

Mesoscale flow modification induced by land-lake surface temperature and roughness differences

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Abstract. Numerical model simulations and airborne measurements have been used to study the mesoscale flow modification of the boundary layer due to the presence of a lake (34 km²) located in the southern boreal zone. The mesoscale model used is a three-dimensional hydrostatic model with a “level 2.5” turbulence closure. The results from this study show that in general, differences in surface roughness are more important than differences in surface temperature for the flow modification. The dynamical effect due to the roughness difference between the lake and the surrounding land creates divergence over most of the lake because of the accelerating flow, while a convergence zone appears at the downwind shore because of the deceleration. This pattern is found during nighttime for all simulated cases with wind speeds from 4 to 17 m s⁻¹. This pattern persists during daytime for moderate to high wind speeds but is reduced in strength owing to stable stratification over the lake. For wind speeds ≤ 6 m s⁻¹ a lake breeze circulation develops during the afternoon. Complicated flow patterns occur when the forces due to friction and thermal pressure gradient are comparable in size. Available measurements qualitatively support the model results.

1. Introduction

Sea and land breezes in coastal areas are good examples of thermal circulations in the atmospheric boundary layer (ABL). Differences in surface temperatures, due to differences in albedo and in heat capacity between surfaces, create air temperature differences between adjacent regions. These temperature differences cause horizontal pressure gradients, which initiate circulations or modify the flow at different scales. The structure and physics of sea and land breezes have been widely studied [Atkinson, 1981]. One could expect that temperature differences between any surfaces should be able to create similar kinds of circulations. Segal and Arritt [1992] introduced the term nonclassical mesoscale circulations (NCMCs) for thermal circulations conceptually similar to sea breezes and used the name perturbed area (PA) for an area whose sensible heat flux differs significantly from the surrounding area and therefore could induce a thermal flow.

Xian and Pielke [1991] investigated the sea breeze development with respect to width of landmasses, geostrophic wind, thermodynamic instability, and latitude. As shown by them, and as will be shown here later, the resulting modification of the flow is highly sensitive to the horizontal dimension of the perturbed area and to the background flow. For a small PA O (kilometers), also other effects such as differences in surface roughness and convective eddies, will be of equal or larger importance for the flow modification than thermal effects associated with a circulation. According to Segal *et al.* [1997], even in the absence of a background flow a NCMC can vanish owing to the horizontal flow associated with convective eddies. They estimate that the width of the PA must be at least twice the depth of the ABL to create a circulation which is not masked by the convective eddies. Results from *Avisar and*

Chen [1993] show that a background wind along the mesoscale flow tends to reduce the thermal gradient while a background wind against the mesoscale flow strengthens the thermal gradient. The flow will also be modified due to surface roughness changes between the lake and its surrounding. This modification is larger for high wind speeds and will, of course, totally dominate any modification when differences in surface temperature are small. The effects on the ABL flow due to changing surface roughness conditions are discussed in the works of e.g., *Kaimal and Finnigan* [1994] and *Wright et al.* [1998].

The aim of this paper is to investigate the degree of mesoscale flow modification caused by a lake during different forcing, represented by horizontal temperature gradients and background flow. It will be shown that for a small lake, as in this study, the difference in roughness between the lake and its surrounding will dominate over corresponding differences in temperature for most cases. We will concentrate on modification of mean conditions and will not consider details in fluxes or aggregation problems. A mesoscale numerical model is used to simulate the flow in connection to the lake. The results from the simulations are compared to in situ airborne measurements, which were performed within the framework of Northern Hemisphere Land Surface Climate Processes Experiment (NOPEX) [Hallidin *et al.*, 1999b].

2. Area Description

Lake Tämnen, which is selected for this study, is located in the northern part of the NOPEX area, 35 m above sea level. Lake Tämnen is 34 km², and its length (southwest/northeast) and width (northwest/southeast) are ~ 7.5 and 4 km, respectively (see Figure 1b). The lake mean depth is only 1.2 m, which means that the lake's response in water temperature to changing radiation conditions is quite fast compared to other lakes with similar size. However, compared to the surrounding land area, it still has a slow response for the timescales considered in this investigation. Figures 1a and 1b show the terrain

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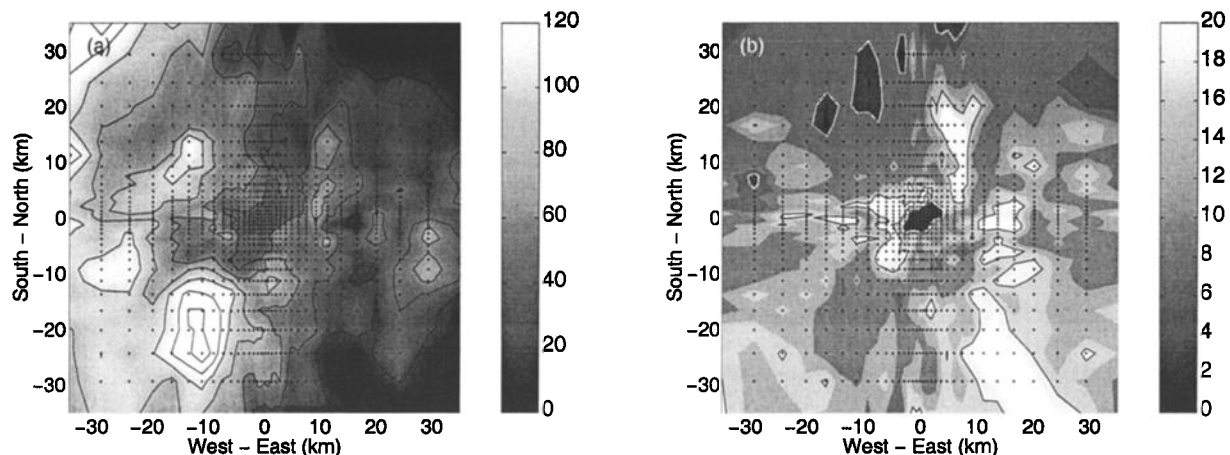


Figure 1. Contour plots of (a) the model terrain and (b) the land use, using surface albedo as reference. The surface albedo is calculated assuming 5, 10, and 20% albedo for the lake, forest, and agricultural areas, respectively. Both plots also include the location of model grid points. Note that the whole model domain is not shown. Lake Tämnaaren is shown by the black area in the center of Figure 1b.

and albedo variations in the NOPEX area as used in the model; here Lake Tämnaaren is located in the middle of the domain where the model resolution is most dense. The terrain and land use descriptions are based on digitized information with 50 m resolution, averaged to fit with the model grid. The NOPEX area is generally flat with a maximum slope in the order of 10 m per kilometer based on the model terrain (see Figure 1a). Thus, orographic effects should be very small as is also shown in the results.

The land use data have 16 classes of surface types, but in this study these are simplified to three main types: Agricultural land (including populated areas), forest (including all forest types, although dominated by conifers), and water surfaces (including lakes, rivers, and wetlands). Figure 1b shows calculated surface albedo in the domain based on this land use information and the albedo values specified in the figure caption. The land use types can be identified from the albedo and shows that the area immediately surrounding the lake is dominated by forest and agricultural surfaces, covering 63 and 32%, respectively, of the area within a ± 10 km square around the lake. The lake, with its low albedo, is clearly seen as the dark area in the center of the domain. The white areas represent agricultural surfaces, and the gray areas forest surfaces. The different gray scales indicate the dominance of either of these surface types.

3. Measurements and Data

In situ airborne meteorological measurements were performed in the ABL with a twin-jet Sabreliner 40A. The aircraft was used on rental basis from the Testing Directorate of the Swedish Defence Material Administration. The three wind components were calculated from pressure fluctuations measured by a radome gust probe [Brown *et al.*, 1983]. The absolute errors of the horizontal wind speed components are ± 0.5 m s⁻¹ each [Samuelsson and Tjernström, 1999a]. Temperature was measured by a Rosemount 102E2AL total temperature sensor and humidity by a dew/frost point hygrometer and a Lyman α hygrometer. Upward and downward short and long wave radiation were measured with instruments mounted on the aircraft fuselage as well as surface temperature measure-

ments with an infrared radiometer. The sampling rate was 6.1–48.8 Hz, depending on required resolution, and the true air speed was 100 m s⁻¹. For more information on airborne measuring techniques, instrumentation, and data, refer to Tjernström and Friehe [1991], Tjernström and Samuelsson [1995], Samuelsson and Tjernström [1999a], and Tjernström and Samuelsson [1999].

The flight plan used for the analysis in this paper is one of the so-called “turbulence flight” patterns, flown during the NOPEX campaigns, which passes over Lake Tämnaaren as seen in Figure 2. This track passes over several surface stations that are not used in this particular study. The other turbulence flight track over Lake Tämnaaren across Norunda, Marsta, and Uppsala was not available, owing to air traffic control considerations. The pattern consists of eight horizontal legs, ~ 60 km in length, flown at three heights in the ABL. The lowest permitted height of 100 m is for some cases on the limit of being

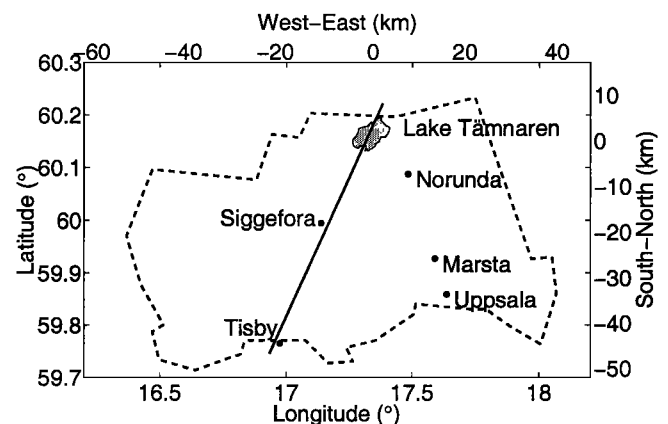


Figure 2. The NOPEX region, shown by its border (dashed line), with the city of Uppsala located in southeast. Lake Tämnaaren is the shaded area in the northern part of the region. Surface measurements were performed at Lake Tämnaaren, at the forest sites in Norunda and Siggefora, and at the agricultural sites in Marsta and Tisby. The airborne measurements were performed along the track indicated by the solid line.

Table 1. Five Cases With Airborne Measurements Used in the Analysis

Date	Time, LST	U/ϕ , $\text{m s}^{-1}/\text{deg}$	$ \psi - \phi $, deg	ΔT_s , $^{\circ}\text{C}$	$\Delta \theta$, $^{\circ}\text{C}$	H_{lak} , W m^{-2}	H_{for} , W m^{-2}	H_{agr} , W m^{-2}	u_* , m s^{-1}	z_i , m	z_s , $L^{-1} \text{m}$
June 13, 1994	1020	5.0/335	50	3.5	0.34	0	325	95	0.60	1250	-6.3
June 13, 1994	1130	4.7/320	65	6.7	0.27	-5	340	90	0.61	1550	-12.4
June 21, 1994	1330	8.7/195	10	6.5	0.10	-15	380	115	0.63	1300	-5.4
May 8, 1995	1220	6.6/20	5	0.6	0.13	30	130	40	0.54	1100	-6.7
June 20, 1995	1230	3.5/280	75	5.9	0.47	0	305	80	0.50	1100	-12.3

Each case represents mean values of two sequential flight legs at the height of 100 m for the mean time indicated as Local Standard Time (LST). The wind speed U and the wind direction ϕ are over-land mean values. The $|\psi - \phi|$ gives the angle between ϕ and flight-track heading, ψ . ΔT_s and $\Delta \theta$ are surface temperature and potential temperature (at 100 m) differences between land and lake, respectively. H_{xxx} represents surface measured values of sensible heat fluxes at Lake Tämna (lak), Norunda (for), and Marsta (agr). The u_* , z_i , and $z_s L^{-1}$ are the friction velocity, the depth of the ABL, and the stability z_i over Monin-Obukhov length, respectively, all estimated from airborne measurements.

too high to detect any effect of flow modifications by the lake for this study. Anyway, as will be seen in the results, it is still possible to detect horizontal differences, due to the presence of the lake, at this level. Each flight mission consisted of four legs at 100 m, three legs around 200 m, and one leg at or above 300 m ($\sim 0.6z_i$, where z_i is the depth of the ABL determined in situ from aircraft profiles during each flight).

In addition to the airborne measurements, data from three land surface sites have been used. These are Lake Tämna, Norunda, and Marsta (see Figure 2). Profile data of temperature and humidity have been used from all these sites for validation of the mesoscale model. Radiosonde data and pibal tracking data from soundings performed at Marsta were also used for this purpose. Sensible heat flux data and some profile data have been used for lake breeze estimations in section 4.

The data used from Lake Tämna were profile data measured at a tower located in the southern part of the lake and turbulence data measured on a float located ~ 500 m south of the tower. The tower fetch varied from 1.5 to 5 km, depending on wind direction. The profile in the tower included wind speed, temperature, and absolute humidity variations measured at 10 Hz at four heights (0.75–7.6 m). Turbulence measurements on the float were performed with an eddy correlation instrument (Applied Technologies) mounted 3 m above the lake surface. When no eddy correlation measurements are available we have used profile data from the tower to calculate turbulent fluxes using the gradient method based on the Monin-Obukhov similarity theory. For more information concerning measurements performed at Lake Tämna, please refer to Venäläinen *et al.* [1998] and Heikinheimo *et al.* [1999].

In Norunda, which is a forest site, a 100-m tower is equipped for both eddy correlation measurements (Solent ultrasonic anemometers in combination with LICOR gas analyzers at three levels) and profile measurements (wind anemometers and thermocouples at 12 levels). The forest around the tower is relatively homogeneous within a fetch of 1–2 km and consists of Norway spruce and Scots pine, which are also the dominating conifer forest types in the whole NOPEX region. See Grelle and Lindroth [1994, 1999], Mölder *et al.* [1999a, 1999b], and Lundin *et al.* [1999] for more information on Norunda tower measurements.

Marsta is an agricultural site where a 30-m tower was instrumented for profile measurements (Pt-500 sensors for temperature and humidity and cup anemometers for wind at six levels) and with an eddy correlation instrument at 10 m (Ultrasonic anemometer). The agricultural fetch varies from 700 m to several kilometers depending on wind direction [Hall-din *et al.*, 1999a; Bergström, 1999; Smedman *et al.*, 1999].

The representativeness of the measurements performed at these sites for similar vegetation types in the rest of the NOPEX area of course depend on surface conditions (i.e., albedo and wetness) and on atmospheric conditions (horizontal homogeneity with respect to cloudiness and other synoptic conditions). For this study, sensible heat flux measurements from Marsta and Norunda are used as representative values for the agricultural and forest areas around Lake Tämna. In the work of Samuelsson and Tjernström [1999b] it is shown that the representativeness of measurements is an important issue when quantitatively comparing results from different stations. Such direct comparisons are not performed here which means that we do not have to take into account all aspects related to the representativeness of measurements and model results. Problems related to the representativeness of surface heat flux measurements performed in the NOPEX area are discussed by Samuelsson and Tjernström [1999b].

For this work, five flight missions on different days have been used in the analysis. Table 1 summarizes some parameters important for these missions. The ABL conditions for June 13, 1994, are described in detail by Samuelsson and Tjernström [1999a, 1999b]. Therefore this day has been chosen for validation of the model simulations in section 5.2.

4. Theoretical Flow Aspects

Considering some conceptual models, the degree of flow modification caused by the lake can be examined as a function of atmospheric conditions. A lake breeze (or any NCMC) is initiated by a horizontal temperature gradient, which creates a potential energy difference between the lake and its surrounding. The lowest lake breeze wind speed must at least be larger than the horizontal flow associated with convective ABL large eddies. Segal *et al.* [1997] concluded that the smallest characteristic lake width required for this to be true must be larger than twice the boundary layer depth. However, even if the lake is large enough to overcome the convective velocity scale, any thermal circulation can still be suppressed by a strong enough background wind. The smallest size of a PA which is on the limit to create a closed circulation for a certain background wind speed has been estimated by Segal and Arritt [1992] and Doran *et al.* [1995]. Their approach was to estimate the length (L), from the PA upwind border, required for the background wind speed (u_b) to decrease to zero when the flow works against a thermally induced mesoscale pressure gradient. In the derivations they differ in some of their assumptions; Segal and Arritt use the geostrophic wind speed and consider a boundary layer developing during cold air advection, while

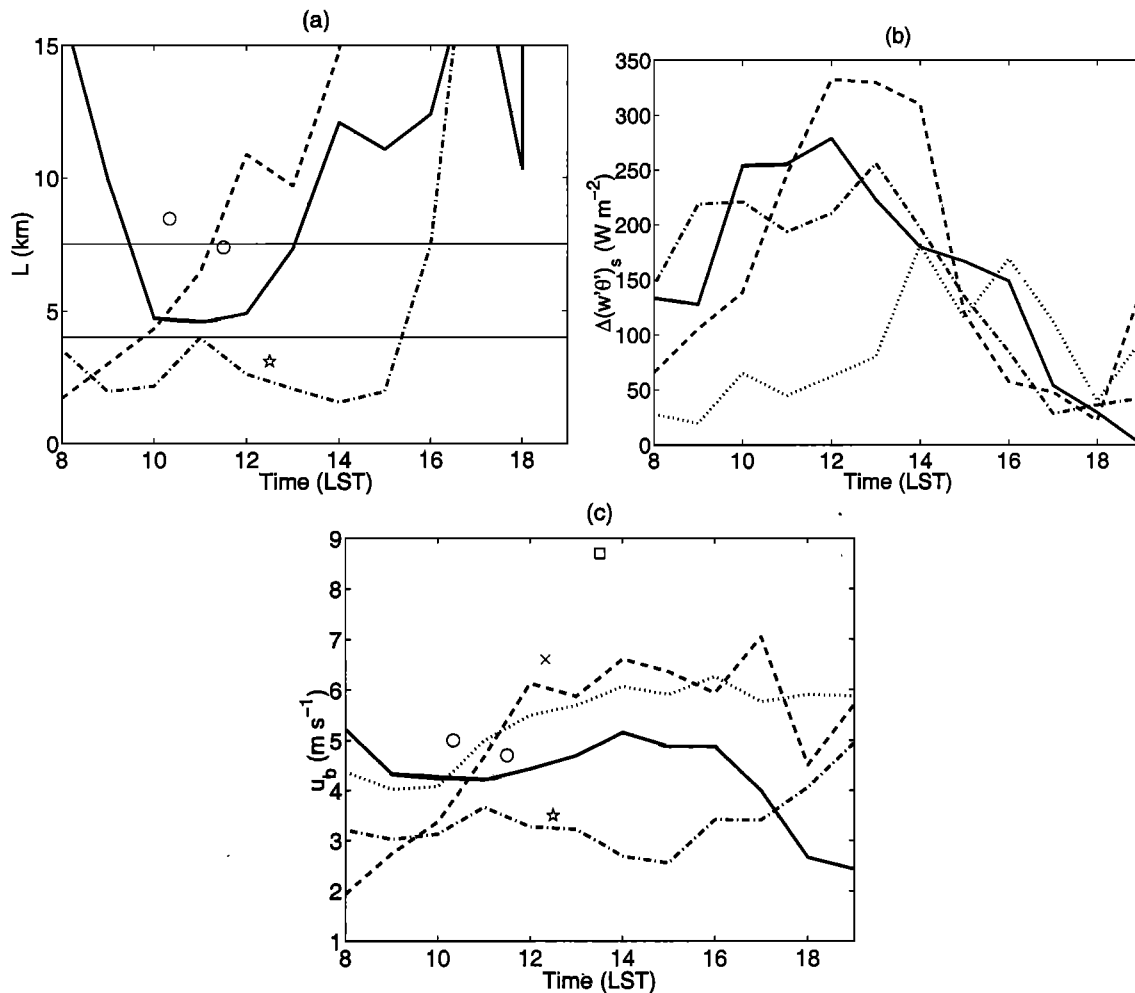


Figure 3. (a) Minimum lake width, L , (b) difference in $(\overline{w'\theta'})_s$ between the lake and its surrounding, and (c) background wind speed u_b . Thick lines represent surface measurements and markers represent airborne measurements for June 13, 1994 (solid and circle), June 21, 1994 (dashed and square), May 8, 1995 (dotted and cross), and June 20, 1995 (dash-dotted and star). Horizontal lines in (Figure 3a) represent length and width of Lake Tämnaaren, respectively.

Doran et al. use the ambient wind speed and consider a boundary layer developing during warm air advection. Thus their resulting estimations are somewhat different but both can be shown to follow

$$L \propto \max \left(\frac{u_b^3}{\Delta(\overline{w'\theta'})_s}, 2z_i \right), \quad (1)$$

where the limitation with respect to boundary layer depth is included. Here $\Delta(\overline{w'\theta'})_s$ represents the difference in sensible heat flux between the PA and its surrounding. The assumptions made by Doran et al. [1995] compare best with the conditions for the present study. Their resulting expression for L is

$$L = \frac{\bar{\theta}}{g\Delta(\overline{w'\theta'})_s} \frac{u_b^3}{4\ln(2)}, \quad (2)$$

where $\bar{\theta}$ is the average potential temperature in the ABL and g is the acceleration of gravity. Doran et al. [1995] also performed numerical computations to determine the values of L and $\Delta(\overline{w'\theta'})_s$ for when the surface wind speed is brought to zero. These computations resulted in L values that were uni-

formly $\sim 25\%$ larger than those estimated from (2). Therefore, when applied to measurements, the estimated values are here increased with this factor.

In a study by Sun et al. [1997] of a lake breeze circulation in the Boreal Ecosystem-Atmosphere Study (BOREAS) region it was found that with weak ambient wind (1 m s^{-1}), the exact temperature difference between the lake and the surrounding land, was not the most important factor but rather the contrast between the development of the internal boundary layers over lake and land. It was also found that for a lake above a certain size, the divergence of the air above the lake decreases with increasing lake size.

Applying (2) to measured data of sensible heat flux, temperature, and wind speed during a day, the minimum PA size required to initiate a closed circulation can be calculated as a function of time. Note that the conditions close to the surface are considered in this case. Figure 3a shows the resulting L values as functions of time for the 4 days listed in Table 1. Here L refers to the “required size” of Lake Tämnaaren. For the calculations of L the wind speed at 10-m height and the potential temperature at 29-m height at the Marsta tower are

used as values for u_b and $\bar{\theta}$, respectively. For the difference in $(\overline{w'\theta'})_s$, between the lake and its surrounding, measurements of sensible heat fluxes at Marsta, Norunda, and Lake Täm-naren have been used. The lake-surrounding sensible heat flux has been estimated from Marsta and Norunda by weighting the flux measurements according to the fractions of vegetation around the lake represented by these sites as given in section 2.

Since $L \propto u_b^3$, the results are very sensitive to the wind speed. In the derivation of (2), u_b is assumed to be the well-mixed boundary layer wind speed, which is probably best described by pibal tracking data. However, to get a picture of the time dependence of L for most of the day, we chose to use tower data since the pibal tracking data are not frequent enough for that purpose. To show the dependence on u_b , the wind speed measurements from the aircraft have been used in the estimation of L , indicating also the variability in u_b . Results based on aircraft measured wind speed (see Table 1) are represented by the markers in Figures 3a and 3c. The horizontal lines in Figure 3a represent the length and width of Lake Täm-naren. For a circulation to develop, the minimum required size estimated has to be below the lower line or at least between the lines.

There are 2 days where surface and airborne measurements both indicate that the lake may be large enough, with respect to atmospheric conditions, to make it possible for a closed, or nearly closed, circulation to develop. For both these days the boundary layer depth is less than half the lake width which means that z , should not be a limiting factor for the development of a circulation. For June 20, 1995, the minimum required lake width is less than the width of Lake Täm-naren, considering both airborne and surface wind measurements. According to surface measurements, the conditions are favorable for a circulation to exist from morning hours until 1500 hours. The abrupt change at 1500 hours is due to the combination of an already decreasing sensible heat flux difference and a suddenly increased wind speed as seen from Figures 3b and 3c. On June 13, 1994, the period of conditions favorable for a circulation to exist, is just a couple of hours. The possible existence is also uncertain since the minimum lake size required is in between, and for the first two aircraft legs even above, the lake width and length.

For the two other days, June 21, 1994, and May 8, 1995, the combination of high wind speed and/or small difference in sensible heat flux will prevent any circulation to develop. Note that for June 21 the heat flux difference is larger than for any other day but that is counteracted by the wind speed later during the day which becomes too high to make it possible for any circulation to persist. For May 8, 1995, the minimum lake size required for a circulation to develop under these conditions is in the interval 20–40 km (not shown). Thus the only effect of the lake in this case should be a perturbation in the ambient flow.

In conclusion, several aspects of the boundary layer flow must be fulfilled in order for a NCMC to develop. In the present case the relatively small size of the lake is a limiting factor and fully developed NCMCs are expected to be rare, but even for a larger lake the winds have to be sufficiently weak while the heat flux difference must be large. Still, even a small lake will always have an influence on the flow although the cases for which it will trigger a NCMC are likely to be relatively rare. The effect of the lake for most cases will thus be a perturbation in the ambient flow.

5. Model Simulations

The main idea is to attempt to simulate as much as possible the observed flow with as simple forcing and background flow conditions as possible, while retaining a high degree of control of the forcing. The purpose is to investigate the flow response of heterogeneous surface conditions on the mesoscale and not to validate the behavior of any surface energy balance schemes for different conditions or resolutions. Furthermore, one aim is to maintain the same surface forcing for different background flow conditions, in different simulations. Unlike, e.g., *Vidale et al.* [1997], the details of the energy balance at the surface were thus not simulated.

The model results rests on the fact that a control simulation can be made to agree reasonably well with observations. The effect of changes in the background flow or in the surface forcing can then be studied through sensitivity simulations. Some of the modeling techniques applied here are motivated by this aim [*Tjernström and Grisogono*, 1996, 2000; *Grisogono and Tjernström*, 1996; *Grisogono et al.*, 1998; *Cui et al.*, 1998; *Tjernström*, 1999].

5.1. Model

The model applied, the MIUU meso- γ -scale model [*Tjernström*, 1987, 1988; *Enger*, 1990a], is a three-dimensional hydrostatic model with a higher-order turbulence closure. The vertical coordinate in this model is transformed into a terrain-influenced coordinate system [*Pielke*, 1984]. The turbulence closure is an improved, consistent version of the “level 2.5” closure [*Mellor and Yamada*, 1982]. The closure carries an improved description for the pressure redistribution terms (the “near-wall” correction) and an algorithm to keep all second-order moments realizable [*Andrén*, 1990]. The model also includes routines for subgrid scale condensation, radiation, and for the surface energy balance; however, the latter is not used here for the sake of simplicity.

The model has previously been applied to a variety of applications, including terrain induced flows [e.g., *Enger et al.*, 1993; *Koracin and Enger*, 1994; *Grisogono*, 1995; *Enger and Grisogono*, 1998], coastal flows [e.g., *Cui et al.*, 1998; *Tjernström and Grisogono*, 2000; *Tjernström*, 1999], dispersion calculations [e.g., *Enger*, 1990b], marine stratocumulus [*Tjernström and Koracin*, 1995], and air chemistry [e.g., *Svensson*, 1998; *Svensson and Klemm*, 1998]. It has thus been thoroughly examined for a variety of flows.

The horizontal grid shown in Figure 1, expands toward the lateral boundaries, to achieve maximum resolution in the central parts of the domain, while locating the lateral boundaries far from the area of interest. A simple radiative boundary condition is applied at the lateral boundaries. The vertical grid also expands, log-linearly, toward the model top. The maximum resolution is located close to the surface at the domain center, located in the middle of Lake Täm-naren. The total domain is $150 \times 150 \times 4 \text{ km}^3$ and is resolved by $41 \times 41 \times 30$ grid points, with a maximum resolution of $600 \times 600 \text{ m}^2$ in the horizontal, and 1 m in the vertical (close to the surface) (see Figure 1).

The use of the hydrostatic assumption with this horizontal resolution is undoubtedly on the limit. *Pielke* [1984, pp. 33 and 87] performed a scale analysis and concluded that for this assumption to be valid $L_x/L_y \gg 1$, where L_x is the horizontal and L_y is the vertical scale of the flow, respectively. Here the horizontal wavelength should be of the order of the lake width,

Table 2. Specification of Surface Temperature and Humidity Variables Used as Surface Boundary Conditions

Land Use	T_0 , °C	ΔT_a , °C	δ , hours	h_0 , %	h_D , %
Forest	15.5	8.0	2.0	20	25
Agricultural	19.0	8.0	1.5	5	10
Lake	14.0	1.2	3.0	100	...

See the text for an explanation.

$L_x \sim 4\text{--}7.5$ km. Taking the maximum vertical wavelength to be the maximum ABL depth, $L_y < 2$ km, gives a ratio $L_x/L_y > 2$ which is indeed marginal. On the other hand, this analysis is based on the absolutely largest observed ABL depth and for much of the day, the ratio is at least twice as large. When the airflow encounters the different surface characteristics over the lake, a circulation may be triggered or an internal boundary layer is formed. In either case, the depth of the simulated flow perturbation, using the hydrostatic assumption, is < 500 m, giving $L_x/L_y \sim 10$. For nocturnal conditions, with an even more shallow active layer, the hydrostatic assumption will be well satisfied. A case when the assumption really fails is related to the sea breeze front for which L_x/L_y can be well below one. In this situation we cannot trust the simulated strength of, i.e., vertical wind speed in the front, but the model should still be able to reproduce the patterns realistically. Admittedly, one can, in general, not judge the appropriate scale for a possibly non-hydrostatic circulation from hydrostatic simulations. The only safe test would be parallel runs with a nonhydrostatic model, but these numbers may still provide some guidance. To keep the study simple, the hydrostatic assumption is retained, which must be kept in mind while analyzing the results.

The energy balance at the surface is not simulated. That means that we are not solving an equation for the surface temperature time tendency, $\partial T_g/\partial t$. Instead, T_g is prescribed for each of the three land use classes, respectively. This is given as $T_g = T_0 - \Delta T_a \cos[2\pi(\tau - \delta)/24]$, where T_0 is the average and ΔT_a is the diurnal range of the surface temperature, τ is the local time, and δ is a time lag. The final surface temperature for each grid point is the linearly area-weighted average, using the fraction of the different land uses. Similarly, the surface saturation humidity for each land use class is calculated from its temperature, and the surface humidity is then calculated using a prescribed fraction of potential evaporation, different for each land use class. The fraction of potential evaporation is composed by a constant background value h_0 and a daytime addition h_D to account for transpiration. The maximum value of a function proportional to the incoming solar radiation is h_D . The final, land use weighted, surface temperatures and humidities are transferred to the lowest model grid points, forcing the profiles to obey Monin-Obukhov similarity from above and Zilitinkevich roughness sublayer similarity [e.g., Pielke, 1984, p. 152] from below. The parameters to specify (T_0 , ΔT_a , and δ for the temperature and h_0 and h_D for humidity) are tuned, by trial and error, in a control simulation, using tower measurements from Marsta, Norunda, and Lake Tämnen. This process, while not ensuring the correct observed fluxes, will provide a model atmosphere with the approximate characteristics of the day in question. The parameter values used are shown in Table 2. It should be underlined that the surface temperature T_g does not necessarily have to correspond to the measured soil or skin temperature. In this context it is just the value that provided to the model will give

the correct low-level air temperature with the present formulation and similarly with the fraction of potential evaporation. The simulated fluxes are based on the differences between the grid-averaged surface values and the values at the lowest atmospheric level corresponding to the parameters relevant for each flux.

The background (synoptic scale) pressure gradient was specified as a geostrophic wind, for simplicity constant in time and in the horizontal. The geostrophic wind was estimated from pibal tracking performed at Marsta. Initial potential temperature and specific humidity profiles were estimated from radiosoundings at Marsta. The simulations were initialized using a dynamic initialization; the model is given horizontally homogeneous temperature, humidity and wind fields and run through a pre-integration period, during which the model fields adjust gradually to a realistic quasi-balance. All simulations presented here were initialized at 1800 local standard time (LST) the day before the event, and no data for the first 10 hours are used in the analysis.

5.2. Control Simulation

For the control simulation (CONT) the day of June 13, 1994, was chosen. The respective components of the geostrophic wind in the run (u_g , v_g) change linearly from (1, -6) m s⁻¹ at the surface to (10, -25) m s⁻¹ at 3500 m and constant above, according to the Marsta pibal tracking. Figures 4–6 provide a model verification for this day. Figure 4 shows the simulated temperatures and humidities from the calibrated control run, along with the measurements from Marsta, Norunda, and Lake Tämnen. Figure 4 shows all the levels in the towers (markers) and the corresponding model levels (solid lines) for the whole day. Between ~0400 and 2200 LST, both temperature and humidity seem to agree with measurements at Marsta and Norunda. Before 0400 LST, the initialization seems not to be completed and after 2200, the synoptic scale flow starts changing, invalidating the assumption of a constant background flow. The temperature at Lake Tämnen has the correct diurnal cycle, but is ~1°C too low, while the relative humidity (specific humidity was not measured here) is ~5% too high. The reason for this discrepancy is not known, although the oversimplified boundary conditions may certainly be one explanation. Increasing the specified diurnal average surface temperature by 1°C provides a more accurate temperature but an even larger deficit in the humidity. More seriously, however, the sensible heat flux then remains positive throughout the day. For the purpose of this study, and comparing model fluxes at Tämnen to measurements, the setup in Figure 4 was retained as a control run.

Figure 5 shows that the simulated wind profiles compare quite well to those from pibal tracking during the day (from 0500 to 1600 LST). Figure 6 shows time-height cross sections of potential temperature and relative humidity from radiosoundings and model results. The simulated ABL growth is slightly slow during the first hours, but reaches the correct height. The heating of the ABL between 0800 and 1200 LST is more rapid in the measurements, while the simulated ABL is too moist. The measurements also show an increase of humidity aloft with time, that can never be achieved with this model without nesting from a larger scale; This is intentionally omitted in the simulations. In summary, although differences occur between the measurements and the simulation, the simulated atmosphere is similar to observations. For the purpose of this study, which is to investigate the perturbation by the lake on

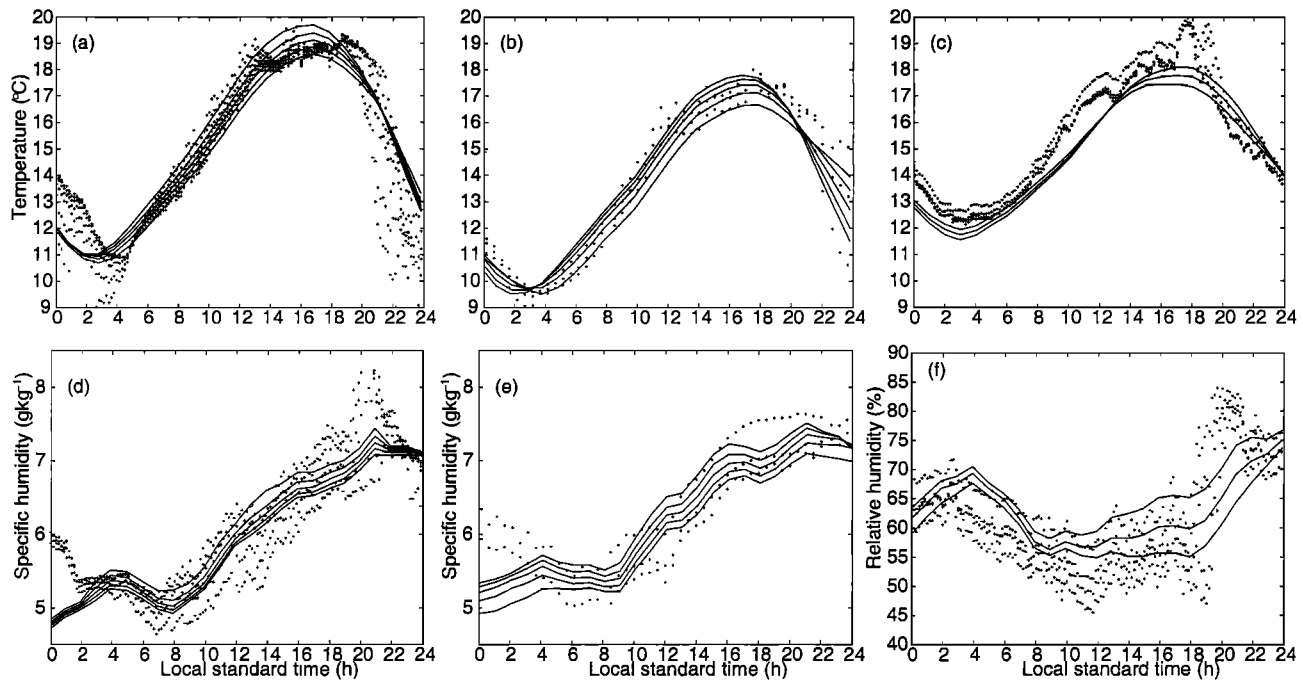


Figure 4. Simulated and measured (a–c) temperature and (d–f) humidity for (a, d) Marsta, (b, e) Norunda, and (d, f) Lake Tännaren. The markers show all the levels at each tower, while the solid lines indicate all the corresponding model levels.

the local flow, this control run is considered accurate enough to serve as a starting point for the sensitivity study.

5.3. Sensitivity Simulations

In all, nine sensitivity simulations were performed (see Table 3). The first sensitivity run (LONG) consists of extending the control simulation into the next day, using the same forcing. The bulk of the ABL is heated during the first day, and when the surface temperature falls to the same level the second night as during the first night, the residual mixed layer remains warm. Thus the first half of the second day will be significantly more stable and then the ABL will

initially grow more slowly. This chain of events was actually observed during June 14, 1994, however, in reality the wind also turned and increased [Samuelsson and Tjernström, 1999a], so that the second day of this simulation is not an exact proxy for the observed conditions. The next six sensitivity simulations were set up with simplified (constant) geostrophic winds: Three different geostrophic wind speed were used (4 (low), 8 (medium), and 17 (high) m s^{-1}), in combination with wind directions from northwest, across the lake main axes, (NWLow, NWMod, NWHi), and from northeast (NELow, NEMed, NEHi). One more simulation (COOL) was run as the control simulation, but with a lake

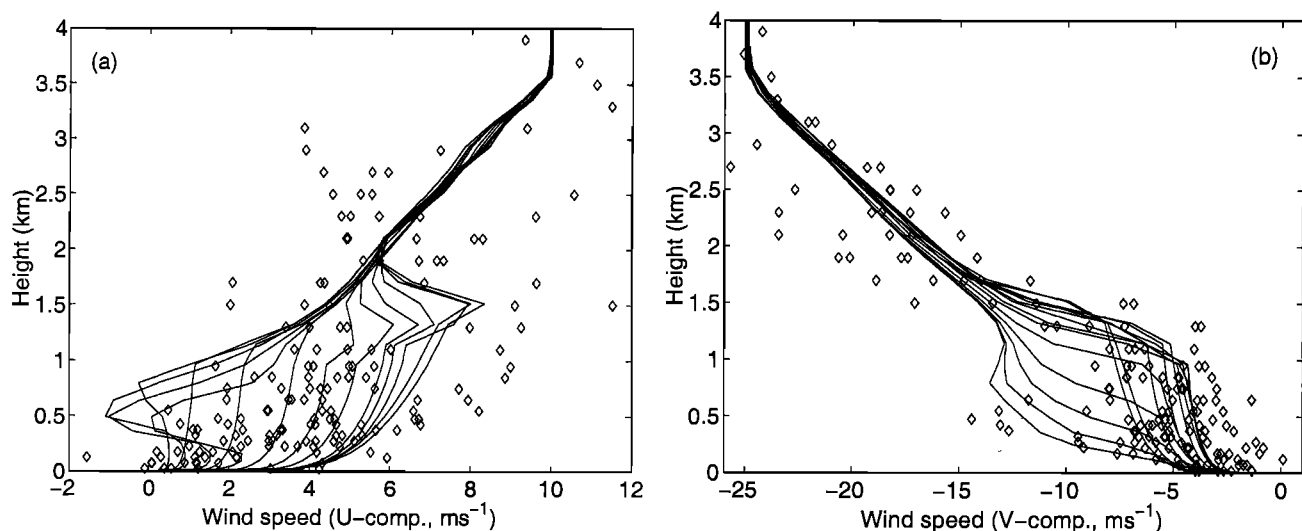


Figure 5. Observed (diamonds, pibal tracking from Marsta) and simulated (solid lines, every hour) wind speed for (a) the east-west and (b) the north-south components between 0500 and 1600 LST.

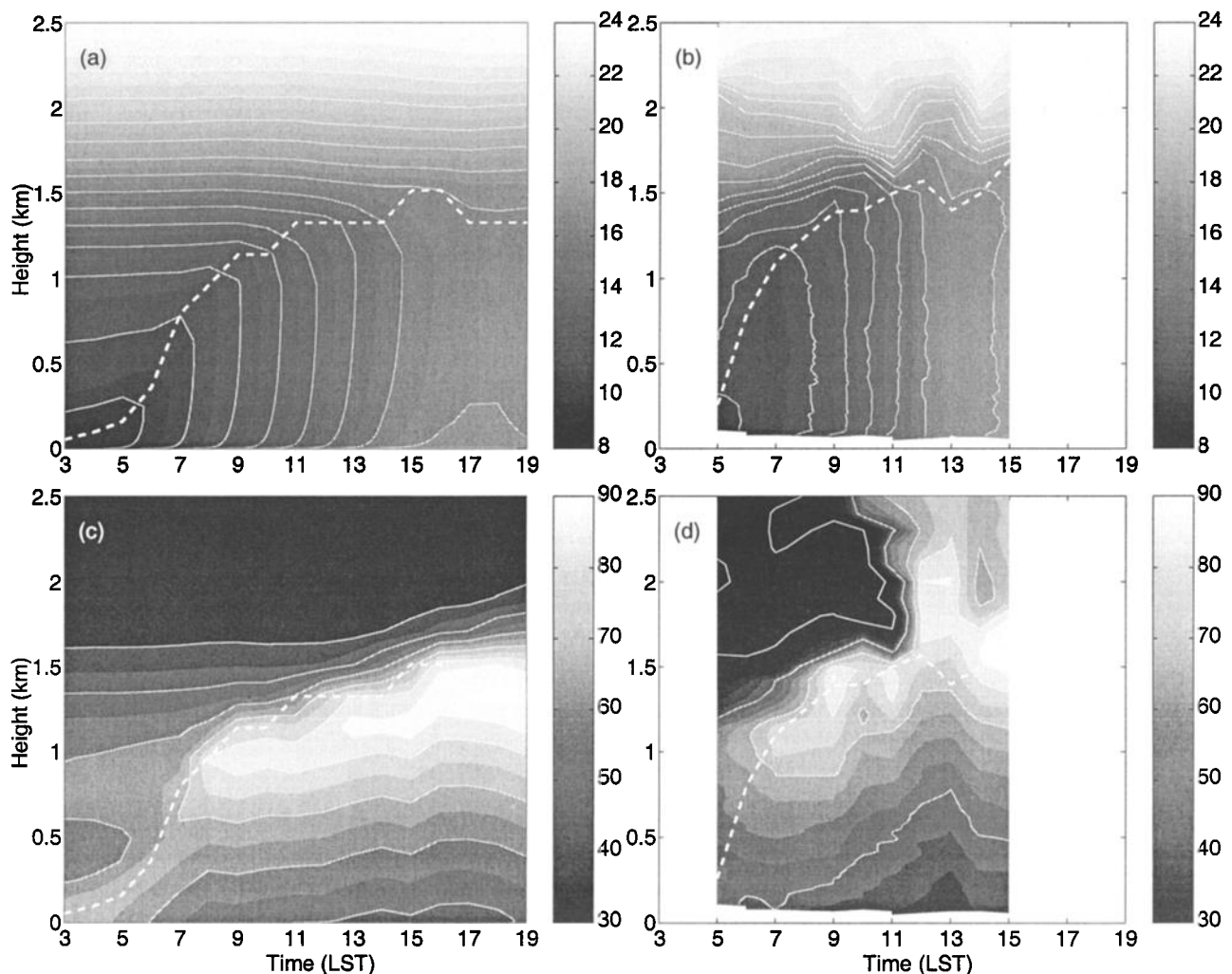


Figure 6. Time-height contour plots of (a–b) potential temperature ($^{\circ}\text{C}$) and (c–d) relative humidity (percent) from the simulations (a, c) and from the measurements with radio-soundings (b, d) at Marsta. The dashed line is the ABL depth, defined as the base of the capping inversion.

surface that was 2°C cooler throughout. The last simulation (NWInt) was run with northwesterly flow at 6 m s^{-1} (in between NWLow and NWMed).

6. Model Results

Figure 7 shows the time-height cross sections of the temperature difference between the lake and the surrounding area for three runs (CONT, NWLow, and NWHi). Here the lake is identified as grid points with $>80\%$ lake fraction, while the surrounding area is defined as all points with a lake fraction $<20\%$, both within a $\pm 10\text{ km}$ square centered around Lake Tämnen. The temperature difference shown is that between the horizontal-average vertical profile from each area, as a function of time. For the control run (Figure 7a) it is evident that the average perturbation is mostly confined below 100 m , while the low-level difference is 2°C (CONT, NWMed, and NWHi are very similar). Figure 7b shows the same for the NWLow run. It is clear that during daytime with weaker winds, the perturbation reaches significantly higher, to $\sim 4\text{--}500\text{ m}$.

The average, minimum, and maximum divergence at all heights above the lake were also estimated for all the runs, and some results are shown in Figure 8. Although, there were slight

differences between all the runs with moderate to high winds across the main axis of the lake (CONT, LONG, NWMed, NWHi, and COOL), they all showed a consistent pattern. Figure 8a shows the composite average of all these runs (similarly, the runs NEMed and NEHi (not shown) were also similar but different to the runs with north-westerly wind due to

Table 3. List of Definitions for the Model Simulations

Name	Wind Speed, m s^{-1}	Wind Direction	Other
CONT ^a			
NWLow	4	northwest	...
NWMed	8	northwest	...
NWHi	17	northwest	...
NELow	4	northeast	...
NEMed	8	northeast	...
NEHi	17	northeast	...
LONG	as in CONT	as in CONT	extended to 48 hours
COOL	8	northwest	lake 2°C cooler
NWInt	6	northwest	...

^aCONT, control simulation. See the text for a specification of the control simulation.

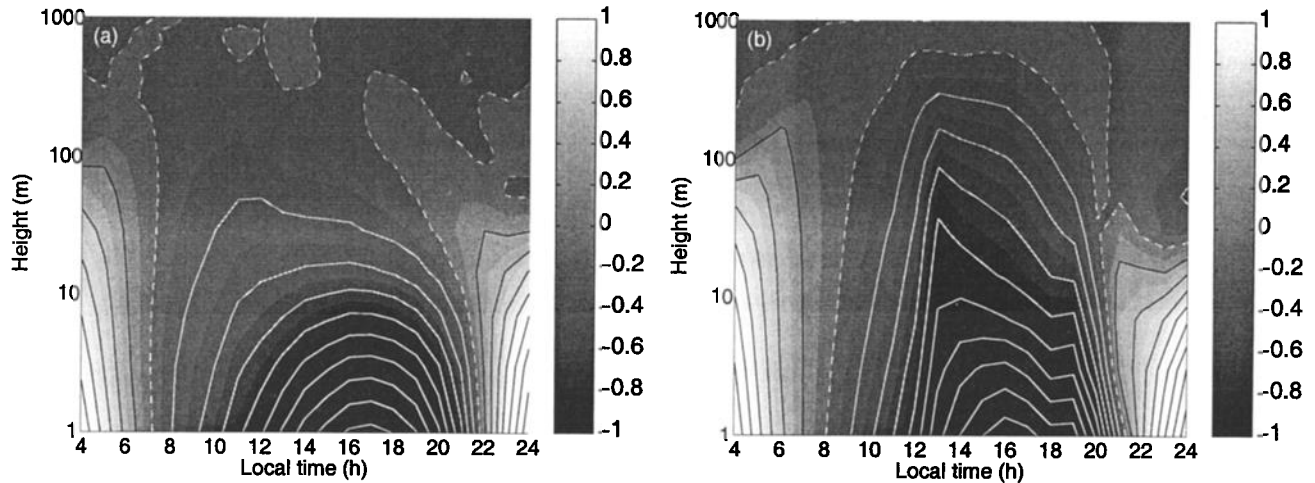


Figure 7. Linear-logarithmic time-height cross section of the mean temperature difference ($^{\circ}\text{C}$) between the air over Lake Tämnaaren and the air in the surroundings for (a) the control run and (b) the sensitivity run NWLow.

the lake geometry). Surprisingly, these runs show on average low-level divergence over the lake during nocturnal conditions (when the lake is warmer than the surroundings) and low-level convergence over the lake during the day, when the lake is cooler than land. This is a more complex pattern than that displayed in the works of, e.g., *Sun et al.* [1997] or *Vidale et al.* [1997]. However, the result for a weak wind case (NWLow, Figure 8b) shows the expected feature, strong daytime divergence over the lake, as would be expected for a lake breeze development. This pattern is quite consistent with that observed [*Sun et al.*, 1997] and simulated [*Vidale et al.*, 1997] over Candle Lake in the BOREAS region for a weak wind case, with a minimum around 1000 LST and a maximum midafternoon [*Vidale et al.*, 1997, Figure 4].

Figure 9 shows the horizontal distribution of the divergence at 15 m over the local surface for the control run. The divergence pattern at nighttime is signified by a large maximum at the center of the lake and a corresponding minimum at the downstream shore. Daytime conditions show a similar pattern but much weaker. The average structure in Figure 8a is the

average result of this dipole. This type of pattern was also observed by *Vidale et al.* [1997, 29,182] for the nighttime case over Candle lake (see their Figure 10), although they analyzed the vertical wind, rather than divergence, and attributed the ascending winds at the downwind shore to be part of a land breeze being “shifted eastward by the prevailing wind.” The horizontal wind appears to have been $\sim 5 \text{ m s}^{-1}$ in their simulation. Here it is suggested that it may be due to a dynamic effect, when the cross-lake wind is significant, and not to a land breeze. At higher altitude, there is mostly convergence when the low-level pattern is strong (at night) and divergence over the whole lake when the low-level pattern is weak (during the day). This can be understood as follows: If everything else (e.g., temperature, stability, etc.) is the same over the whole area, the decreased roughness over the lake will cause the low-level wind speed over the lake to be higher. With a mean flow across the lake this must give rise to a divergence at the upstream shore and a convergence at the downstream shore: the dipole pattern. During nocturnal conditions this pattern becomes stronger; the stable stratification over land inhibits vertical

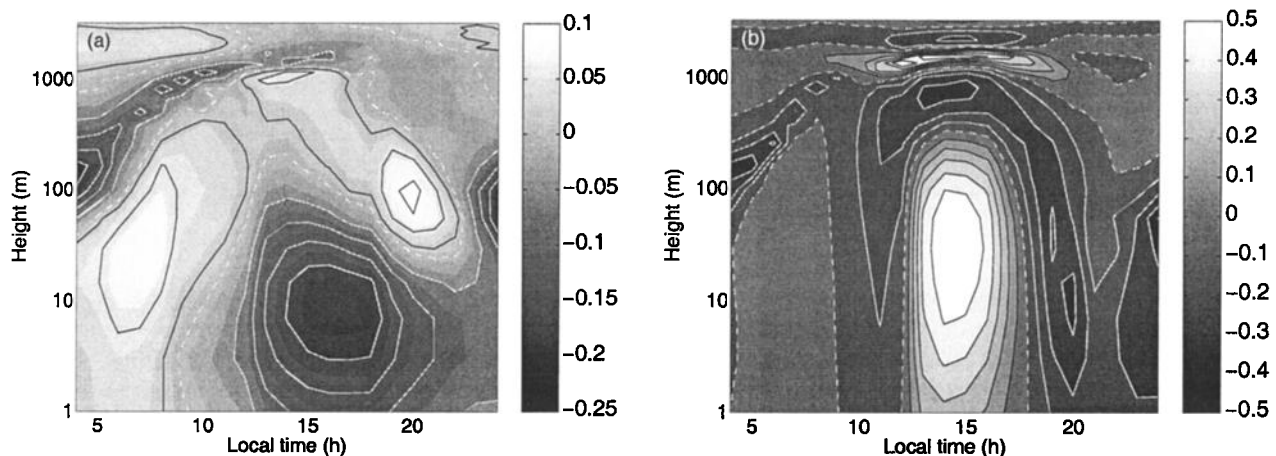


Figure 8. Linear-logarithmic time-height cross section of the divergence (10^{-3} s^{-1}) over Lake Tämnaaren showing (a) the average divergence averaged over the five runs CONT, LONG, NWMed, NWHi, and COOL and (b) mean divergence from NWLow. See the text for a discussion.

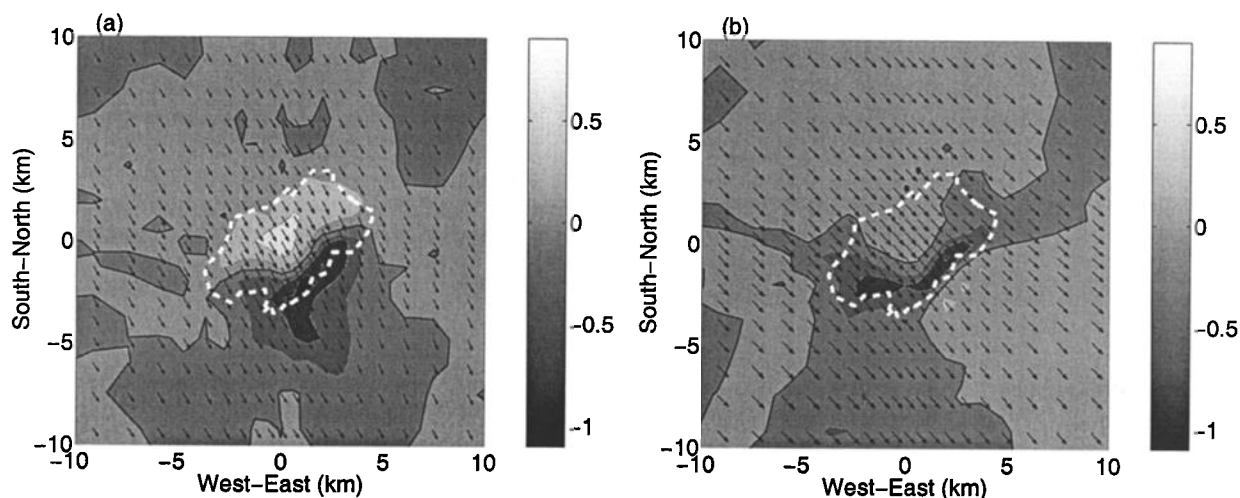


Figure 9. Contour plots of divergence (10^{-3} s^{-1}) at the height of 15 m along with wind vectors for the control run. The times are (a) 0400 and (b) 1600 LST. Lake Tämnaaren is indicated by the dashed line.

momentum transport by turbulence, and the winds here become weak. At the same time, the air over the lake is unstably stratified, thus enhancing vertical momentum transport here, and the acceleration of the wind at the upstream shore thus becomes more pronounced. As the convective internal boundary layer over the lake becomes deeper with downwind distance, momentum transport becomes even more efficient, and the low-level wind will continue to accelerate over most of the lake. While approaching the downwind shore, however, the deceleration in the stable internal boundary layer downwind of the lake becomes strong, causing convergence. This explains the nocturnal dipole pattern, with stronger divergence than convergence on average. The divergence pattern will promote vertical winds over the lake; downward and upward over the upstream and downstream shores, respectively. This will tilt the isotherms aloft “backward” relative to the flow, so for some height interval the flow will sense a “thermal hill” with subsequent deceleration (convergence) aloft.

During daytime conditions the opposite happens; the low-level wind speed over land is relatively high as a result of effective vertical momentum transport during convective conditions. A stable internal boundary layer forms over the lake, where acceleration of the wind over the upstream shore occurs due to decreasing roughness. However, the acceleration is reduced compared to nighttime conditions due to suppressed momentum transport in the stably stratified air. As the stable internal boundary layer over the lake grows, vertical momentum transport becomes progressively impeded, and the acceleration decreases downwind; it may even turn into deceleration before the flow approaches the downstream shore. Thus the divergence at the upstream shore is reduced and may even vanish, while the convergence over the downwind part of the lake is also reduced but remains finite. The dynamic perturbation of the PBL top now becomes weaker, owing to the weaker low-level pattern. However, the air over the lake at some height (e.g., below 100 m) is cooler, and the associated baroclinicity will result in a weak divergence above the lake. Overall, the vertical propagation of this dynamic lake effect reaches higher during the night, when the stable ambient conditions can sustain a weak gravity-wave pattern; this can actually be seen all the way to the model top (not shown). During daytime conditions the forcing becomes smaller and the lake-effect is

mostly confined to below 2–300 m, unless the thermodynamic effect takes over and a proper lake breeze forms.

Overall, it is the conditions over land, due to low thermal inertia of the surface, that determine the flow structure by changing the relative stability between the air over land and over lake. Above the blending height, with respect to the size of the lake, it is also the land conditions that determines the state of the flow simply because the land dominates the area.

The contrasting wind conditions over the lake during daytime and nighttime are illustrated in Figure 10, showing the wind speeds at 2, 10, and 100 m (note the offset scales) in a transect along the wind, across the lake. Figure 10a shows the conditions at 0400 LST in the morning. Note the low-level acceleration with a wind speed maxima close to the middle of the lake and the deceleration over the entire lake at 100 m. To check that this dynamic lake-effect is not just an artifact caused by the model boundary conditions, Figure 10b shows the average difference between the measured wind at 4 m over Lake Tämnaaren and the corresponding wind at ~ 5 m at Marsta, averaged over the month of June 1994. As expected, the low-level winds over the lake are significantly stronger during the night, while there is, on average, almost no difference between 0900 and 1800 LST. The dotted lines on either side of the average is one standard deviation; This is almost constant throughout the day.

In Figure 10b, note the behavior of the wind speed difference around 1500–1800 LST. This suggests that something may happen to the flow regime around this time, at least on some of the days. In Figure 8b it was evident that for the weak wind case (NWLow), a lake breeze formed. Figure 11 shows the divergence patterns for the two runs NWLow and NWInt at the height of 100 m at 1600 LST. NWInt was actually run to investigate the threshold wind speed for the lake breeze, and to see if there was an intermediate background wind speed, where a gradual transition between lake breeze and no lake breeze conditions could be seen. The low-level nocturnal divergence patterns for the NWLow and NWInt runs (not shown) are very similar to the control run pattern in Figure 9a. At 1600 LST, however, lake-breeze circulations have developed in both the NWLow and the NWInt runs. The lake-breeze front appears at the upwind shore, similar to the results of *Vidale et al.* [1997], *Taylor et al.* [1998], and *Shen* [1998]. Although the pattern is

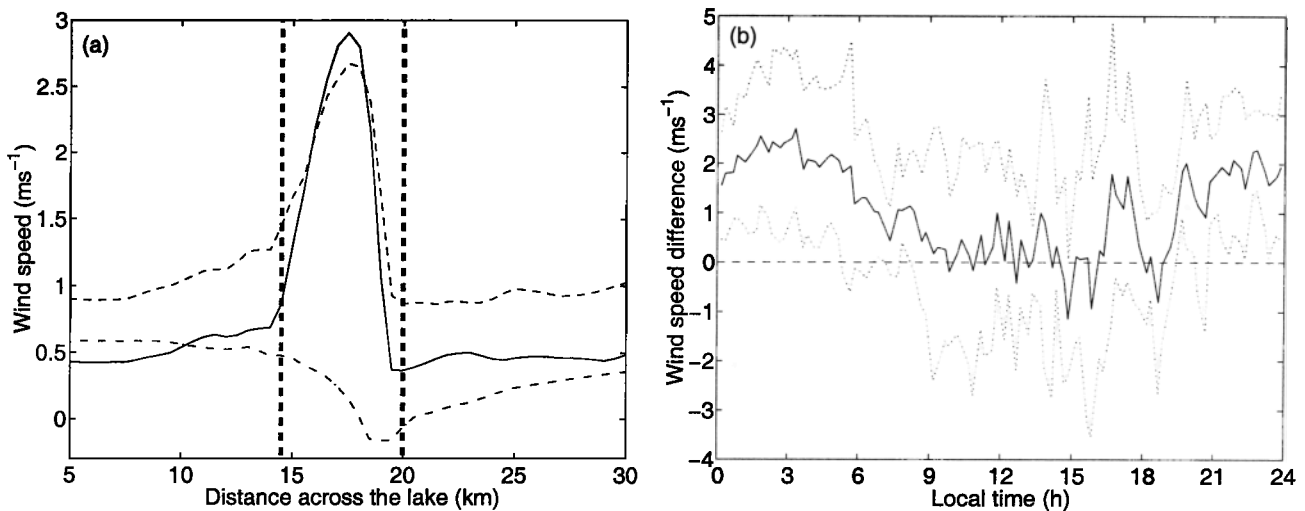


Figure 10. Plots of wind speeds over Lake Tämnen and the surroundings: (a) Simulated winds at 2 m (solid), 10 m (dashed), and 100 m (dashed-dotted), in an along-wind cross section over the lake taken at 0400 LST. Some lines are offset to increase resolution: the 10- and 100-m winds are offset by -1 and -6 m s^{-1} , respectively. (b) Mean measured wind speed difference between Lake Tämnen and Marsta for the month of June 1994 is shown, measured at 4 and ~ 5 m, respectively. The dotted lines are ± 1 standard deviation.

somewhat weaker for NWInt than for NWLow, both cases show well developed circulations, with horse-shoe shaped convergence zones at the upwind shore. During the night the low wind cases behave similar to the cases with higher winds and no land breeze appears. As the COOL run (with a reduced temperature over the lake) did not deviate significantly from the control run results, it may be concluded that the strength of the wind speed is more important in determining the flow regime, than the temperature difference (within reasonable bounds). The lake breeze seems to vanish here as the background flow increases above 6 m s^{-1} .

Sun *et al.* [1997] suggest that the measured divergence from an aircraft (of the wind component along the flight track) flying over Candle Lake could be predicted using the wind speed and the lake-land temperature difference. Similar scaling is attempted in Figure 12, using all the model data. In Figure 12a the mean divergence over the lake at 15 m is plotted against

lake-land temperature difference (ΔT), which shows the relationship between Figures 7 and 8. Even though the scatter is large, a few things can be noted. First, there appears to be a significant hysteresis in the low wind cases; Following the time development from the arrows in Figure 12a, we can see that in the morning the divergence does not increase until the temperature difference at 15 m has reached 0.5°C . However, in the afternoon the divergence starts to rapidly decrease although the temperature difference is not changing very much. The reason for this behavior is that the convergence area associated with the sea breeze front is initially included in the over-lake mean value of divergence but later on the front propagates over land and the associated convergence does not affect the over-lake mean value. When the temperature difference reaches its highest value the intensity of the sea breeze is also at maximum as shown by the maximum in mean divergence. For the moderate to strong wind cases, low-level divergence

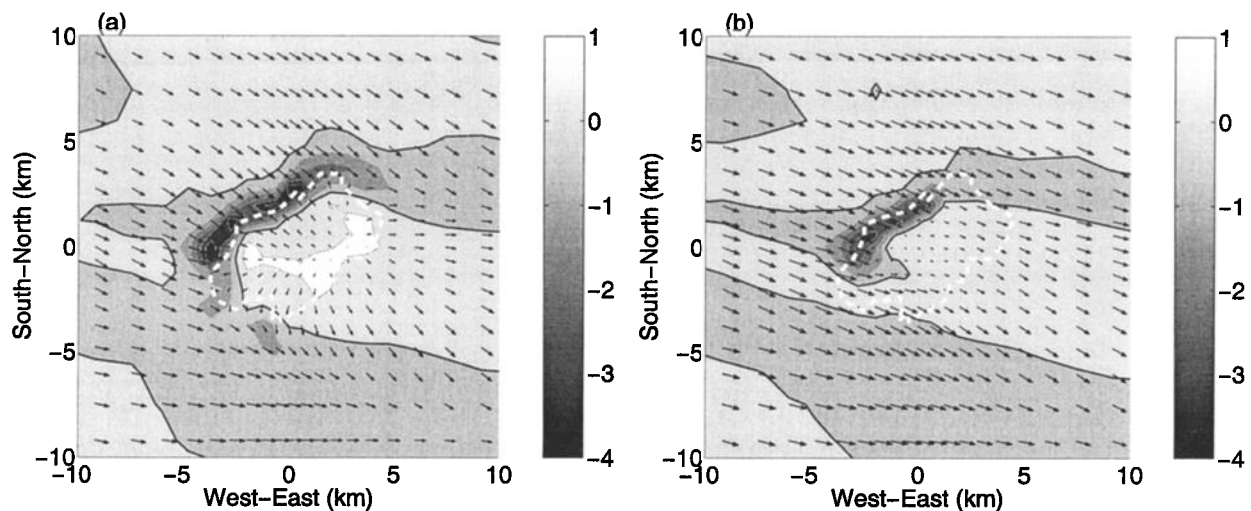


Figure 11. As Figure 9, but for (a) NWLow and (b) NWInt, at the height of 100 m at 1600 LST.

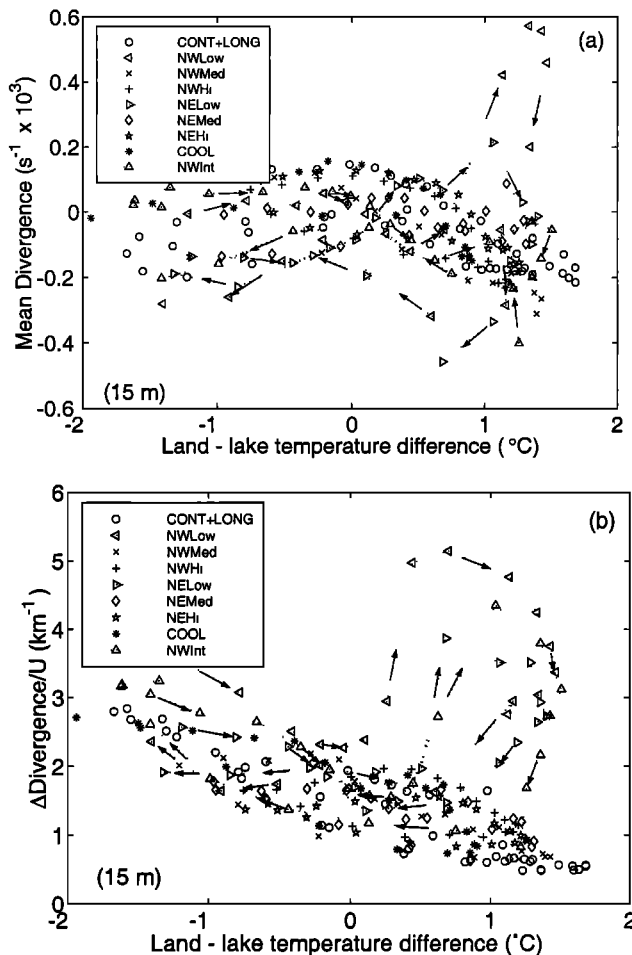


Figure 12. Divergence from all simulations at 15 m: (a) mean divergence against land-lake temperature difference ΔT and (b) the difference between the maximum and minimum divergence, scaled by the mean wind speed, against ΔT . Different runs are shown using different symbols as described in the legends. The arrows show the direction of time development.

has a tendency to decrease with increasing ΔT . The reason is that the divergence over the upwind part of the lake decreases in intensity during daytime while the convergence zone at the downwind shoreline does not change that much in intensity.

In Figure 12b the difference between the maximum and minimum divergence over the lake, divided by the wind speed, is plotted against ΔT . This scaling works very well for the moderate to strong wind cases since the intensity in the divergence pattern increases with wind speed. It also nicely shows the larger intensity of the dipole pattern during nighttime. For the low wind speed cases the scaling fails because the mean wind speed is not a relevant scaling parameter for a developed sea breeze. The hysteresis is due to the fact that the sea breeze front leaves the lake and propagates in over land in the afternoon, which causes the difference in divergence over the lake to be reduced.

While turbulent fluxes that are given by the model closure appear roughly consistent with the measurements, there is always the possibility that mesoscale fluxes may be significant over a domain of this size, which is typically the size of a grid square in a general circulation model (GCM). The model out-

put here provides a possibility to estimate mesoscale fluxes by directly correlating the resolved scale fluctuations in the different variables and averaging over some area, the whole model domain or parts thereof [Vidale *et al.*, 1997]. However, for the present simulations such mesoscale fluxes appeared to be insignificant (not shown), regardless of averaging domain. Mesoscale sensible and latent heat fluxes were generally smaller than $\pm 5 \text{ W m}^{-2}$, even for the lake vicinity, and momentum fluxes were typically $< 5\%$ of the turbulent flux, except for at the ABL top and aloft, where a small upward flux ($< 0.05 \text{ m}^2 \text{ s}^{-2}$) persisted above the ABL to $\sim 2500 \text{ m}$.

7. Aircraft-Model Results Comparison

Except for the control run, none of the runs correspond to meteorological conditions for any specific day. Attempts were made to quantitatively compare airborne observations with model results corresponding to similar geostrophic wind conditions as in the observations. However, it was concluded that details in the flow patterns are very sensitive to variations in

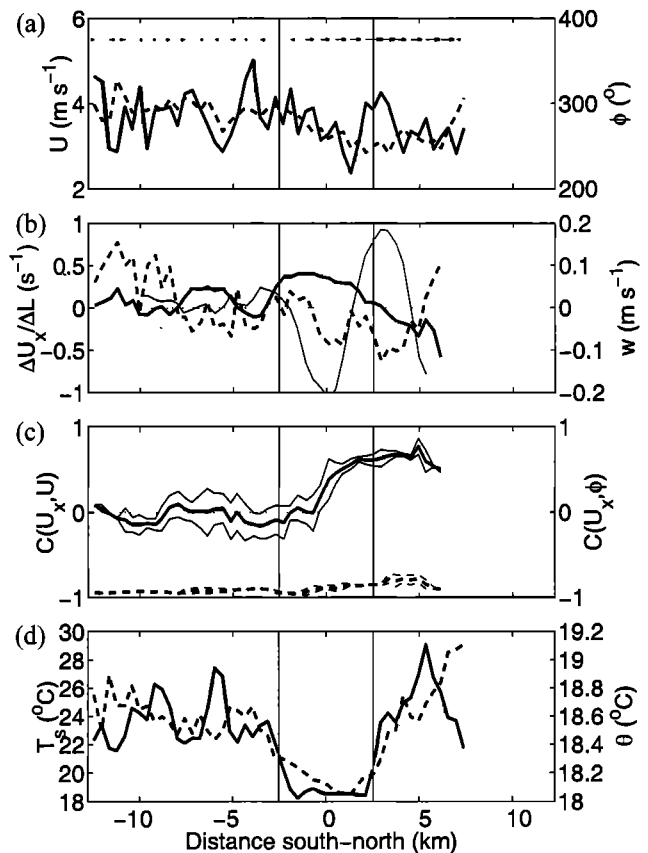


Figure 13. Airborne measurements on June 20, 1995, at 1230 LST. Thick solid and dashed lines refer to left and right axes labels, respectively. (a) total wind speed and wind direction plus the along-flight-track component of the wind speed (U_x) as arrows, (b) divergence, estimated from U_x and vertical wind speed, (c) correlation coefficient between U_x and the total wind speed U , $C(U_x, U)$, and between U_x and the wind direction ϕ , $C(U_x, \phi)$, along with standard deviations (thin lines), and (d) surface temperature and potential temperature. The vertical lines indicate the lake location (based on albedo). The light solid curve in Figure 4b is the aircraft measured divergence at 850-m height.

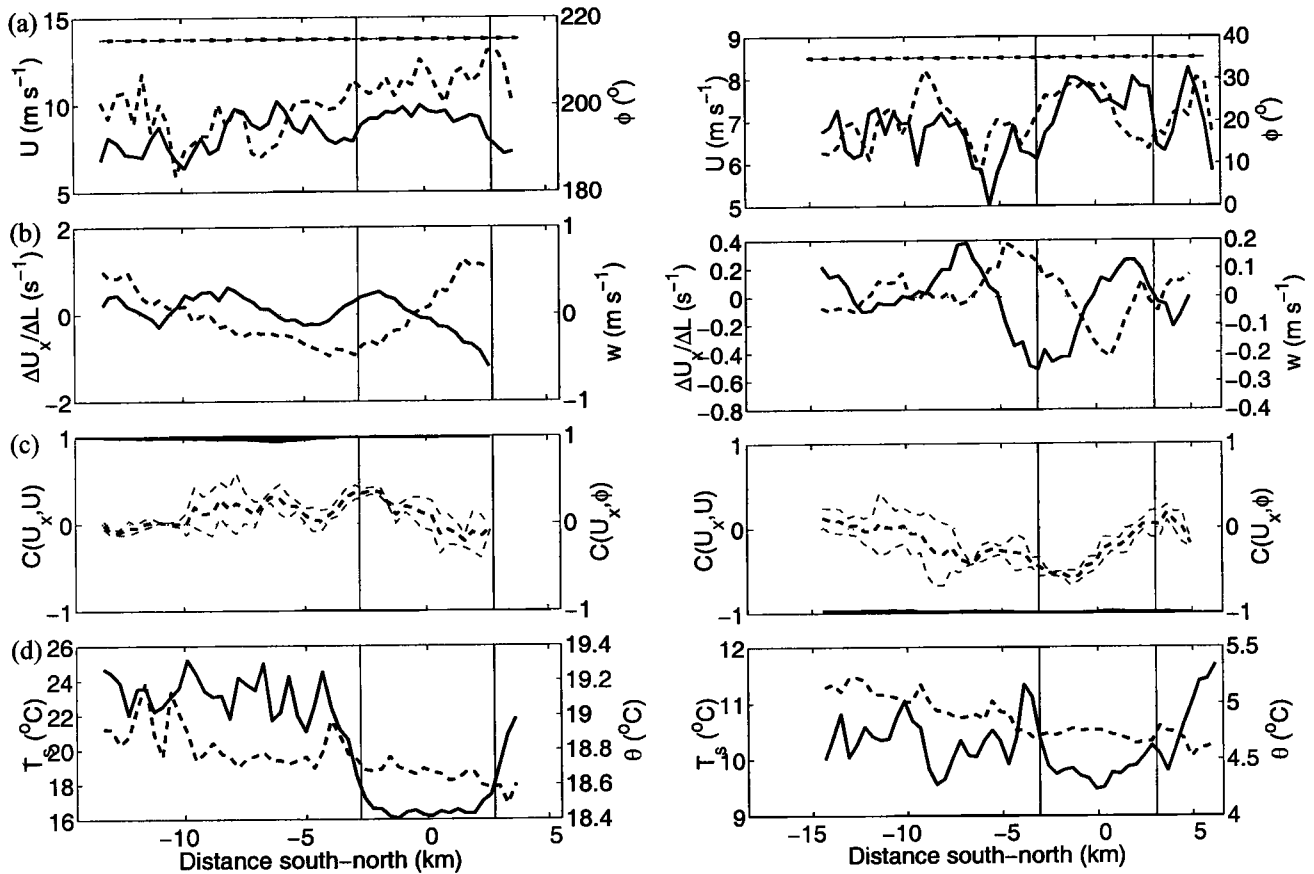


Figure 14. As Figure 13 but for (left) June 21, 1994, at 1330 LST and (right) May 8, 1995, at 1220 LST.

both wind speed and wind direction, which makes it difficult to compare actual numbers. Therefore we have chosen to perform a qualitative comparison of measurements and model results for two extreme cases among the cases available, representing one low (June 20, 1995) and two high (June 21, 1994, and May 8, 1995) wind speed cases, respectively (see Table 1).

Figures 13 and 14 show block averaged data (~ 500 -m aver-

aging length) calculated from two consecutive flight legs in each case. Each plot contains four panels. Figures 13a and 14a show the scalar wind speed, U , wind direction, ϕ , and the along-track wind-speed component, U_x , the latter as arrows. Figures 13b and 14b show the measured divergence along the track, calculated from least squares fits of U_x in overlapping windows of 5 km [Sun *et al.*, 1997]. Also shown here is the

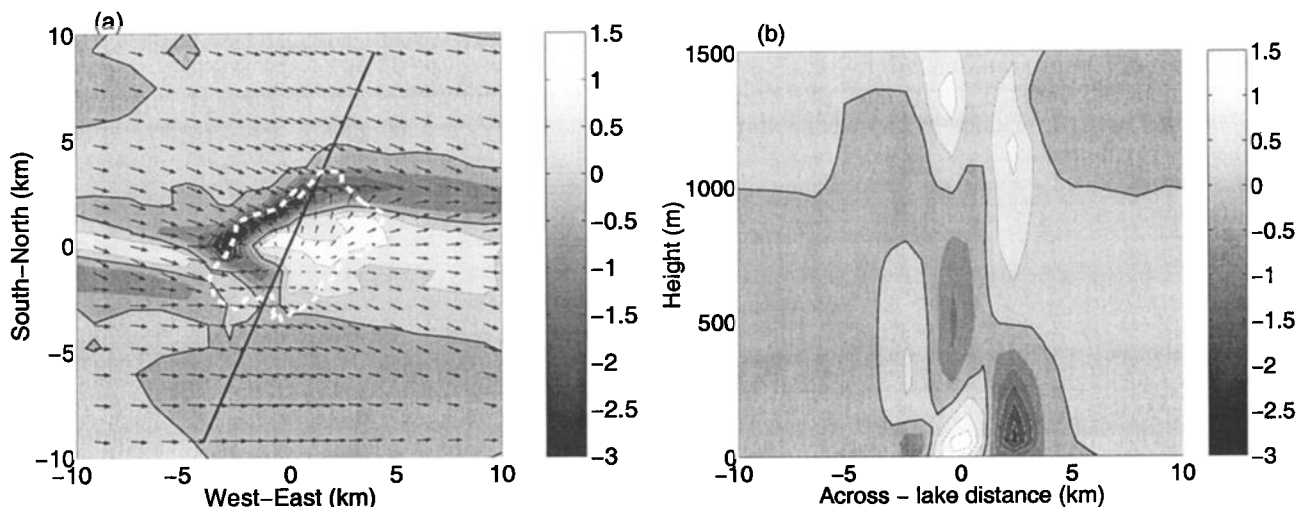


Figure 15. Divergence (a) at the height of 100 m at 1300 LST for the NWInt run and (b) across flow (looking downwind) distance-height cross section at the same time.

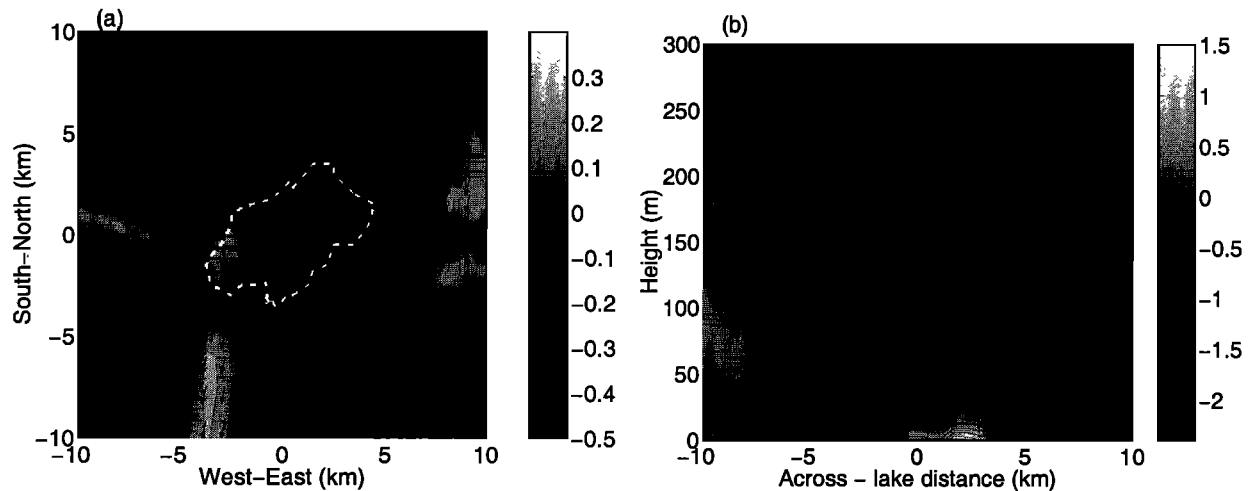


Figure 16. Divergence (a) at the height of 100 m at 1300 LST for the NEHi run and (b) along flow (from right) distance-height cross section at the same time.

vertical wind speed component. Figures 13c and 14c show the correlation coefficients for the correlation between along-track and scalar wind speed, $C(U_x, U)$, and between the along-track wind speed and the wind direction, $C(U_x, \phi)$. The correlation coefficients are calculated for each 5 km window. Figures 13d and 14d show the flight level potential temperature and the surface radiation temperature, clearly indicating the location of the lake. The thin vertical lines in all panels outline the section of the flight leg that is directly over the lake, as indicated by albedo values $< 6\%$ from the shortwave radiation measurements on the aircraft.

Comparing divergence from the model simulations with aircraft data is far from straightforward. The only divergence an aircraft will be able to measure is that in the wind speed component along the track of the aircraft. First, this means that the measured divergence may depend either on changes in the wind speed for a constant wind direction or on changes in the wind direction for a constant wind speed; in reality, both factors contribute to a varying degree. Second, note in Table 1 that the wind direction does in general not coincide with the track of the aircraft, which makes this issue even more difficult.

7.1. Low Wind Case

Airborne measurements for the weak wind case (June 20, 1995, at 1230 LST) are shown in Figure 13. According to the result in Figure 3a, a lake breeze circulation should exist under these circumstances. That implies divergence over the lake and convergence over the upstream lake shore which is also the case as shown in Figure 13b. The model results for the NWInt run at 1300 in Figure 11 give a qualitative picture of the 3-D divergence pattern that can be expected for this case. The model fields have approximately the same wind direction and horizontal temperature gradient, but 0.5 m s^{-1} lower wind speed compared to the airborne measurements. The low-level divergence over the lake, seen as the light region in Figure 15b, should generate sinking motion over the lake which is also observed by the aircraft. Note that the model results indicate that the convergence zone is not symmetric, it is stretched downwind north of the lake but not south of the lake. That is also seen in the airborne measurements as relatively strong convergence north north-east of the lake while it is close to zero south south-east of the lake. However, the aircraft ob-

served convergence zone is located north of the lake and not at the northern lake shore as given by the model. It may be that the lake-breeze front propagates more upwind in reality but less in the model due to the fact that there is a feedback to the surface temperature in reality but in the model the surface temperature is prescribed.

According to Figure 13, U_x is generally very well correlated with wind direction. That is obvious considering the wind field arrows in Figure 15a. Over the northern lake shore the correlation between U_x and wind speed has a local maximum, which can be related to the convergent wind field in this region. The low standard deviation of $C(U_x, U)$ at the same time as the correlation is strong indicates that wind speed variations are nearly as important for the resulting divergence in this region as wind direction variations.

Figure 15b shows that the low-level divergence pattern is compensated with an opposite pattern above, with an area of maximum convergence located around 700 m. This pattern corresponds well to aircraft measured divergence at 825 m in this case, indicated by the thin line in Figure 13b.

7.2. High Wind Case

Figure 14 shows two examples of high wind cases; June 21, 1994, with a south south-westerly wind speed of $\sim 9 \text{ m s}^{-1}$ and May 8, 1995, with a north northeasterly wind speed of $\sim 8 \text{ m s}^{-1}$. Both wind directions are thus approximately along the flight track, but from opposite directions. The measurements show that there is a clear convergence zone over the downstream shore of the lake in both cases, and a less pronounced divergence zone over the upstream shore. In the modeled divergence in Figure 16a, based on the NEHi run, the convergence zone is clearly seen while the divergence over the lake is very weak. As a consequence of low-level convergence and divergence, respectively, the vertical wind speed is positive over the downstream half of the lake and negative over the upstream half of the lake. The along-flow distance-height cross section in Figure 16b shows that both divergence and convergence decrease rapidly with height. However, the convergence regime extends higher than the divergence regime. As seen in Figure 14c the measured divergence is dominated by changes in wind speed.

8. Discussion

Many of the variations in the airborne measurements, close to the lake, have been compared with corresponding model results, even though there are as large variations farther from the lake, which do not have any correspondence in model results. The reason is that model simulations are always more or less idealized and do not include many of the processes that take place in the atmosphere. In this case that concerns for example convective eddies, which are neither parameterized nor resolved in the model. Another simplification in the model is the surface boundary conditions. In these simulations we do not solve a local energy balance in every grid point; thus the lower boundary is not allowed to adjust locally with respect to vegetation (type and density), soil moisture, and roughness. It will thus not be possible to resolve all changes in the surface conditions. Thus changes observed in aircraft measured parameters that are due to locally changing surface conditions or convection over land, cannot always be expected to be described by the model. Close to the lake, however, measurements and simulations should be comparable.

To make sure that the scales of interest are resolved we chose a relatively high resolution in the center of the model for this study ($600 \times 600 \text{ m}^2$). As discussed in the model description, this makes the use of a hydrostatic model just barely justified for some flow regimes. Further, when the low-level wind direction is oriented along the x or y axis of the model the divergence in the best resolved areas seems to be sensitive for model resolution. That indicates that the modeled divergence pattern may be unrealistic in some areas for some situations. However, for the results discussed here such errors do not change the qualitative picture of divergence patterns.

In the model simulation results it has been shown that horizontal variations in temperature and divergence are very pronounced close to the surface but decrease significantly with height. Above 200–300 m the simulations indicate compensating divergence patterns with respect to the patterns close to the surface. Thus the 100- and 200-m airborne measuring levels are located in between, sometimes causing the signals to be relatively weak. In that sense, it would have been better to perform measurements below 100 m and above 200 m. However, as mentioned before, the 100-m level was the lowest permitted flight level and the 200-m level was chosen with respect to turbulence fluxes which show large vertical divergence in the lower ABL [Samuelsson and Tjernström, 1999b].

Venäläinen *et al.* [1998] studied the wind speed dependence on overwater fetch (Lake Råksjön and Lake Tämnaaren in the NOPEX region) for off-shore wind from a forested area. They observed that the wind speed acceleration is very large for the first 100 m and that the wind speed continues accelerating off-shore but with reduced magnitude. This pattern appears to be at least qualitatively verified in the present model simulations.

Complicated flow patterns occur when forces due to the friction and the thermal pressure gradient are comparable in size. Results from a model run with simplified land use, just forest and lake, exclude the possibility that such patterns would be an effect of heterogeneity in land use. Also the shape of the lake does affect the patterns. Since the terrain variations in this area are small it is unlikely that terrain variations would have any impact on the flow regimes, especially not during daytime. That has also been confirmed from results based on a model run with constant terrain height but realistic land use.

9. Conclusions

In this paper the modification of the ABL flow due to the presence of a lake has been investigated. Lake Tämnaaren, located in the northern part of the NOPEX region in southern Sweden is studied. The lake is 34 km^2 with a width in the range $\sim 4\text{--}7.5 \text{ km}$.

For this study a mesoscale numerical model, the MIUU model, was used to simulate the dependence of the ABL flow on surface temperature and roughness differences at different background wind speeds and directions. The model results have been qualitatively compared with in situ airborne measurements performed along a track passing over the lake. Surface observations of wind speed and sensible heat flux have also been used to estimate the minimum lake size required to develop a lake breeze circulation.

A general result from this study is that the effect of surface roughness changes are very important for the observed flow modification and in most situations dominate over the effect of diurnal variations in surface temperature difference between land and lake. The most common pattern that results from the simulations is a low-level dipole divergence pattern over the lake, with divergence over the upwind half of the lake, due to acceleration of the wind that occurs when it is advected over the smooth lake surface. Close to the downwind shore a convergence area occurs due to the increased roughness. The dipole pattern is strongest during night time since the unstable stratification over the lake induce effective vertical momentum transport, while the momentum transport over land is suppressed due to stable stratification. This situation was observed for low-level wind speeds down to 4 m s^{-1} . The low-level divergence pattern induce upward vertical motion over the downwind shore and downward vertical motion over the lake. The dipole pattern becomes stronger with increasing wind speed and weaker with increasing land-lake surface temperature difference.

For low and moderate background wind speed (geostrophic wind $\leq 6 \text{ m s}^{-1}$ according to the simulations) a lake breeze circulation develops during the afternoon with low-level divergence over the lake and low-level convergence over the upwind shore. This pattern is compensated with an opposite divergence pattern at $\sim 700 \text{ m}$, however, the depth of the circulation often seems to be less than the ABL depth. When the friction (background wind speed) and pressure gradient forces (horizontal temperature gradient) are of similar importance the flow pattern becomes more complicated than those described by the dipole or lake breeze patterns.

The model simulated flow modifications are partly supported by airborne measurements, although there are for some situations difficult to judge if the measured values are due to effects not considered by the model. It has also been shown that an expression derived by Doran *et al.* [1995], to estimate the smallest required size of a lake to induce a mesoscale thermal circulation, gives good results in this case. The expression uses information on background wind speed and surface sensible heat flux difference.

Many studies focus on weak wind cases with well developed lake or land breeze circulations. The results from this study indicate more complex patterns in relation to horizontal pressure gradient and friction forces.

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