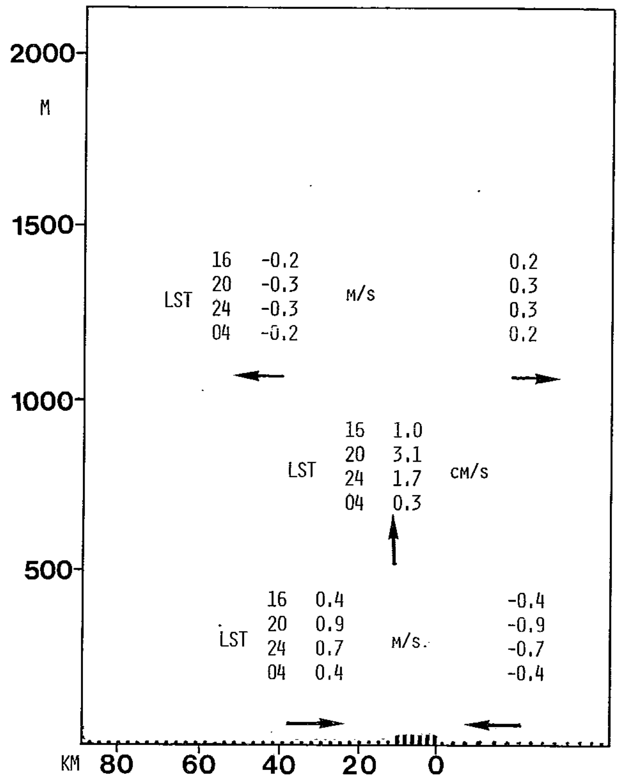

Fig. 2. The urban heat island circulation according to the model.

3.2 Sea and land breezes

Fig. 3 shows the modelled sea and land breeze circulation (*u* and *w* components) at 16 and 04, respectively. The surface wind speeds are qualitatively similar to the observed weak breezes on the southern Finland coasts during late summer when the sea is relatively warm. The sea breeze front (the rising motion maximum) advanced 25 km inland by 20 and disappeared thereafter. The land breeze cell, at 04, is also weak but extends further inland and seaward than the sea breeze. The maximum rising motion is at 16, 1.1 cm/s 10 km inland, and at 04, 0.1 cm/s 30 km to the sea. The behaviour of the modelled sea breeze under variable external conditions (heating, large scale wind, baroclinicity, steep coast etc.) will be a subject for further articles.

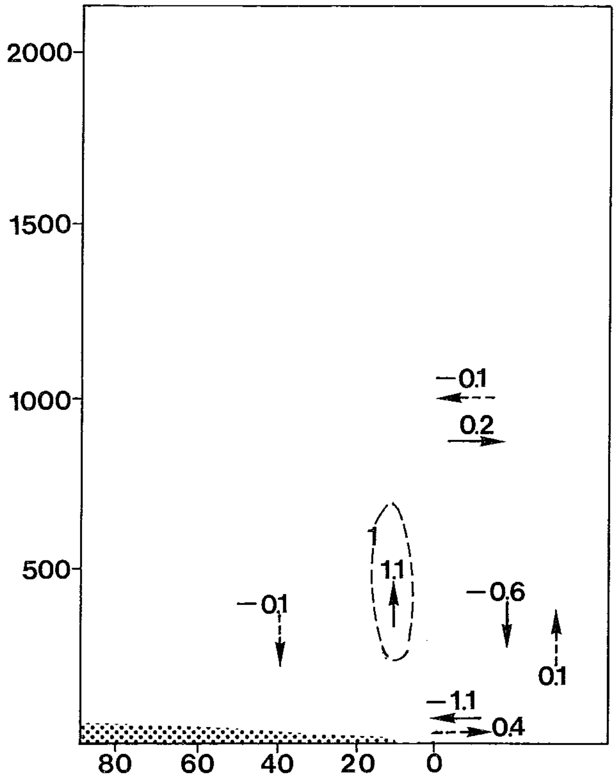

Fig. 3. The wind u (m/s) and w (cm/s) components in the sea breeze at 16 LST (heavy arrows) and in the land breeze case at 04 LST (dashed arrows) according to the model.

3.3 The combined sea and city breeze

When a 12 km wide city is set on the coastline, the extra heating zone activates the sea breeze more than just adding the city breeze effect. This can be seen in Fig. 4 where the afternoon flow field (u and w at 16) is shown over a coastal city. Compared to the »no city» case (Fig. 3), the surface wind near the coastline is increased from 0.9 to 3.7 m/s and the maximum rising motion, from 1.1 cm/s to 9.2 cm/s at the 600 m height above the city. On the landward side of the city there is also »city breeze»; inflow (0.9 m/s) towards the city against the sea breeze.

The dynamical reason for the strong sea breeze is that the warm city creates and maintains an unstable surface layer above the coastline early in the morning so that the sea breeze mechanism can set in even if the actual temperature difference between land and sea is not large and becomes positive quite late.

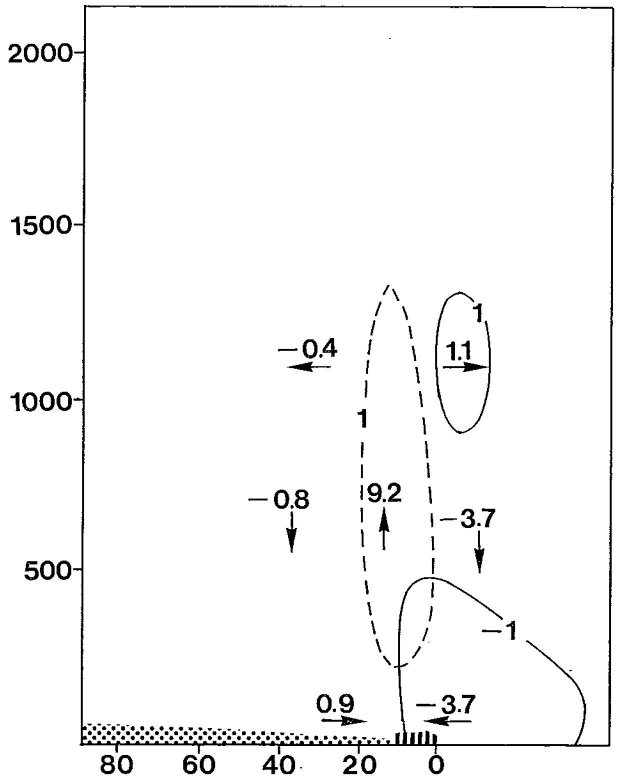

Fig. 4. The wind u (m/s) and w (cm/s) components in the sea breeze case at 16 LST when a 12 km wide city is situated on the coast.

The warm city creates yet another effect: it tends to lock the sea breeze to the coastline above the city. The sea breeze advance inland slowed down in the present experiment, and at 20 in the evening the coastal wind is 2.5 m/s from the sea, the inflow from the countryside is 1.2 m/s and the rising motion is 11 cm/s in a deep 300–1500 m layer only 12 km from the coast. A somewhat similar »locking» effect has been reported on the tidal beaches (ATKINSON, 1981), where the sea breeze is stronger and stays near the coastline during low tide days (when the sand beaches become hot) compared to high tide days when the beaches are covered by water.

3.4 The combined land and city breeze

The modelled sea breeze calms down quickly after 20 o'clock when the land becomes colder than the air. The city breeze continues, however, and the developing land breeze now sets in above the city breeze circulation, which is compressed near the surface. Fig. 5 shows the situation at 04 o'clock. The city breeze blows above the city in a 300 m deep layer with inflow surface winds of 0.7 m/s and rising motion 0.2 cm/s. The land breeze circulation is aloft with rising motion (0.2 cm/s) above the sea and sinking motion (w = −1.2 cm/s) 20 km inland. The circulation is stronger than the pure land breeze in the »no city» case (Fig. 3). The reason for the increase is presumably in the decreased friction as the land breeze rises above the surface in the city area, and in the city breeze, which helps to drive the lower branch of the land breeze. The strong land breeze activates also a secondary weak indirect circulation cell over land, due to mass continuity.

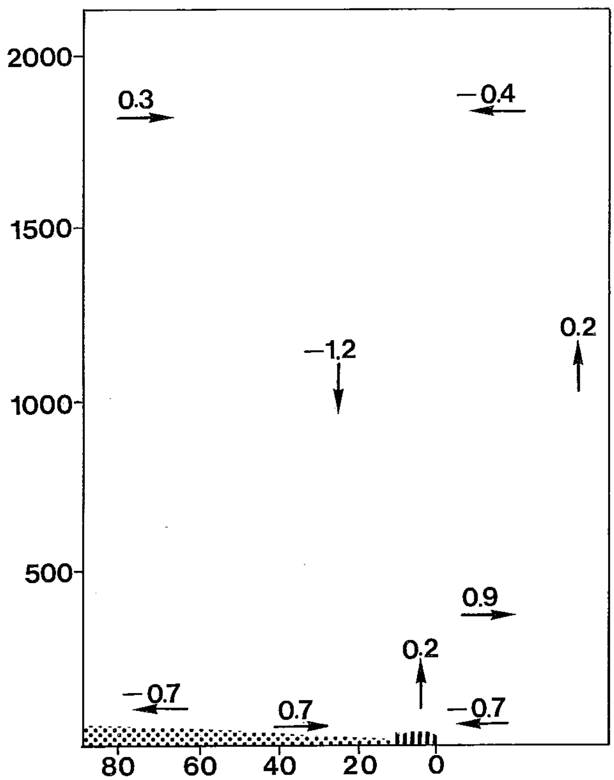

Fig. 5. The wind u (m/s) and w (cm/s) components in the land breeze case at 04 LST when a 12 km wide city is situated on the coast.

4. *The area averaged flow*

In large scale forecast models the subgrid scale effects include also mesoscale phenomena, which should interactively affect (»feed back» to) the large scale flow. The common parameterizations of the subgrid mixing phenomena (*e.g.* those used in the present model) have been developed for homogeneous small scale turbulence and thus the mesoscale organised circulation is not being described if they are used. Normally this is not so serious as the large area average around the gridpoint in a typical 200 x 200 km grid smooths out small nonsystematic features. The present experiments give, however, an opportunity to check what the large scale flow should look like under the influence of local mesoscale phenomena, simply by taking horizontal averages over the 200 km area which includes the mesoscale circulation. A good subgrid parameterization in the large scale model should be able to reproduce the averaged values, given the external conditions.

4.1 Sea breeze

Fig. 6a shows the mesoscale model wind hodographs (wind vectors at 4 hour intervals) for the sea and city breeze, in a single coastal grid point and the horizontally averaged wind over the 200 km coastal area, at the heights of 10, 200 and 1800 m. The single point hodograph shows a clear sea-land breeze loop while the area average remains small. There is, then, only very weak feedback to the large scale flow, which was calm (no background wind) and stable. In some circumstances, however, the area average may become large. This is shown in Fig. 6b where the sea and city breeze effects appear in a 3 m/s background geostrophic wind from land. A weak offshore wind is known to enhance the sea breeze and this is clearly seen in the single point hodographs of Fig. 6. Moreover, the area averaged hodograph now shows large variation around the frictionally reduced mean wind vector, which would prevail in the absence of the sea breeze (and would be produced by the usual parameterization schemes).

In this case the feedback to the large scale is not simple nor small: A strong transverse wind component develops over the sea at low levels in the evening and the area averaged low level wind is almost perpendicular to the geostrophic wind. Even at midnight, the 200 m area averaged wind is 5.7 m/s and 45° left from the geostrophic 3 m/s wind, while in the absence of the coastal effect the 200 m wind would be only about 2.5 m/s. Such a transverse coastal midnight low level jet is not an artifact of the present model but is frequently observed during offshore winds (*e.g.* Hsu, 1979). The parameterization of this effect is obviously

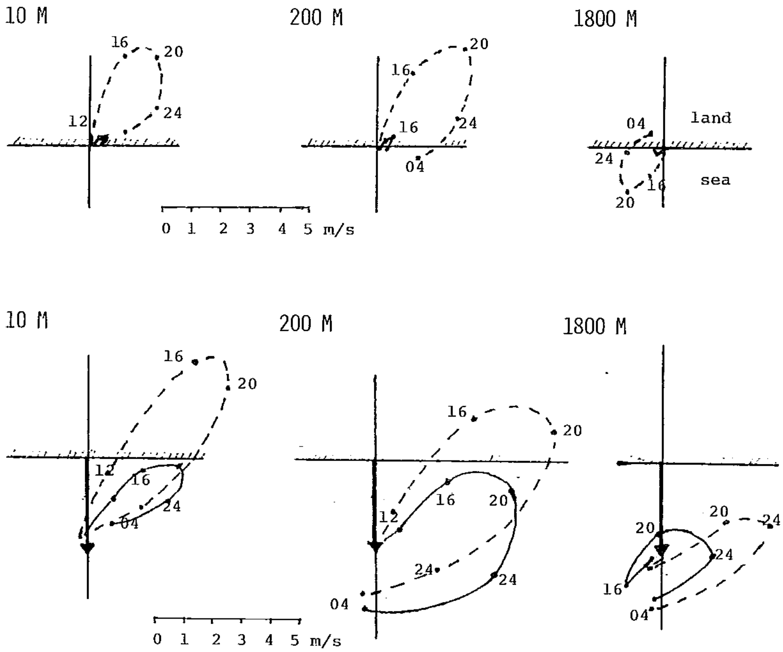

Fig. 6. The model wind hodographs on the coastal zone in one gridpoint (dashed lines), and averaged over a 200 km wide area, at the heights of 10, 200 and 1800 m.
Upper frame: no prevailing flow
Lower frame: prevailing geostrophic flow from land, 3 m/s (weak offshore flow).

important for the planetary boundary layer flow simulation in the coastal area. Mesoscale models could be used to monitor the conditions for such phenomena in order to develop parameterization schemes that would reproduce them.

4.2 Lakes and hills

As a second example, consider forced flow over hills and lakes. The common surface layer parametrization schemes (bulk aerodynamic formulae, Monin-Obukov-theory) have been developed for homogeneous terrain. In nature, hills, rivers and lakes make the surface less homogeneous. Moreover, in stably stratifield flow hills smaller than about 6 km length do not influence the flow much above the surface layer while larger hills tend to produce mountain or hill waves (gravity waves) in a deep layer (*e.g.* HOLTON, 1979). The effects of hills have been studied recently both by observations and models (*e.g.* MASON and KING, 1984) but much more needs to be done. In this section an unhomogeneous surface was used in the

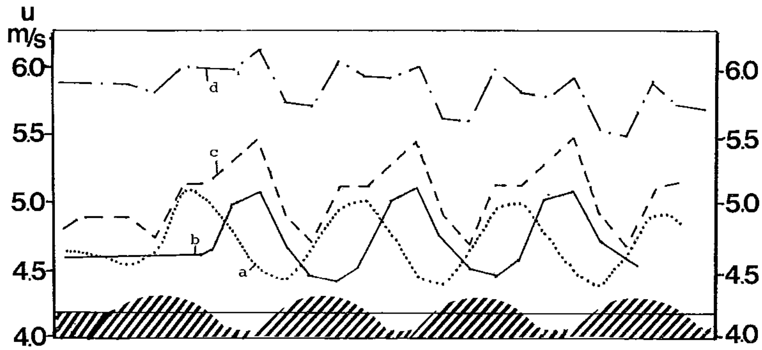

Fig. 7. The 10 m u wind component in the steady state of the model when 10 m/s geostrophic wind blows (from left to right) over an area filled with lakes and/or hills as indicated below.
a) hills only, no lakes
b) lakes only, flat topography
c) hills and lakes, stable stratification
d) as in c) but warm hills (22 °C) and cold lakes (10 °C).

present model to study the response in the large scale flow.

The model was applied to geostrophic flow over unhomogeneous terrain by introducing small hills (height 37 m, length 12 km) and/or lakes (4 km wide) in the valleys of the hills. The model gridlength was reduced to 2 km, so the subgrid scale features fall within the regime, which is presumably well represented by the model parameterizations while the hill waves and »lake breezes» are explicitly resolved.

Fig. 7a–c shows the 10 m steady state u-velocity in the grid points over the area filled with either hills or lakes or both, when a steady 10 m/s geostrophic wind is applied. There are no heat contrasts so the turbulence is mechanically driven in stably stratified boundary layer ($N = 10^{-2}\ s^{-1}$) as the wind is blowing over topography changes and surface roughness differences. Here z_0 is 10 cm over land (hills) and 1 mm over lakes.

Over the flat inflow land area the 10 m wind speed is 4.9 m/s, 20° left from the geostrophic 10 m/s wind and the 10 m u-velocity is 4.6 m/s. The presence of hills but no lakes (Fig. 7a) reduces the surface flow speed in valleys and on uphill slopes and increases it on hilltops and downhill slopes so that the area average does not change. In the case of lakes but no hills (Fig. 7b), the wind speed increases over lakes and is then slightly decelerated over land. On the average, the wind speed increases a bit. When the two effects are combined (Fig. 7c) the u-velocity increases more than just adding the two anomalies. The area averaged 10 m u-velocity is in this case about 0.5 m/s or 10 % stronger than without surface inhomogeneities. Much weaker variation is similarly found in the v-component. The variation in the surface wind disappears quickly upward, and could, therefore, be simulated by decreasing the effective roughness length over the lake and hilly areas in a coarser mesh model.

The vertical velocities that arise in the modelled flow over the unhomogeneous surface are small but logical: over the windward shores of each lake there develops weak rising motion ($w \sim 0.5$ cm/s) at 200 m level with sinking motion on the respective leeward shore while the inflow side of the whole lake district is characterized by sinking motion ($w \sim -0.7$ cm/s) at 1 km level, and the outflow region, by rising motion. These vertical motions are caused by the »coastal convergence» due to the decreased friction over each individual lake, and over the lake area as a whole.

When the land area is made warm (22 °C) compared to the lakes (10 °C) simulating early summer daytime conditions, the flow becomes unstable near the land surface but remains stable over lakes. The lake dimensions are, however, too small for proper lake breeze development. In Fig. 7d, simulation in such a case leads to stronger surface winds and the 10 m u-velocity is now around 6 m/s for u_g = 10 m/s with only small variation. The most interesting aspect in this experiment is that the u-velocity is slowly decreasing into the area of warm hills and cold lakes, indicating increasing roughness inside the lake district with temperature differences. This feature was not present in the stably stratified flow.

5. *Concluding remarks*

In this work, the given surface temperatures and roughnesses have caused circulations in a high resolution numerical model in a calm case. The simulated city breeze (over a warm city) was strongest in the evening as the consequence of unstable daytime stratification even if the (observed) temperature difference itself was largest overnight, but it continued throughout the day. The pure sea and land breezes were weakish in the late summer case chosen but the inclusion of a warm city zone on the coastline enhanced the sea breeze considerably, by getting it started almost immediately after sunrise, and locked it on the coastline above the warm city. The land breeze developed above the suppressed city breeze circulation and enhanced thereby slightly.

According to the results, the sea breeze could be expected to be stronger and more common over coastal city areas than over uninhabited coasts. The vertical motion and mixing, and, *e.g.,* the vertical advection and dispersion of pollutants is strong over the city even at nighttime.

These mesoscale circulations are often expected to smooth out under horizontal averaging over, say, 200 km. In the calm case this was so, but an offshore prevailing wind enlarged the dimensions and intensity of the modelled sea breeze to the extent that the 200 km average clearly showed sea breeze features, which should be parameterized in a coarse mesh model.

Finally, the sea area was transformed to a region of small lakes (4 km) and hills (12 km length) to study the area averaged flow in such a »lake district» case. In stable, mechanically driven flow without temperature differences, the main effect was a large decrease in the effective roughness of the area, and vertical motions analogous to coastal convergence, while warm hills and cold lakes induced an extra decrease which was most evident in the inflow area.

REFERENCES

ALESTALO, M., 1975: Esimerkkejä ns. lämpösaarekeilmiöstä Helsingin alueella. *Report No. 7, Department of Meteorology, University of Helsinki.* 15 pp.

ALESTALO, M. and H. SAVIJÄRVI, 1985: Mesoscale circulations in a hydrostatic model: coastal convergence and orographic lifting. *Tellus,* 37A, 156–162.

ALPERT, P., COHEN, A., DORON, E. and J. NEUMANN, 1982: A model simulation of the summer circulation from the Eastern Mediterranean past Lake Kinneret in the Jordan Valley. *Mon. Wea. Rev.,* 110, 994–1006.

ATKINSON, B.W., 1981: *Meso-scale Atmospheric Circulations.* Academic Press, London.

HOLTON, J.R., 1979: *An introduction to dynamic meteorology.* 2nd ed. Academic Press, London. 391 pp.

HSU, S.A., 1979: Mesoscale nocturnal jetlike winds within the planetary boundary layer over a flat, open coast. *Boundary Layer Met.,* 17, 485–494.

MASON, P.J. and J.C. KING, 1984: Atmospheric flow over a succession of nearly two-dimensional ridges and valleys. *Quart. J. Royal Met. Soc.,* **110**, 821–846.