Boundary layer meso-scale flows

Hannu Savijärvi
University of Helsinki

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- Mechanical channeling
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- Thermal channeling on a sea gulf
  - spring, early summer (**cold sea**)
  - late autumn, winter (**warm sea**)

(Turbulence is needed to transfer to the air the effects of surface inhomogeneities, which drive these meso-scale phenomena)

Mechanical channeling

Conservation of mass (continuity equation):

$$\frac{\partial \rho}{\partial t} = -\nabla_3 \bullet (\rho \vec{V})$$

If density is nearly constant (e.g. in the PBL; Boussinesq approximation), then: div(**V**)=0.

If, moreover, stratification is stable, air is hard to lift vertically so w remains small, w=0.

It follows that:

Stable PBL air flowing along inhomogeneous ground must increase its speed to conserve mass e.g.:

- in topographic gaps: strong ”gap winds” in mountain passes

- in straits at the sea: stronger winds between islands than over the more open sea, especially when sea is cold (i.e. in springtime)

(Note that roadworks cause jams in motorways because the traffic must slow down for safety. For no increase in ’car density’, speeds should be increased instead!)

This mechanical effect also channels the prevailing winds to blow along longish valleys and lakes.

Two examples:

1) HIRLAM (5 km Δx) simulation of Lake Tanganyika fine-scale climate: July 1994 average of four 7-day restart cycles using ECMWF boundaries and initial states. Lake T is a rift valley lake with a gap in the middle; Southeasterly trade winds => ”Gap winds”. (Savijärvi and Järvenoja, Meteor.Atmos.Phys. 2000)

2) AROME (2,5 km Δx) forecast of 10 m winds along the Gulf of Finland during easterly flow 14.8.2006

Example 1: Lake Tanganyika surface winds

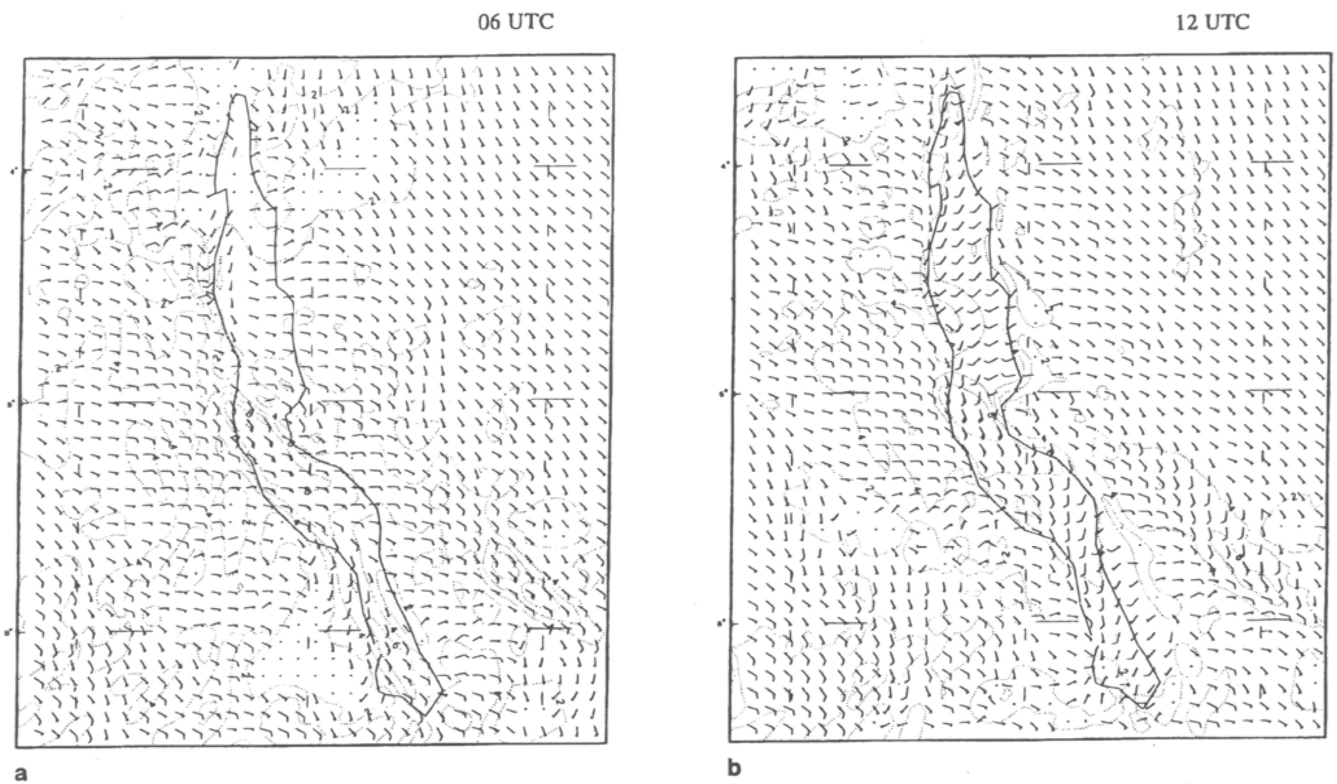
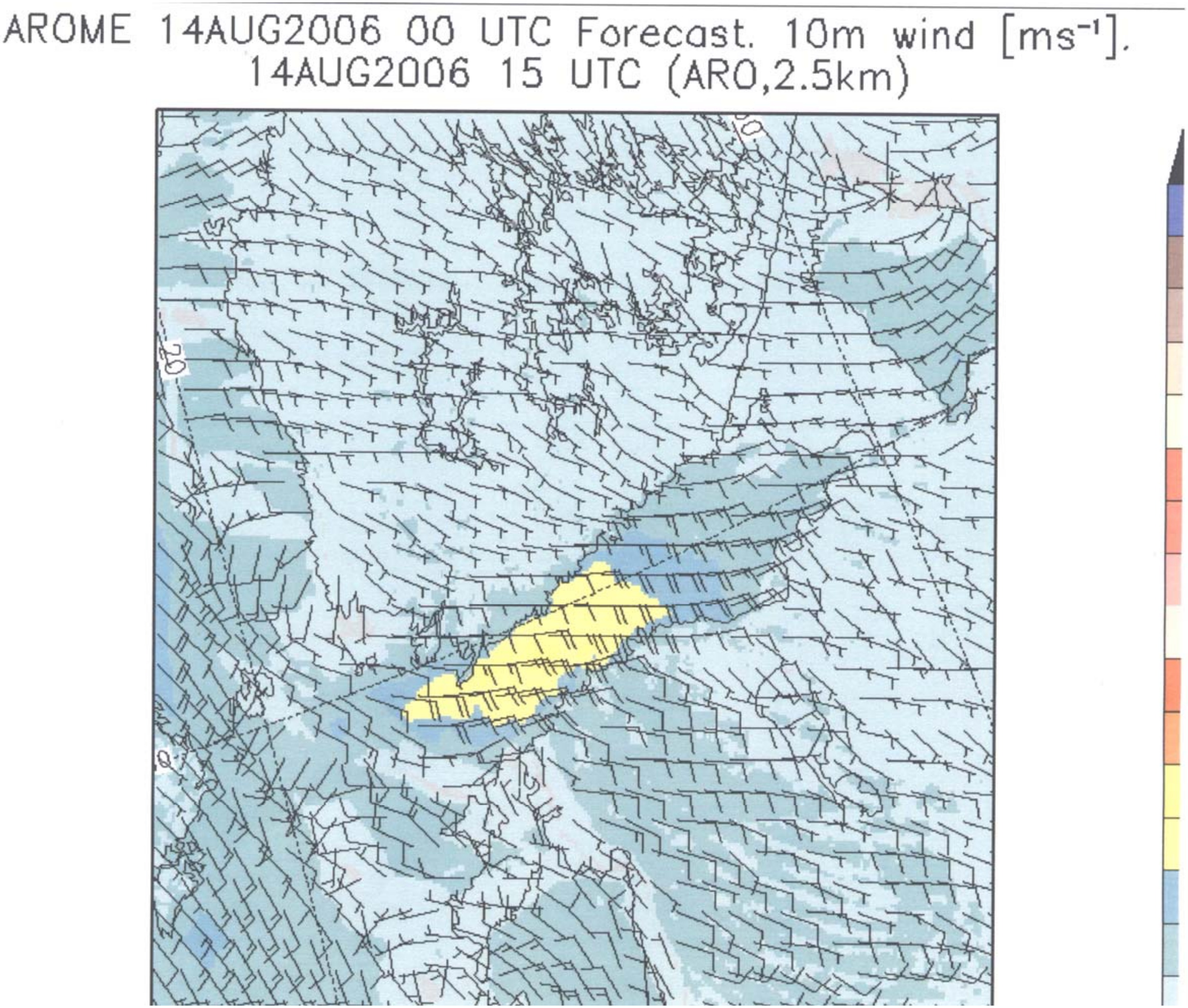


Fig. 8. The modelled 10 m mean winds and isotachs (contour interval 2 m/s) in the dry season (July 1994) in 16.5 km resolution. (a) Local morning, 06UTC; (b) Local afternoon, 12UTC

note stronger ’gap winds’ in the narrow part of the lake; note coastal land breezes/katabatic winds in the morning (06UTC), sea breezes/upslope winds during local afternoon (12UTC).

Example 2:


AROME 14AUG2006 00 UTC Forecast. 10m wind [ms⁻¹].
14AUG2006 15 UTC (ARO,2.5km)
20
20

Thermal 'heat island' circulations

Basic flow U past a warm/cold surface strip: a steady-state perturbation flow u',w' is created by heat transfer:

warm strip, cold strip,

heat island circulation, 'warm plume' downstream

anti-heat island circulation 'warm plume' upstream

(Internal boundary layers start to grow downstream. Turbulence is essential in transferring the surface temperature and roughness differences to the air)

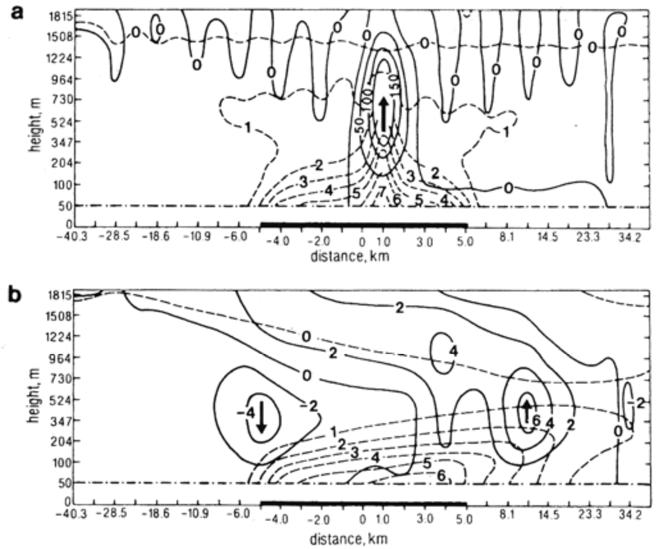
Coriolis force produces also a **v'-component**.

If there is **no basic flow**, a warm strip triggers flow near the surface toward it, convergence leads to rising motion above it, air returns to the cool sides at the top of the PBL.

Urban heat island circulation is one example of heat island circulations. Monsoons can be considered as a large-scale example.

Linear theory and nonlinear numerical models (LES, meso-scale) can be used to study the physics of heat island flows.

(e.g. Savijärvi and Matthews, 2004 JAS: local flow effects around small tropical islands, downstream cloud plumes etc.)

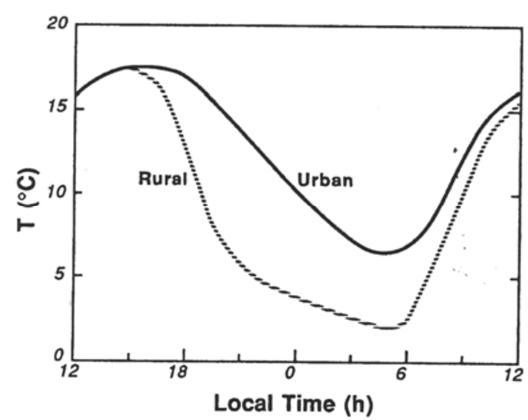


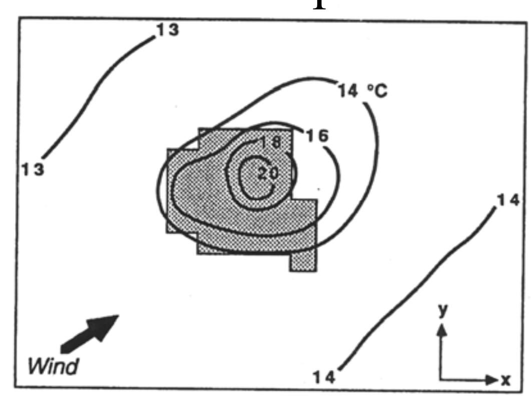
U = 1 m/s, warm dome; nearly symmetric heat island circulation

U = 5 m/s, warm plume downstream

Fig. 23. Development of a local circulation over a heated island (a) with $\bar{u}$ = 1 m s^{-1} and (b) with $\bar{u}$ = 5 m s^{-1}. In both cases ΔT = 10°C, and the simulation is for 5 hours after onset of island heating. Solid lines indicate vertical velocity perturbations in centimeters per second, and broken lines temperature disturbance in degrees Celsius [after *Tanouye*, 1966].

urban heat island, note warm plume downstream





Slope winds

-Thermally-cooled nighttime slopes trigger shallow downslope (katabatic, drainage) winds (strong in Antarctica, Greenland)

- Sun-warmed daytime slopes trigger weaker daytime upslope winds

night

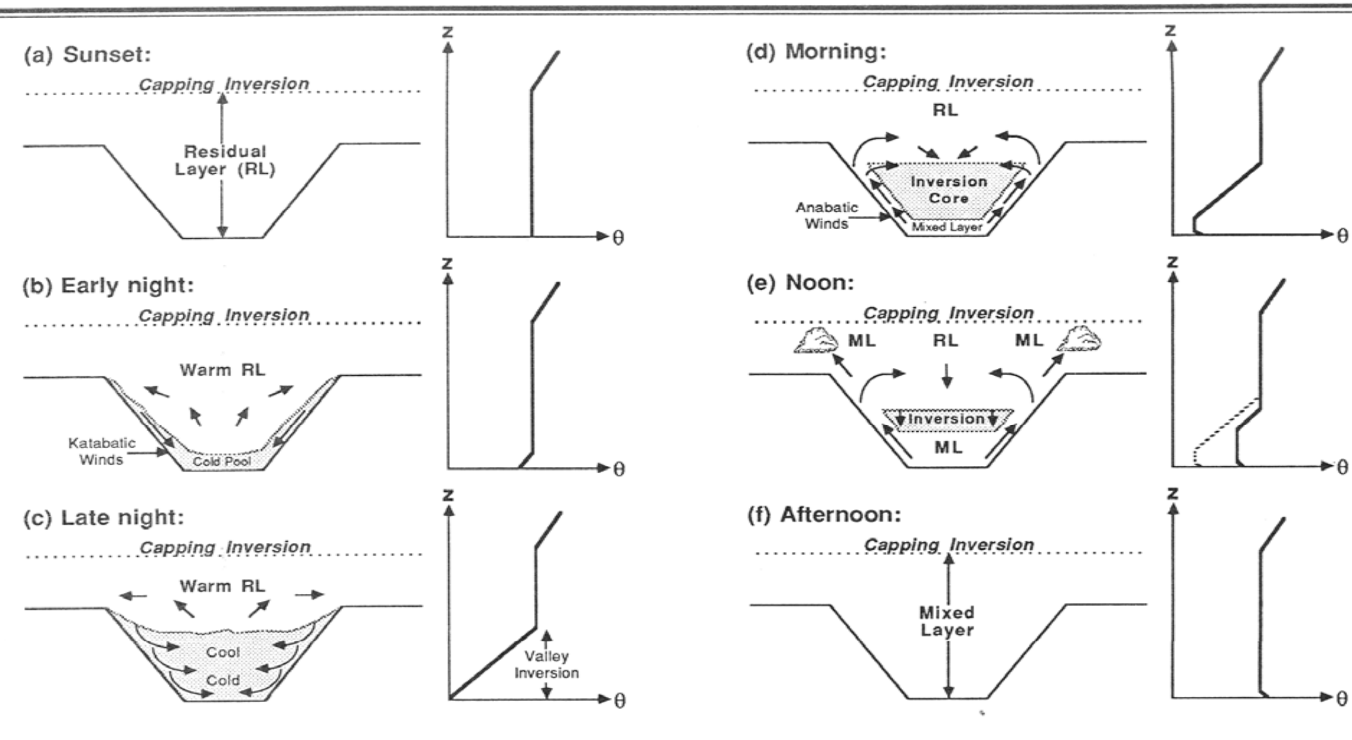


Fig. 14.2 Idealized evolution of the cross-valley circulations during a diurnal cycle. Potential temperature profile corresponds to sounding made from the center of the valley (after Geiger, 1965; and Whiteman, 1982).

day

(Stull 1988)

Interactions are possible: Consider a city in a river valley bottom (e.g. Lanzhou, N-W China). Then:

-During daytime and weak or no basic flow, near the surface urban heat island circulation blows **toward** the city centre, while upslope cross-valley winds blow uphill, **away** from the centre. These opposite circulations nearly kill each other so there is no wind (no ventilation). Hence daytime pollution can increase.

-During night-time, cross-valley downslope winds help the urban heat island circulations to concentrate pollution to the city centre, where it accumulates due to higher night-time stability (i.e. less ventilation).

(Savijärvi and Jin, BLM 2001: Local wind in a valley city;
’smog trap’ situations)

Sea breeze

Sunny, calm morning: Convection heats PBL air over land => high pressure at the top of PBL => flow to sea => PBL mass loss => '**thermal low**' (1 mb) in land surface pressures => near-surface flow from cool sea => sea breeze circulation cell.

Coriolis force turns the winds. The whole cell moves inland at about 3 m/s.

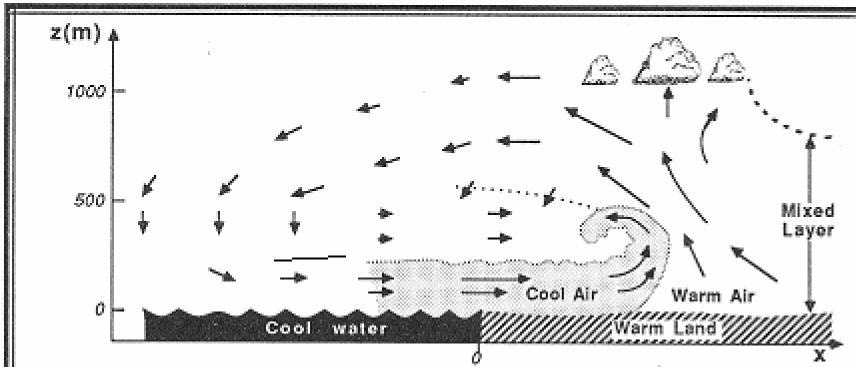
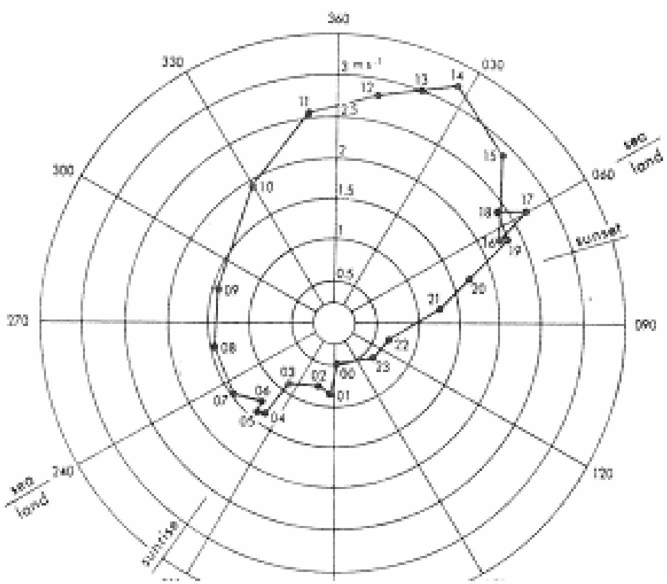


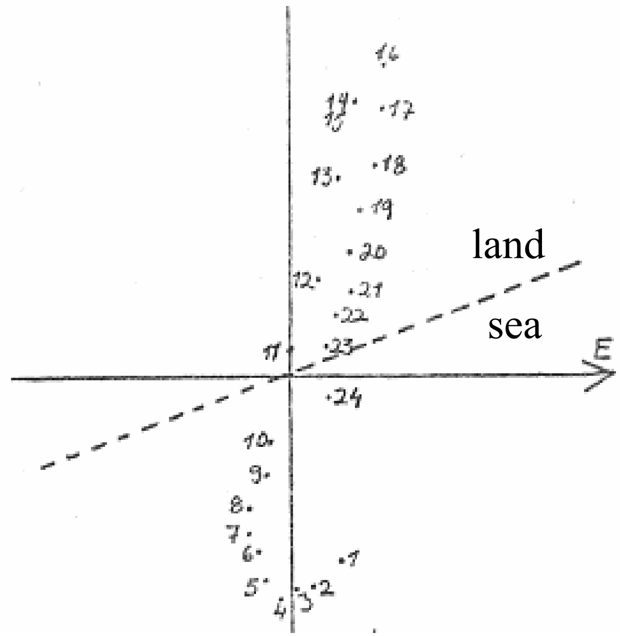
Fig. 14.7 Idealized sea or lake-breeze circulation. (After Lyons (1975) and Ogawa, et al., 1986).

Stull (1988)

Land breeze blows from cool land (e.g. during nighttime)

Examples of observed calm-case diurnal sea and land breezes: mean hourly wind vectors to/from origo





Wind rose: Kinloss, Scotland

Wind hodograph: Helsinki, Finland

Sea breeze is about 4-7 m/s, return flow 2-4 m/s at 1-2 km.

Controlling environmental factors:

-Temperature difference between sea air and land air

Should be at least 8 K (No SB in cloudy weather).

Best conditions: during weak, sunny cold outbreaks

-Prevailing large-scale flow

Tail wind from sea: weakens sea breeze, carries it inland

Wind along coast: sea breeze adds to the large-scale wind

Head wind from land: when weak, strengthens the sea breeze, by keeping the cell at the coastline.
When strong ($V_g > 8$ m/s), destroys the sea breeze.

Example 1: UH 2D model sea breeze hodographs (4 h intervals) for Helsinki during 5 m/s prevailing flow (**V**g) from many directions relative to coastline:

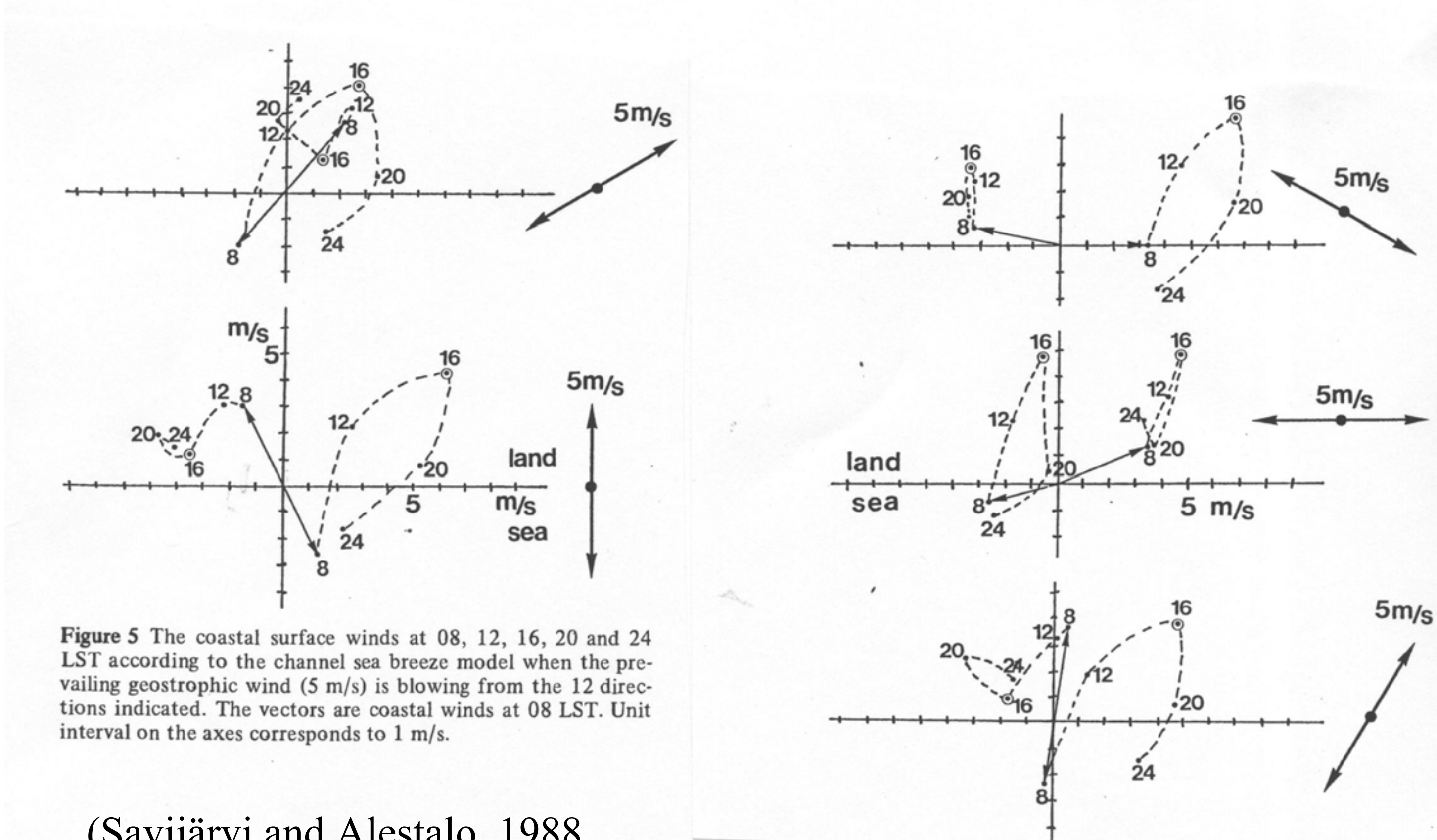
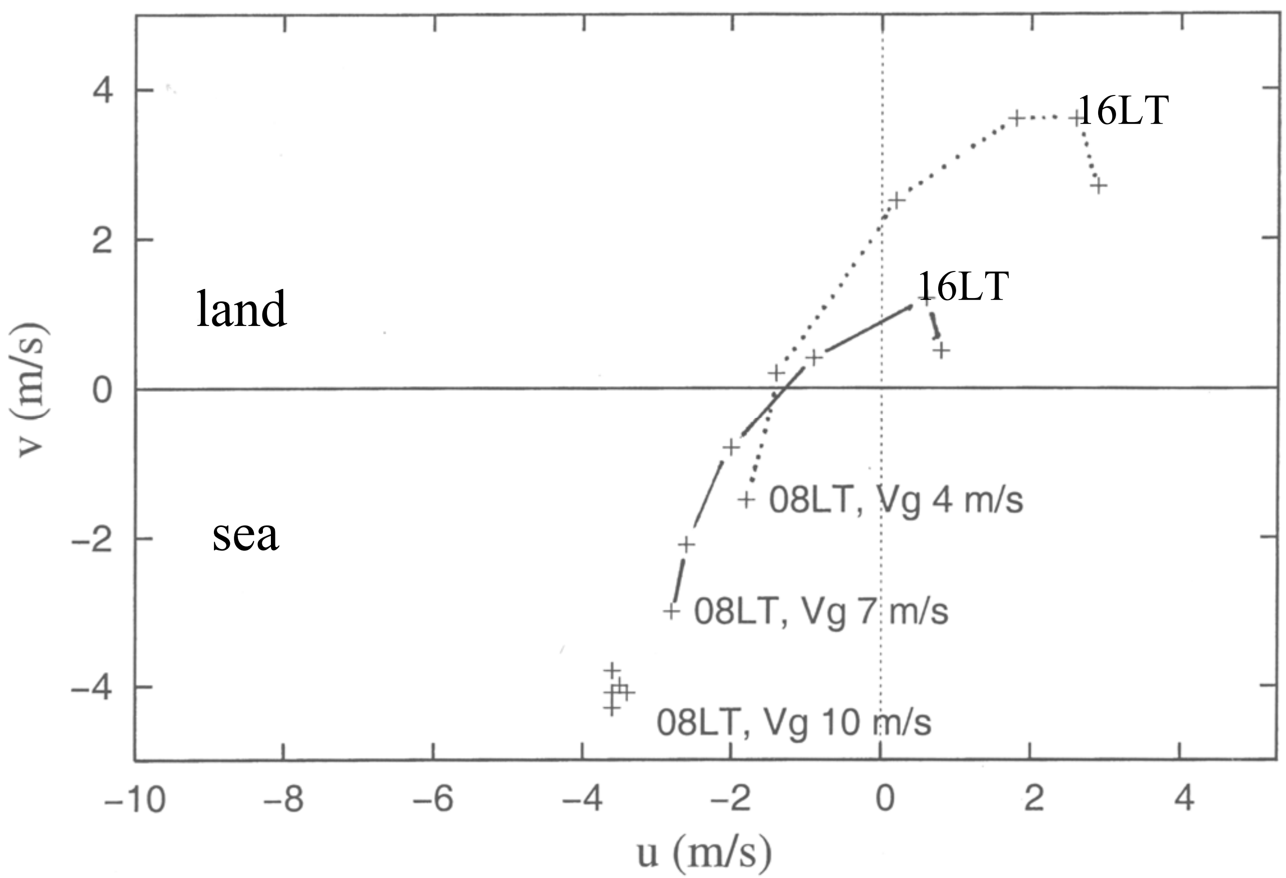


Figure 5 The coastal surface winds at 08, 12, 16, 20 and 24 LST according to the channel sea breeze model when the prevailing geostrophic wind (5 m/s) is blowing from the 12 directions indicated. The vectors are coastal winds at 08 LST. Unit interval on the axes corresponds to 1 m/s.

(Savijärvi and Alestalo, 1988
Contr.Atm.Phys.)

Example 2: Model hodographs (10 m wind vector end-points at 2 h intervals) for Helsinki during **Vg from northeast**. Note: **Strong afternoon sea breeze for Vg of 4 m/s, Nearly calm for Vg of 7 m/s, No sea breeze for Vg of 10 m/s.**



Example 3:

Clear sky, moderate southerly basic flow over the Gulf of Finland:

SB cell of the Finnish coast disappears quickly inland in tail wind. In contrast, SB cell of the Estonian coast is strong and it may be pushed over to Finland. Then, **easterly winds** are observed **on the Finnish coast in the evening**, with an **easterly low-level jet** developing over the Gulf of Finland.

(Savijärvi and Alestalo, 1988 Contr.Atm.Phys.)

Thermal channeling on sea gulfs

When windy, well-mixed PBL air from over rough land crosses over to smooth sea, vertical mixing and surface drag (friction) is suddenly reduced by a factor of 10-100 => PBL air accelerates over the sea in inertial oscillation with period T=12 h/sin(latitude), forming a low-level jet a few hours later:

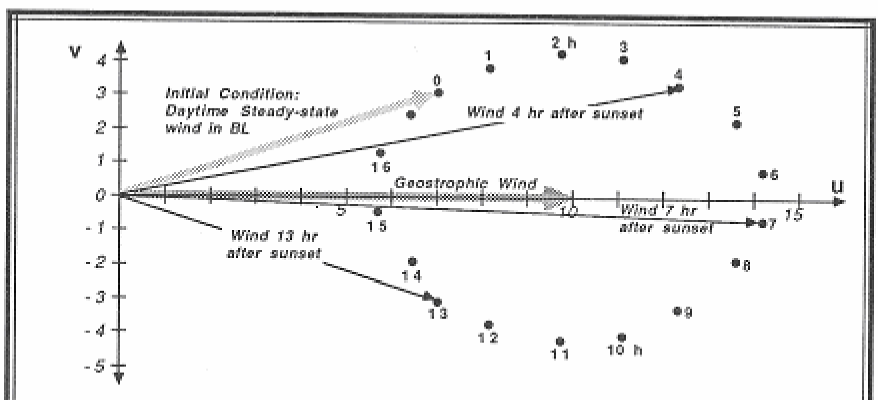


Fig. 12.18 Diurnal variation of the low level jet associated with the inertial oscillation.

from Stull (1988)

In moderate southeasterly geostrophic flow over the Gulf of Finland (60N) the wind maximum occurs (at about 200 m height) after a 4-5 h travel. Air is then near the Finnish coast, forming a shallow easterly **inertial low-level jet** (iLLJ), which is strongest during late afternoon.

If the day is sunny, sea breezes and enhanced convection may further increase the iLLJ speeds to 5-6 m/s over Vg, with **supergeostrophic surface easterlies resulting along the north coast in the afternoon.**

(Savijärvi, Niemelä and Tisler, QJRMS 2005)

An analogous effect may also occur for westerly basic flow, and of course on the southern (Estonian) coast as well.

Example: a sunny August day:

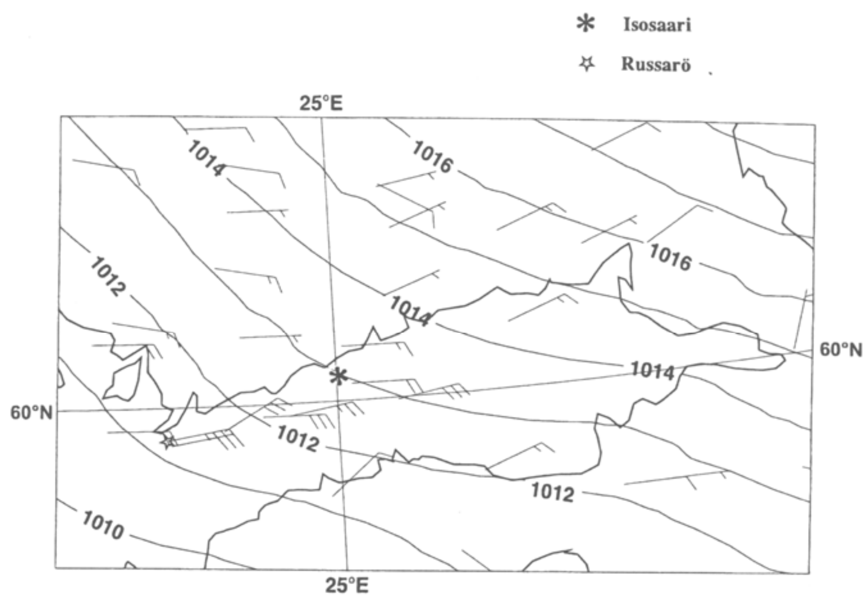


Figure 7. Observed surface winds on 29 August 1997, 1200 UTC, and HIRLAM surface pressure +12 h forecast (hPa) from 29 August 1997 0000 UTC with 7.7 km horizontal grid length. Stations Isosaari and Russarö (Fig. 11) are marked.

note strong 10-14 m/s surface winds with large cross-isobar angles on the Finnish side of the Gulf.

Vg is 10-12 m/s from SE over the Gulf of Finland

(from Savijärvi, Niemelä and Tisler, QJRMS 2005)

HIRLAM (7.7 km) +12h winds at 30 m height:

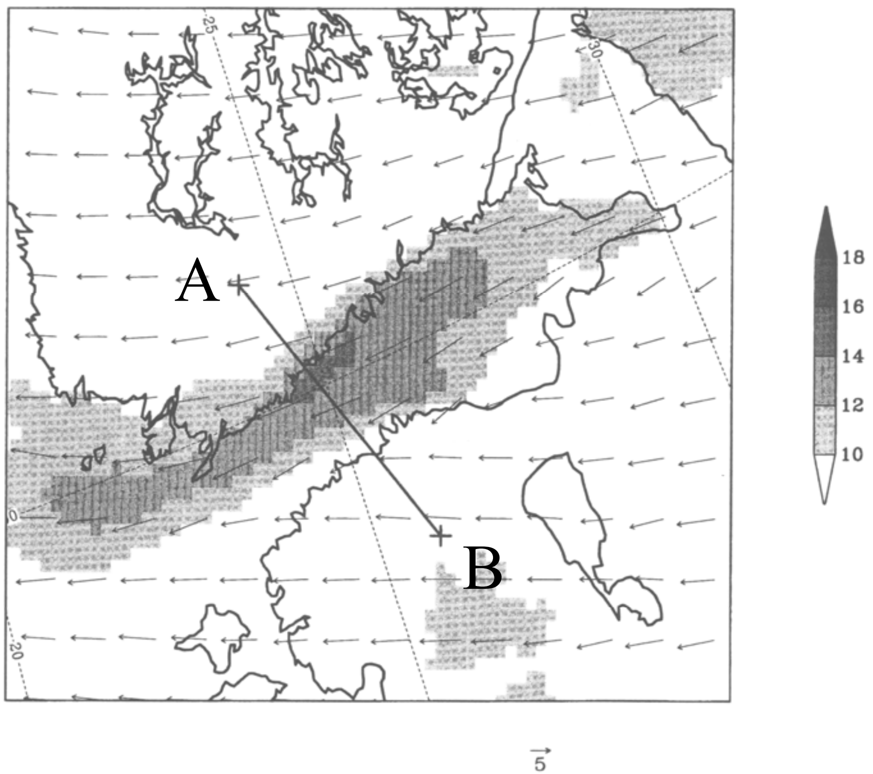


Figure 8. Wind vectors and wind velocity (grey scale, m s^{-1}) on the lowest model level (about 30 m) in the HIRLAM +12 h forecast valid on 29 August 1997, 12 UTC. The cross sections of Figs. 9 and 10 are along the line shown (from Finland in the north to Estonia in the south).

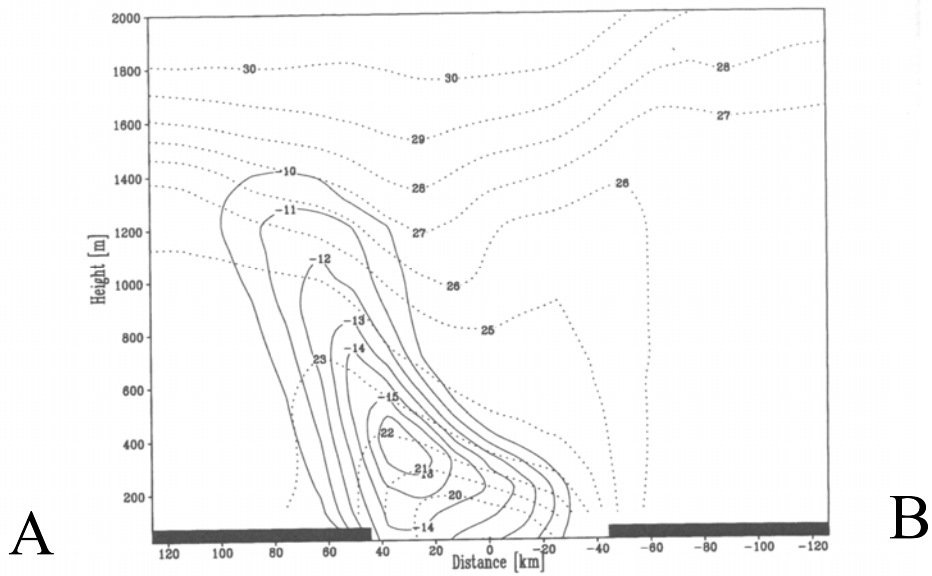


Figure 9. Cross sections (Finland on the left and Estonia on the right) of the HIRLAM potential temperature (dotted lines, °C) and the wind component parallel with the coastline (solid lines, m s^{-1}). The negative values are towards the viewer and only the values $|v| > 9$ m s^{-1} are shown for clarity.

along-coast wind component:
16 m/s easterly LLJ along
the Finnish coast (left)

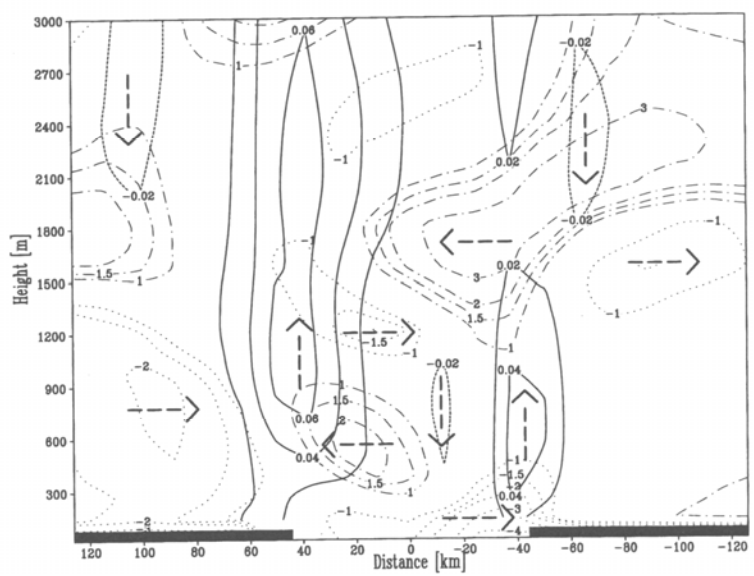


Figure 10. Cross sections (Finland on the left and Estonia on the right) of the anomaly (wind minus area-averaged wind) component perpendicular to the coastline (m s^{-1}), and the vertical velocity (m s^{-1}, solid). The positive (dot–dashed) and the negative (dotted) wind values are towards Finland and Estonia, respectively.

across-coast circulations:
sea breezes near both coasts

Five-year observed May-June wind roses for two islands near the Finnish coast, and a 2D model simulation for constant Vg-speed

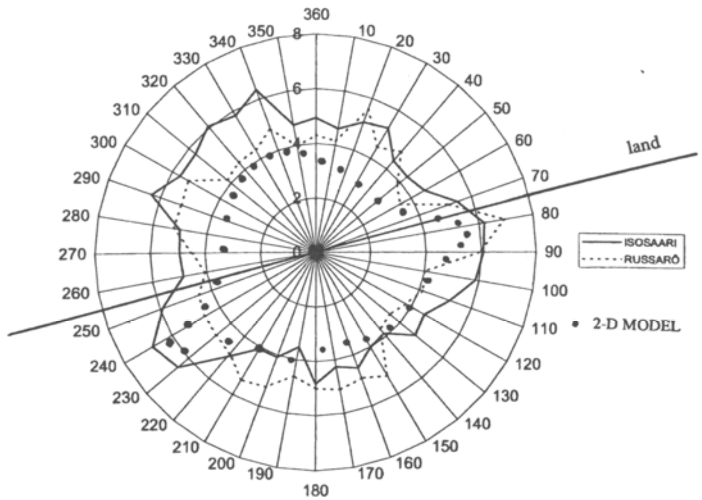


Figure 11. The wind rose of the observed mean surface wind speeds (m s^{-1}) for May–June 1999–2003 from the islands Isosaari (solid line) and Russarö (dashed line). Also shown is the wind rose of the overcast channel 2D simulations (10 m wind vectors 3 km out at sea) for moderate $\mathbf{V}_g$ (10 m s^{-1}) from 32 directions (dots; maximum value of wind from 238° corresponds to 8.5 m s^{-1}).

SST effects on thermal channeling

Gulf of Finland, south-easterly basic flow:

Cool sea (spring, early summer):

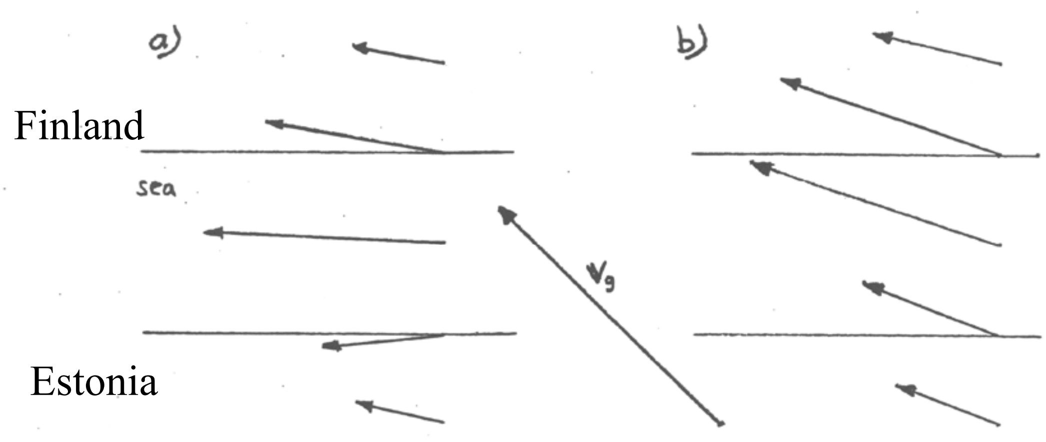
-**Cloudy weather**: Basic flow over cool sea creates an anti-heat-island circulation. This enhances the inertial LLJ and channels surface winds to blow strongly along the Gulf (as is observed)

-**Clear weather**: Also a sea breeze is triggered on the head wind (Estonian) side. This enhances the downstream iLLJ even more.

Warm sea (late summer): iLLJ is now weaker (no anti-heat island circulation), but vertical mixing is enhanced by the warm surface, so surface winds are strong over the sea, albeit with a smaller cross-isobar angle.

Example: 2D model 10 m afternoon winds over the Gulf of Finland, for 10 m/s Vg from SE. Summer case, overcast, T2m about 17 C.

a) SST cool (13 C) b) SST warm (20 C)



(note 'thermal channeling' in the cool sea case and strong coastal winds on the north coast in both cases)

SST effects on thermal channeling (2)

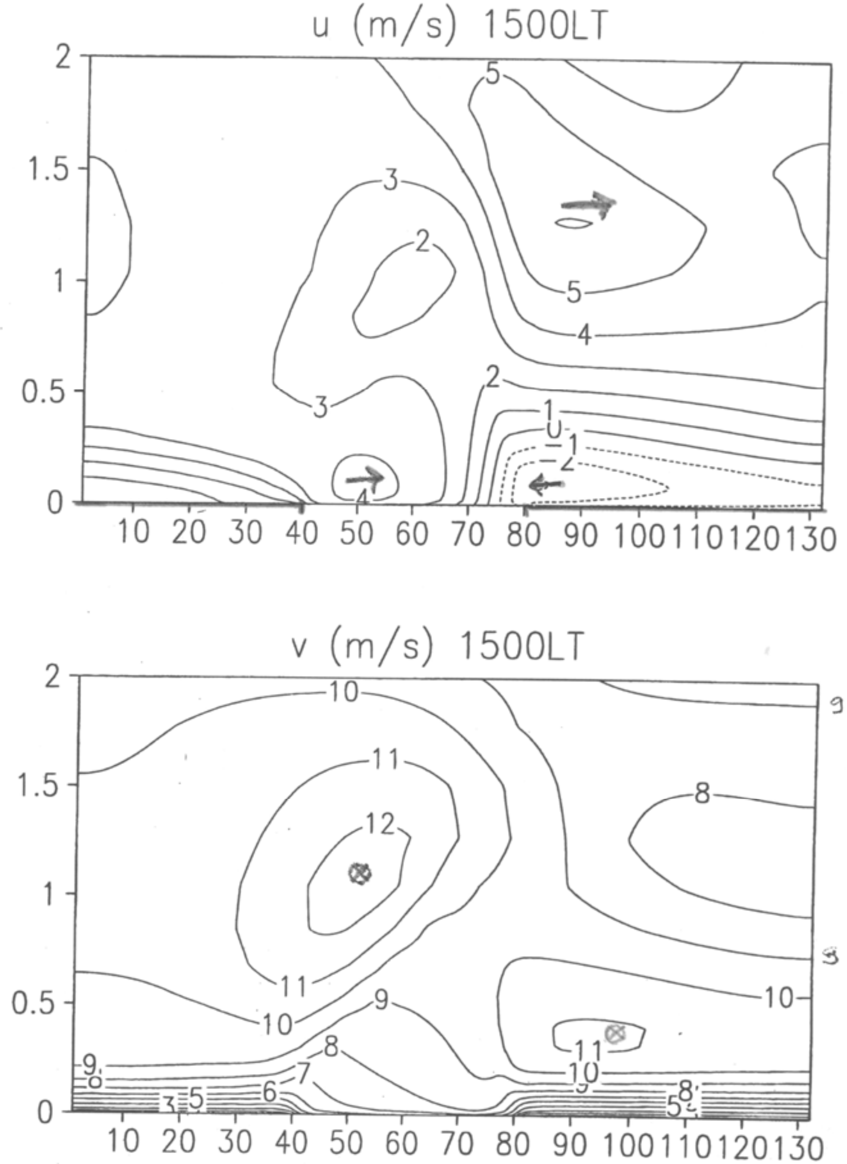
Warm sea, cold air (Early winter, sea not yet frozen)

Calm:

Land breeze cells develop over both coasts. They converge in the middle of the Gulf, creating rising motion. This triggers **lines of convective clouds** and **bands of rain/snow in the middle of the Gulf of Finland.**

Moderate cold outbreak from ESE:

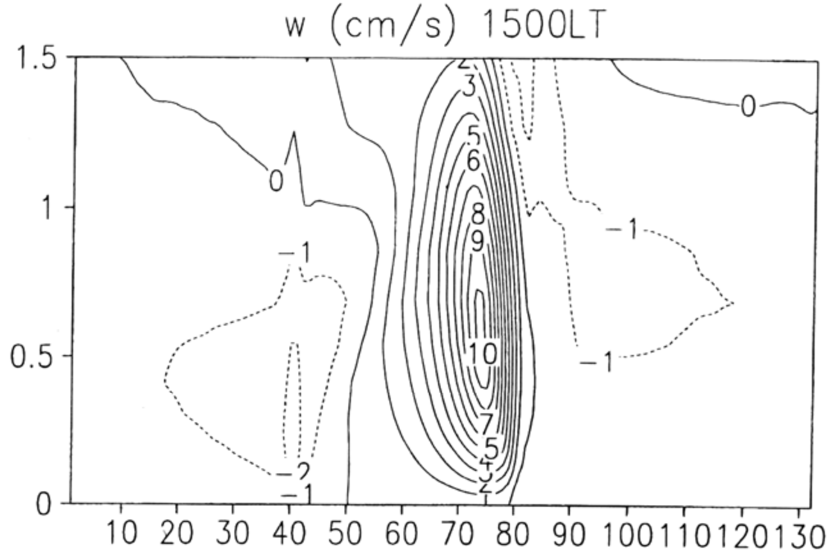
Land breeze cells are now advected to the north by the southerly wind, and are enhanced by the induced heat island circulation => A land breeze front sets near the north coast with strong easterly winds and strong rising motion => **heavy snow showers along the north coast**



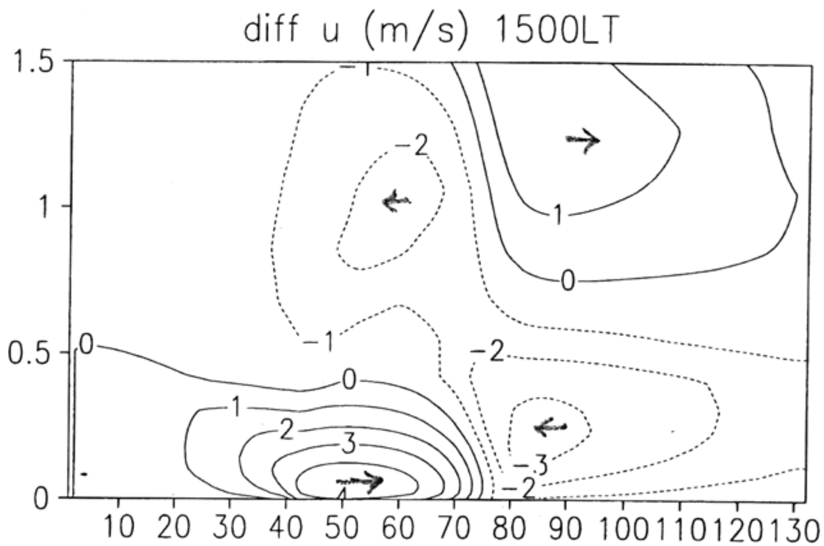
Idealized January cold air outbreak from ESE along open Gulf of Finland. SST +1C, air –16C, Vg 10 m/s.

Finland (north) on the right, warm Gulf at grid points 40-80,
(ug, vg) = (+4, +9) m/s

Note land breezes in u and strong values of v at the sea next to the north coast



Strong rising motion forces line convection and bands of heavy snow along the north coast



Difference of u from its inflow profile (perturbation u') clearly shows two land breeze cells converging strongly in a 'land breeze front' near the north coast

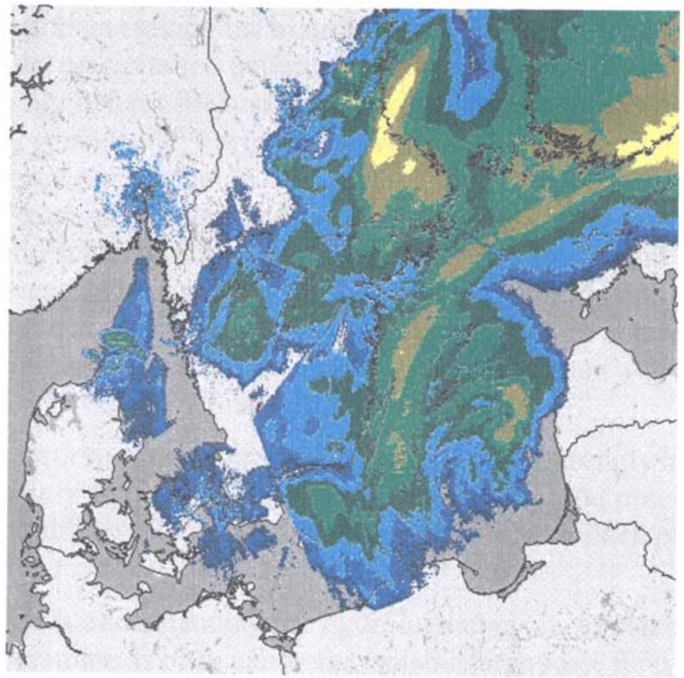
(A linear-model-like heat island circulation pattern, nonlinearly amplified on the north coast)

17-18 Jan 2006 was probably such a case. Snow bands and roll vortices were observed along the Finnish coast with strong easterly winds. The UH Doppler radar observed a convergence zone 10 km off the coast.

\- - -

The 6-8 Dec 1998 Gävle snowstorm during arctic north-easterly outbreak along the non-frozen Gulf of Bothnia (which runs roughly SSW-NNE) may have been a similar case.

A band of heavy snowfall was seen by radar at the sea along the Swedish coast. It hit the curving coastline at Gävle continuously for three days, accumulating more than 1 m of snow, paralysing the city.



Accumulated precipitation from the Radar composite between the 6th and the 7th of December 1998.

Summary

1) Topographic gaps can make air to converge and to increase speed in order to conserve mass, especially during stable conditions (cold sea, cold land)

2) Heat islands create local direct circulations, forming downstream warm plumes during basic flow.

3) Warm/cold slopes tend to create uphill/downhill flows

4) Sea breezes blow toward warm, sunny land (day-time), land breezes toward warm sea (night; winter). They may interact strongly with the large-scale flow.

5) Sea gulfs: On top of all the above: inertial low level jets, thermal channeling, quasi-stationary fronts, snowstorms,.. Coasts provide a rich plethora of interacting meso-scale phenomena.