ISSN No. 0737-5352-9

The Sea-Land Breeze as Local Wind, the Numerical and Analytical Approach to its Modeling

Giovanni A. Dalu

CIRA Cooperative Institute for Research in the Atmosphere

Colorado State University

INTERNATIONAL CENTER FOR THEORETICAL PHYSICS - TRIESTE

16-20 MAY 1988

WORKSHOP ON MODELING OF THE ATMOSPHERIC FLOW FIELD

The sea-land breeze as local wind, the numerical and analytical approach to its modeling

Lecture by Giovanni A. Dalu

Cooperative Institute for Research in the Atmosphere, CSU
Fort Collins, Colorado 80523, U.S.A.

Institute for Atmospheric Physics, CNR
00144 Rome, Italy

SUMMARY

These lectures on the sea-land breeze wind are organized in the following way:

1. A general description of the sea-land breeze phenomenon and of its impact on the human activities.
2. A first theory based on energy consideration, approachable with pocket calculators.
3. An analytical theory approachable with desk computers.
4. A numerical approach to modeling with large computers.

1 - A description of the sea-land breeze wind

The sea-land breeze wind is common atmospheric phenomenon in coastal regions and it is typical of fair weather conditions and therefore more frequent in the summer season. The sea breeze is a diurnal wind, which affects a depth of the atmosphere of the order of 1000 meters for a horizontal extension of the order of 100 kilometers. The observed wind intensity is of the order of 6-10 meters per second. The land breeze is mainly a nocturnal wind and it is much shallower and less extended horizontally then the sea breeze. Its depth being of the order of few hundreds meters with a horizontal extension of about 70 kilometers or less. Its intensity, weaker than the sea breeze, is in the range of 3-5 meters per second.

However, in some places, because of the high depth of the boundary layer and the low latitude, like for instance in the gulf of Carpenteria in Australia (Smith, et al., 1972), the sea breeze is particularly spectacular and the above mentioned figures for observed wind intensity and inland penetration are there more than double.

The sea breeze was known to the ancient sailors like the Greeks and the Fenicians (Herodotus, 485-425 B.C.), as a friendly and convenient wind. Its gentle intensity and its duration, from few hours after sunrise to sunset, well suited the frail vessels of those times. Nowadays the sea breeze is highly appreciated by leisure sailors.

The sea-land breeze play an important role in stirring the coastal waters (Mizzi and Pielke, 1972, Dalu, et al., 1989) and its very important because it keeps the coastal waters well oxygenated right where there is the breeding site for a lot of species. Furthermore, the coastal currents and the vertical mixing within the diurnal thermocline (Dalu and Purini, 1980, 1981) are crucial for the dispersion of the abbondant pollutants in coastal waters close to large ports or heavily industrialized and urbanized areas.

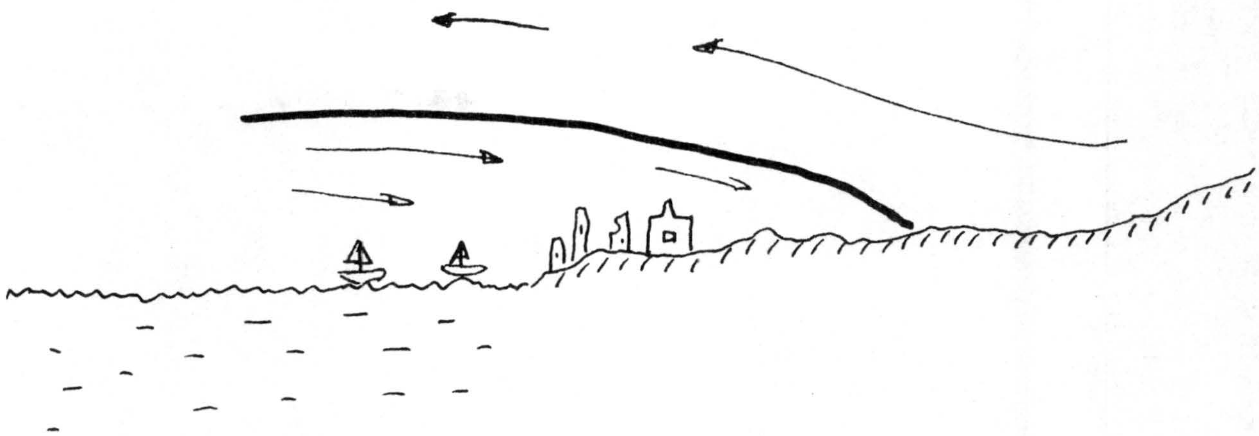

Figure 1. The maritime air penetrates inland proceeding with a speed of the order of 5 meters per second for about 100 kilometers, with a depth of the order of 600 meters. The return flow above is weaker and deeper. The sea air is preceded by a region where the heating is intense. The penetration starts a few hours before sunrise and stops at sunset.

From a meteorological point of view, the weather conditions favorable for the sea breeze are when the synoptic are weak and the days are warm. Also for the land breeze the large scale winds have to be weak, possibly accompanied to a clear sky night, in order to have a fair amount of radiative cooling. These conditions are usually met when an anticyclone, i.e. a high pressure region with subsidence, sits over a coastal area.

2 - An energy model for the sea-land breeze wind

The sea breeze results from the diurnal heating which occurs over land during the sunny hours. As a consequence the air over land becomes warmer and less dense in comparison to the adjent air above the sea, Figure 2, through some depth H.

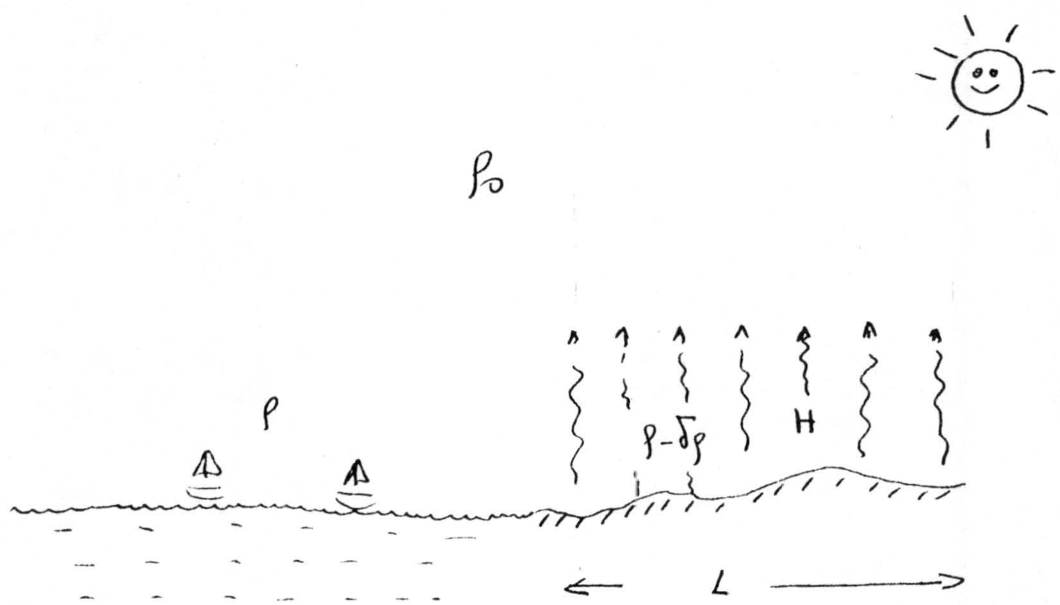

Figure 2. The diurnal heating warms the air and reduces the density of the air over land through a depth H. The free atmosphere density ρ_o is less than perturbed density $\rho-\delta\rho$ which is less than the over the sea ρ.

In the configuration depicted in Figure 2, there is some potential energy available A.P.E. for conversion into kinetic energy K.E.

$$A.P.E. = 1/2\ g'H(HL) \quad (1)$$

where g' is the reduced gravity:

$$g' = g(-\delta\rho/\rho) = g\ \delta\theta/\theta \quad (2)$$

ρ is density, $\delta\rho$ is its perturbation

θ is potential temperature, $\delta\theta$ is its perturbation

H is the depth of the convective boundary layer over land, C.B.L.

L is the horizontal scale of motion, i.e. the land penetration of the sea breeze.

As the potential energy is generated by the diabatic process over land, the air moves converting the potential energy into kinetic energy, as shown in Figure 3.

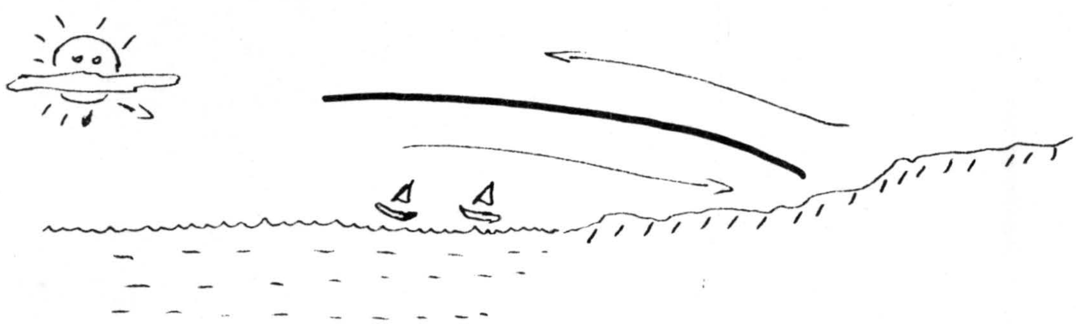

Figure 3. The dense maritime air moves inland subsiding, the warm air over land is lifted and moves towards the sea.

An evaluation of the means intensity of the motion can be done equating the potential energy with the kinetic energy:

$$A.P.E. - K.E. = 1/2\ v^2(2LH) \tag{3}$$

$$v = (g\ H/2)^{1/2} \simeq 7\ m/sec. \tag{4}$$

when $g' = 10^{-1}\ m/sec^2$ and $H = 1000\ m$.

The inland penetration of the sea breeze L stops when the Coriolis force comes into balance with the pressure force, i.e.:

$$fv = \frac{1}{\rho} \cdot \frac{\partial p}{\partial x} \simeq g'H/(2L); \quad f = 2\ \Omega\ \sin(\lambda) = 10^{-4}\ s^{-1} \tag{5}$$

$$L = g'H/(2fv) \simeq 70\ kilometers \tag{6}$$

For the land breeze the inverse process takes place, cooling occurs over land through a depth h of the nocturnal boundary layer, which is typically smaller than the depth H of the diurnal convective boundary layer. The nocturnal configuration is depicted in Figure 4.

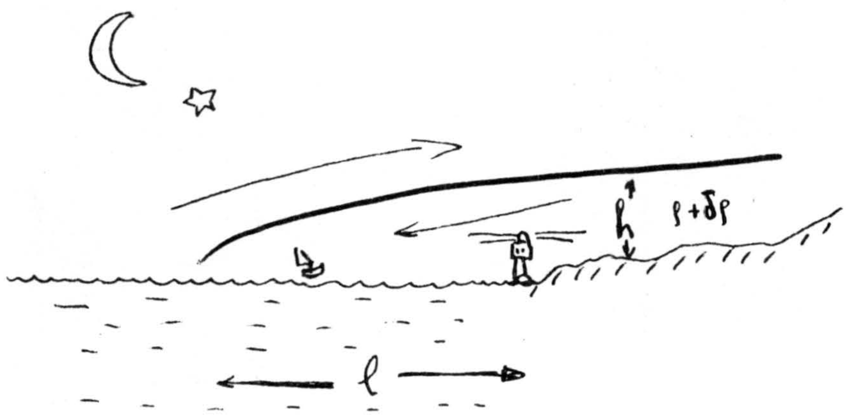

Figure 4. During the night the air over the land cools and becomes denser than the air above the sea. Consequently, the cold air moves towards the sea and the warm air moves aloft towards the land.

The calculation for computing the intensity and the horizontal extension of the induced circulation follow the same pattern:

$$\text{for } h = 500 \text{ m and } g' = 10^{-1} \text{ m/sec}^2$$

$$v = (g\ h/2)^{1/2} \simeq 5 \text{ m/sec and } \ell = g'h/(2fv) \simeq 50 \text{ km} \qquad (7)$$

From (5-6-7) it clearly results that the horizontal scale of motion and the intensity of the flow are smaller during the night because the nocturnal cooling is shallower than the diurnal warming over land. For a more complete study of the sea-land breeze energetics and to its extension to a stratified atmosphere we refer to the following two papers: Green and Dalu (1980) and Dalu and Green (1982).

3 - The analytical theory of the sea-land breeze

From Rotunno (1983) the linearized equation describing the sea land breeze flow are:

$$\left(\frac{\partial}{\partial t} + f\lambda\right) u - f\ v + \frac{\partial}{\partial x} \Phi = 0 \qquad (8)$$

$$\left(\frac{\partial}{\partial t} + f\lambda\right) v + f\ u = 0 \qquad (9)$$

$$\left(\frac{\partial}{\partial t} + f\lambda\right) w - b + \frac{\partial}{\partial z} \Phi = 0 \qquad (10)$$

$$\left(\frac{\partial}{\partial t} + f\lambda\right) b + N^2\ w = Q \qquad (11)$$

$$\frac{\partial}{\partial x} u + \frac{\partial}{\partial z} v = 0 \qquad (12)$$

Where f is the Coriolis parameter, b is the buoyancy force, Φ is the geopotential, Q is the diabatic heat forcing, which warms the air during the day and cools it during the night, $(f\lambda)^{-1}$ is the damping time, due to friction. Defining the stream function:

$$u = \frac{\partial}{\partial z} \psi \; ; \quad w = - \frac{\partial}{\partial x} \psi \qquad (13)$$

the primitive equations equations (8-12) can be reduced to the equation for the stream function:

$$\left[\left(\frac{\partial^2}{\partial t^2} + f\lambda \right)^2 + N^2 \right] \frac{\partial^2}{\partial x^2} \psi + \left[\left(\frac{\partial^2}{\partial t^2} + f\lambda \right) + f^2 \right] \frac{\partial^2}{\partial x^2} \psi = - \frac{\partial}{\partial x} Q \qquad (13)$$

We then define the following adimensional quantities:

$$\tau = ft; \; s = \frac{\partial}{\partial \tau}; \; p = s + \lambda; \; \eta = z/h;$$

$$\xi = f/N \; (1 + p^2)^{1/2} \; x/h$$

$$\text{where } L \{f(t)\} = \int_0^\infty \exp(st) \, f(t) \, dt = \hat{f}(s)$$

$$\text{and } L \, L^{-1} \{f(t)\} = f(t)$$

$$\hat{\psi} = L\{\psi h^{-2} f^{-1}\}; \; \tilde{\psi} = h^{-2} f^{-1}$$

$$(\hat{u}, \hat{v}, \hat{w}) = L\{(u,v,w) h^{-1} f^{-1}\}; \; (\tilde{u},\tilde{v},\tilde{w}) = (u,v,w) h^{-1} f^{-1}$$

$$\hat{b} = L\{b h^{-1} f^{-2}\}; \; \tilde{b} = b h^{-1} f^{-2} \tag{14}$$

$$\hat{\Phi} = L\{\Phi h^{-2} f^{-2}\}; \; \tilde{\Phi} = \Phi h^{-2} f^{-2}$$

$$\hat{Q} = L\{Q h^{-1} f^{-3}\}; \; \tilde{Q} = Q h^{-1} f^{-3}$$

Where h is the vertical scale of motion, which is the depth through which the diabatic heating Q is acting. We denote the Laplace transform with L, with the hat and Laplace transformed variable and with the tilde the corresponding adimensionalized variable.

Assuming that the system is in good approximation hydrostatic, $(\partial^2/\partial t^2 + N^2) \simeq N^2$ and using the definitions stated in (14) for the nondimensional variables and their Laplace transform, the stream equation (13) can be written as:

$$\hat{\psi}_{\xi\xi} + \hat{\psi}_{\eta\eta} = -\beta(p)\,\hat{Q};$$

with

$$\beta(p) = f/N \frac{1}{(1 + p^2)^{1/2}} \tag{15}$$

The Green function for equation (15), i.e. the stream function, resulting from a Dirac δ-function perturbation, is:

$$g_\psi = -\frac{1}{2\pi}\,\ell n \left\{\frac{(\xi-\xi')^2 + (\eta+\eta')^2}{(\xi-\xi')^2 + (\eta-\eta')^2}\right\}^{1/2} \tag{16}$$

From which we deduced, through derivation, the Green function for the horizontal and for the vertical velocities:

$$g_u = \frac{\partial}{\partial\eta}\, g_\psi = \frac{1}{2\pi}\left\{\frac{\eta-\eta'}{(\xi-\xi')^2 + (\eta-\eta')^2} - \frac{\eta+\eta'}{(\xi-\xi')^2 + (\eta-\eta')^2}\right\} \tag{17}$$

$$g_w = -\frac{\partial}{\partial\xi}\, g_\psi = \frac{1}{2\pi}\left\{\frac{\xi-\xi'}{(\xi-\xi')^2 + (\eta+\eta')^2} - \frac{\xi-\xi'}{(\xi-\xi')^2 + (\eta-\eta')^2}\right\} \tag{18}$$

Then the Laplace transform of the horizontal and vertical velocity, in adimensional units, are:

$$\hat{u} = \hat{Q} \cdot \beta(p) \cdot a(p) \tag{19}$$

$$\text{where } a(p) = \frac{1}{2\pi}\left\{\ell n\left[\frac{\xi^2 + (\eta+1)^2}{\xi^2 + \eta^2}\right]^{1/2} + \ell n\left[\frac{\xi^2 + (\eta-1)^2}{\xi^2 + \eta^2}\right]^{1/2}\right\}$$

$$\hat{w} = \hat{Q} \cdot \beta(p) \cdot b(p) \tag{20}$$

$$\text{where } b(p) = -\frac{1}{2\pi}\left\{\tan^{-1}\left(\frac{\eta-1}{\xi}\right) - 2\tan^{-1}\left(\frac{\eta}{\xi}\right) + \tan^{-1}\left(\frac{\eta+1}{\xi}\right)\right\}$$

The horizontal and vertical velocity in adimensional units, through Laplace inversion, are:

$$\tilde{u}(\tau) = \tilde{Q} * \exp(-\lambda\tau) * A(\tau)$$

$$\tilde{w}(\tau) = \tilde{Q} * \exp(-\lambda\tau) * B(\tau) \tag{21}$$

where * denotes convolution product. $A(\tau)$ and $B(\tau)$ are:

$$A(\tau) = L^{-1}\{\beta(s) \cdot a(s)\} = \frac{1}{2\pi} \int_0^\tau du\, J_o\left(\sqrt{\tau^2 - u^2}\right) \frac{1}{u} \cdot \left\{2 \cos\left(\frac{\eta}{\xi} u\right)\right.$$

$$\left. - \cos\left(\frac{\eta+1}{\xi} u\right) - \cos\left(\frac{\eta-1}{\xi} u\right)\right\} \tag{22}$$

$$B(\tau) = L^{-1}\{\beta(s) \cdot b(p)\}$$

$$= - \frac{1}{2\pi} \int_0^\tau du\, J_o\left(\sqrt{\tau^2 - u^2}\right) \frac{1}{u} \cdot \left\{\sin\left(\frac{\eta+1}{\xi}\right) - 2 \sin\left(\frac{\eta}{\xi} u\right) + \sin\left(\frac{\eta-1}{\xi}\right)\right\}$$

$$\xi = \frac{f}{N} \frac{x}{h}$$

Equations (21) gives the required dynamical field for the sea breeze, when used in conjunction of (22) and (14). However it may be more explicatory to give the asymptotic behavior of equations (19-20). If the diabatic heating is a step function:

$$\tilde{Q}(\tau) = \tilde{Q}_o He(\tau) \text{ then}$$

$$\hat{Q}(s) = \tilde{Q}_o / s \tag{23}$$

The asymptotic behavior for $\tilde{u}$ and $\tilde{w}$ is:

$$\tilde{u}(\tau \to \infty) = \lim_{s \to 0} [s\, \hat{u}(s)]$$

$$= \tilde{Q}_o\ f/N \frac{1}{(1+\lambda^2)^{1/2}} \cdot \frac{1}{2\pi} \left\{\ell n\left[\frac{\xi^2 + (\eta+1)^2}{\xi^2 + \eta^2}\right]^{1/2} + \ell n\left[\frac{\xi^2 + (\eta-1)^2}{\xi^2 + \eta^2}\right]^{1/2}\right\} \tag{24}$$

$$\tilde{w}(\tau\to\infty) = \lim_{s\to 0} [s\ \hat{w}(s)] =$$

$$= \tilde{Q}_o\ f/N \frac{1}{(1+\lambda^2)^{1/2}} \cdot \frac{1}{2\pi} \cdot \left\{\tan^{-1}\left(\frac{\eta-1}{\xi}\right) - 2\tan^{-1}\left(\frac{\eta}{\xi}\right) + \tan^{-1}\left(\frac{\eta+1}{\xi}\right)\right\}$$

$$\text{with } \xi = f/N \cdot (1+\lambda^2)^{1/2}\ x/h \tag{25}$$

From (24) we see that the frictional forces reduce the intensity of the velocities and the horizontal scale of motion (25) by a factor of $1/(1+\lambda^2)^{1/2}$. The sea breeze penetration stops when

$$\frac{\xi^2 + 1}{\xi^2} \simeq 1 \tag{26}$$

i.e. when $x > R/(1+\lambda^2)^{1/2}$; $R = h\ N/f$ is the Rossby radius.

The sea breeze is felt inland as far as the Rossby deformation radius corrected by the factor $1/(1+\lambda^2)^{1/2}$.

When the diabatic heat is periodic:

$$\tilde{Q}(\tau) = \tilde{Q}_o \sin(\omega\tau) \text{ then}$$

$$\hat{Q}(s) = \tilde{Q}_o \frac{\omega}{s^2+\omega^2} \tag{27}$$

The behavior of $\tilde{u}$ and $\tilde{w}$ is periodic:

$$\tilde{u}(\tau) = \tilde{Q}_o f/N \sin(\omega\tau) \cdot \left(\frac{1}{1-\omega^2}\right)^{1/2} \cdot \frac{1}{2\pi} \left\{\ell n\left[\frac{\xi^2+(\eta+1)^2}{\xi^2+\eta^2}\right]^{1/2} + \ell n\left[\frac{\xi^2+(\eta-1)^2}{\xi^2+\eta^2}\right]^{1/2}\right\} \tag{28}$$

$$\tilde{w}(\tau) = \tilde{Q}_o f/N \sin(\omega\tau)\cdot\left[\frac{1}{1-\omega^2}\right]^{1/2}\cdot\frac{1}{2\pi}\left\{\tan^{-1}\left(\frac{\eta-1}{\xi}\right) - 2\tan^{-1}\left(\frac{\eta}{\xi}\right) + \tan^{-1}\left(\frac{\eta-1}{\xi}\right)\right\}$$

with $\xi = f/n \cdot (1-\omega^2)^{1/2}\ x/h$ (29)

From (27) we see that periodicity amplifies the intensity of the velocities and the horizontal scale of motion (28) by a factor of $1/(1-\omega^2)^{1/2}$. The sea breeze front stops when

$$\frac{\xi^2 + 1}{\xi^2} \simeq 1 \tag{30}$$

i.e. when $x > R/(1-\omega^2)^{1/2}$; $R = h\ N/f$ is the Rossby radius.

Under periodic forcing, the sea breeze is felt inland as far as the Rossby deformation radius corrected by the factor $1/(1-\omega^2)^{1/2}$. From (30) we see that at 30 degree latitude the penetration becomes infinite, and at lower latitude we don't have sea breeze anymore but waves. These waves are not observed, but sea breezes are observed as far as the lower latitudes. In fact when we take into account friction and periodicity simultaneously, the sea breeze penetration stops when

$$x = R/(1 + \lambda^2 - \omega^2)^{1/2} \tag{31}$$

and the amplitude amplification factor is

$$1/(1 + \lambda^2 - \omega^2)^{1/2} \tag{32}$$

Since $\omega \approx \lambda$ we always have sea breezes below 30 degree latitude, and not propagating waves. Figure (5) shows the structure of the stream function of the sea breeze.

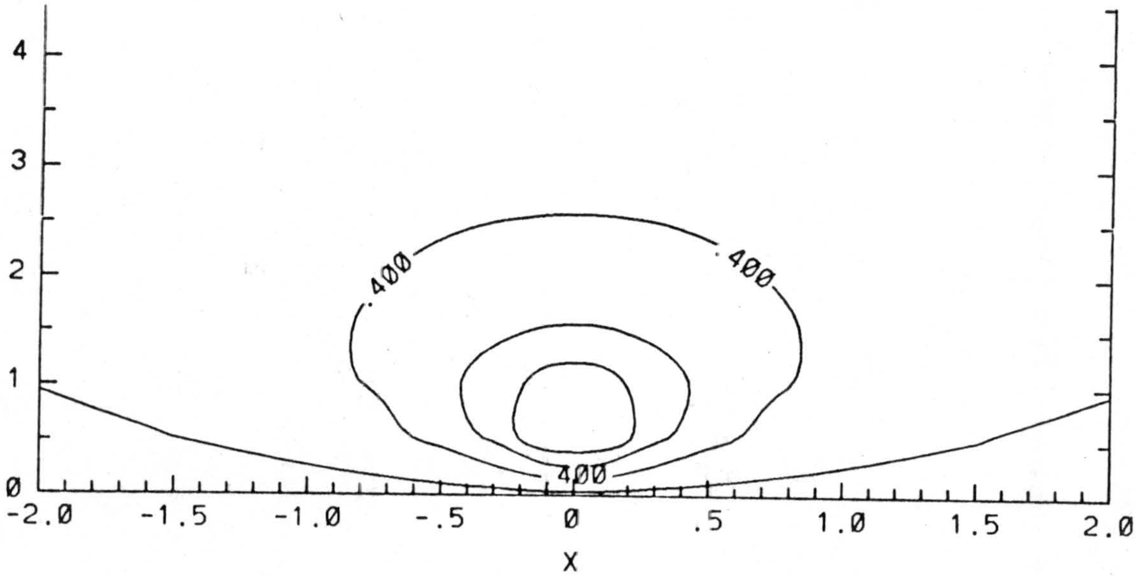

Figure 5. Sea breeze stream function.

4 - A numerical approach to the sea-land breeze problem

There are a number of models which deal with the problem of sea-land breeze, we refer to Pielke model, because is one of the more complete in its kind, and its continuously updated. The basic literature is contained in a series of papers by Pielke and Mahrer (1975-76-77).

The model is a primitive equation mesoscale model. It contains also the large-scale flow, which is deduced from the analysis of the synoptic maps. The large-scale field is assumed not to evolve during the simulation time. The model has an appropriate boundary layer.

Before the simulation itself, a few hours run ensured that the meteorological fields were almost in balance with the lower boundary drag due to the topography and to the surface roughness. Furthermore, the atmosphere, initially barotropic, had to acquire the appropriate baroclinicity corresponding to the observed shear. Simulations start at sunrise and the fields evolve interactively with the fluxes in the lower layers. Heat budget, short- and long-wave radiation, latent-heat and sensible-heat diffusion in the soil and in the atmosphere are included.

The model equations are hydrostatic and in terrain following coordinates:

$$\frac{d\mathbf{v}}{dt} = f\mathbf{k} \times (\mathbf{v} - \mathbf{v}_g) - \theta_v \nabla\pi + g\,\frac{z^* - \bar{s}}{\bar{s}}\nabla z_G - gz^*\frac{\nabla s}{\bar{s}} +$$

$$+ \left(\frac{\bar{s}}{s - z_G}\right)^2 \frac{\partial}{\partial z^*}\left(K_z^m \frac{\partial \mathbf{v}}{\partial z^*}\right) + \nabla K_H \nabla \mathbf{v}, \qquad (33)$$

$$\mathbf{v} = (u,v),\ \nabla = \left(\frac{\partial}{\partial x}, \frac{\partial}{\partial y}\right),\ \frac{d}{dt} = \left(\frac{\partial}{\partial t} + \mathbf{v}\cdot\nabla + w\frac{\partial}{\partial z}\right),$$

where θ_v and π are the potential virtual temperature and the Exner function, respectively.

$$\frac{d\theta}{dt} = \left(\frac{\bar{s}}{s - z_G}\right)^2 \frac{\partial}{\partial z^*}\left(K_z \frac{\partial \theta_v}{\partial z} + \nabla(K_H \nabla\theta)_v\right). \qquad (34)$$

The vertical velocity is computed through continuity:

$$\nabla \cdot \mathbf{v} + \frac{\partial w^*}{\partial z^*} - \frac{1}{s - z_G} (\mathbf{v} \cdot \nabla z_G) + \frac{1}{s - z_G} \left(\frac{\partial s}{\partial t} + \mathbf{v} \cdot \nabla s \right) = 0, \qquad (35)$$

The vertical coordinate is z^*:

$$z^* = \bar{s} \frac{z - z_G}{s - z_G}, \qquad (36)$$

where z_G is the height of the topography at the grid points and $\bar{s}$ is initial total, height of the model.

The material surface s at the top of the model evolves in time to satisfy the vertically integrated continuity equation.

The evolution of soil temperature is described by

$$\frac{\partial T_s}{\partial t} = \frac{\partial}{\partial z} \left(K_z \frac{\partial T_s}{\partial z} \right) \qquad (37)$$

and the land temperature is computed with Newton-Raphson iterative method from the heat balance equation

$$R_s + R_L + \rho L u_* q_* + c_p u_* \theta_* - \rho_s c_s K_s \frac{\partial T_G}{\partial z} - \sigma T_G^4 = 0. \qquad (38)$$

Short-wave radiation R_s, long-wave radiation R_L, latent heat, sensible and soil heat fluxes (Eq. (4), respectively, are treated as in Mahrer and Pielke (5).

The integration scheme is forward in time and semi-implicit. Advection terms are upstream computed with a cubic spline interpolation technic. The flow chart proceeds as in Fig. 6. An example of the use of Pielke's model is given in Figures 7-8 where the sea breeze over Sardinia island is simulated, Figure 9.

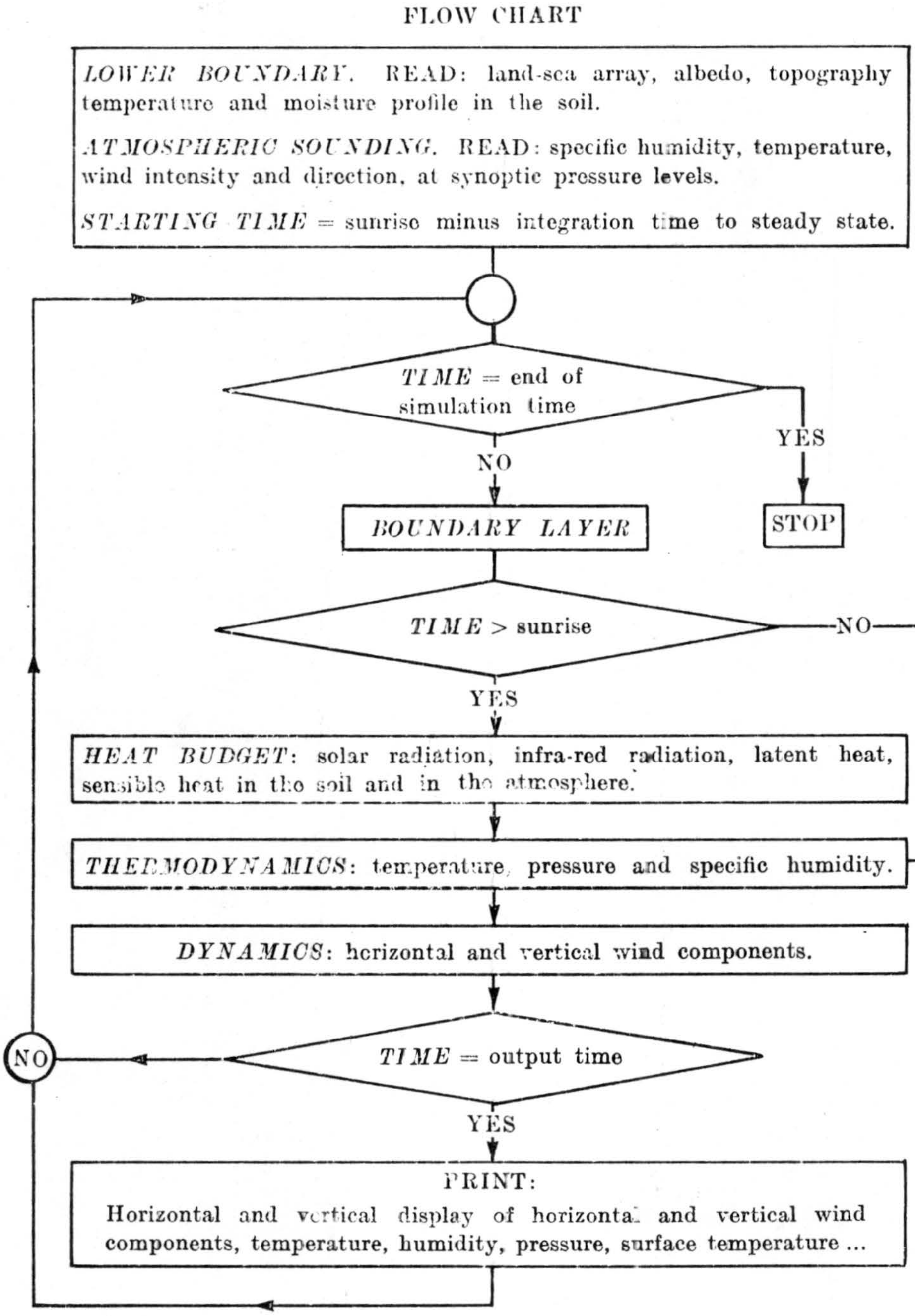

Figure 6. Flow chart.

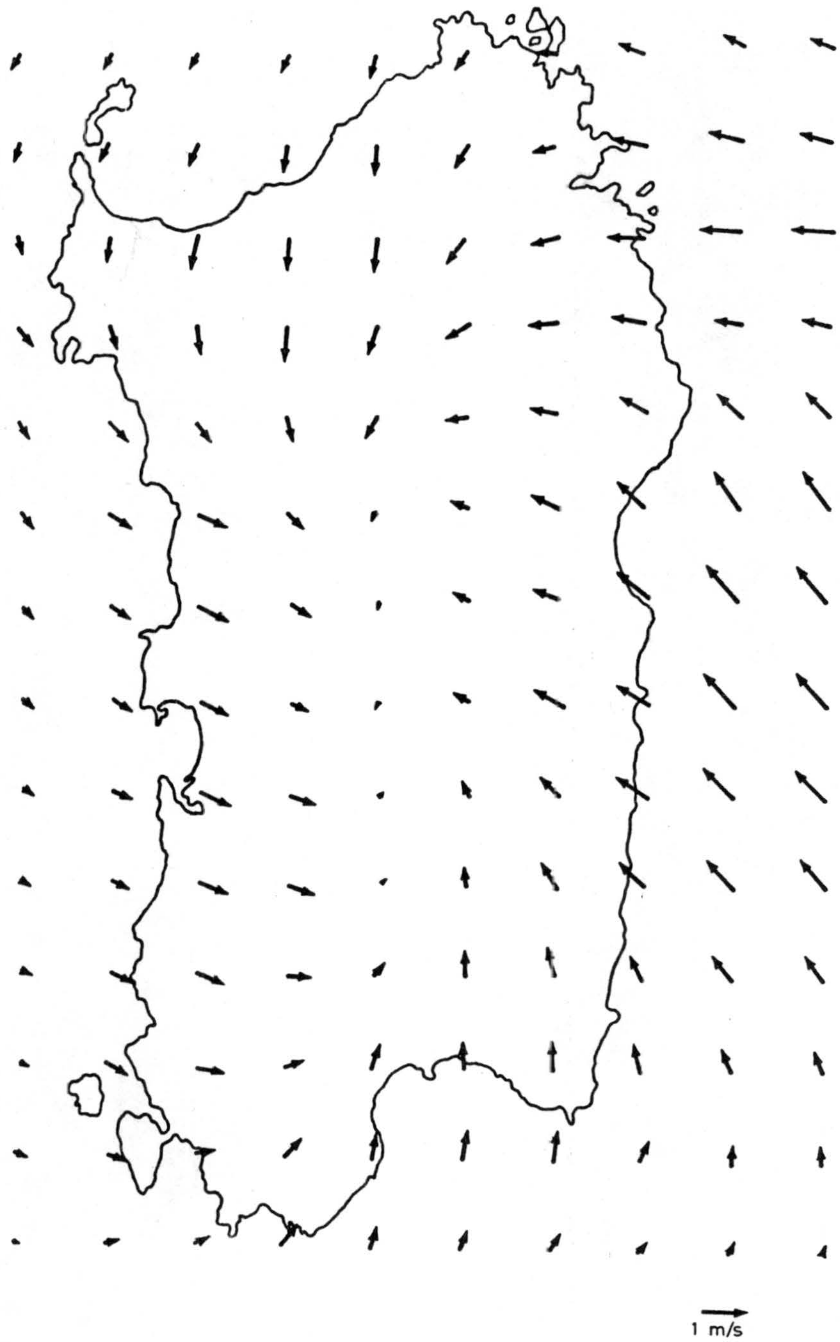

Figure 7. Sea breeze over Sardinia.

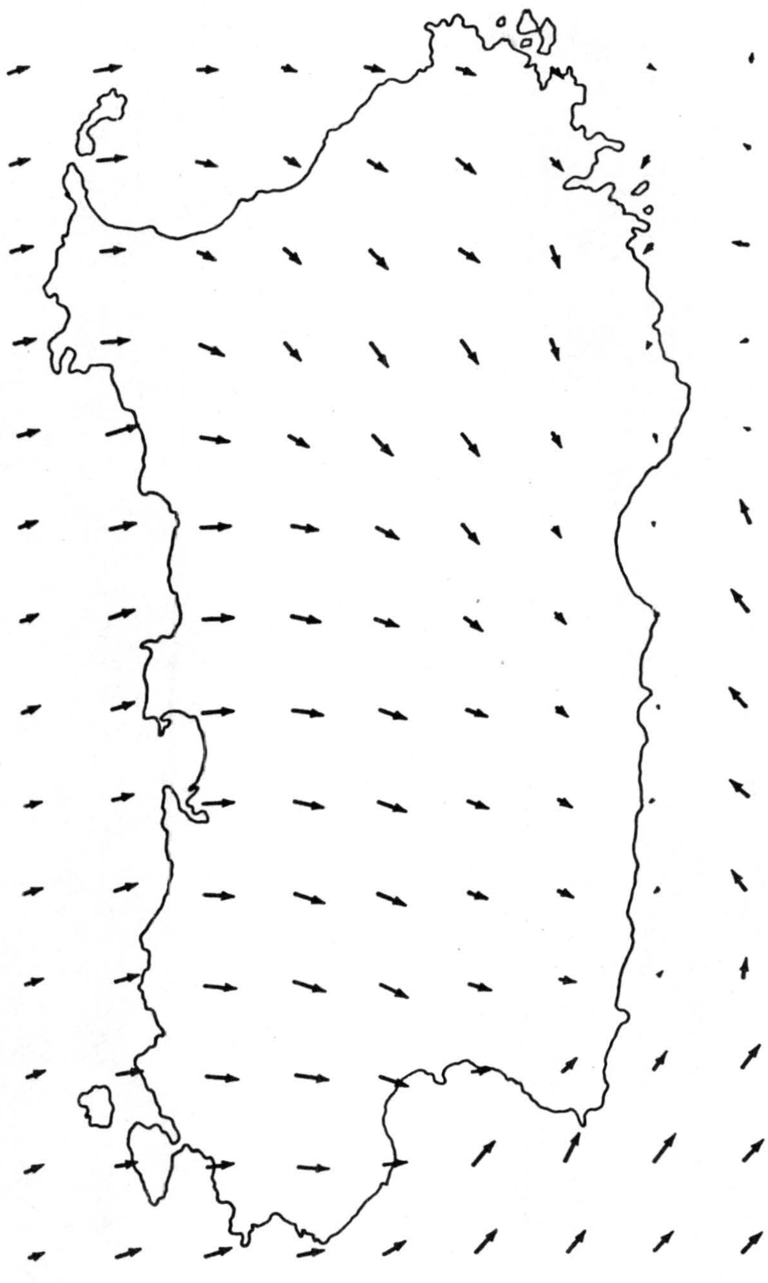

Figure 8. Sea breeze over Sardinia in a north westerly synoptic flow.

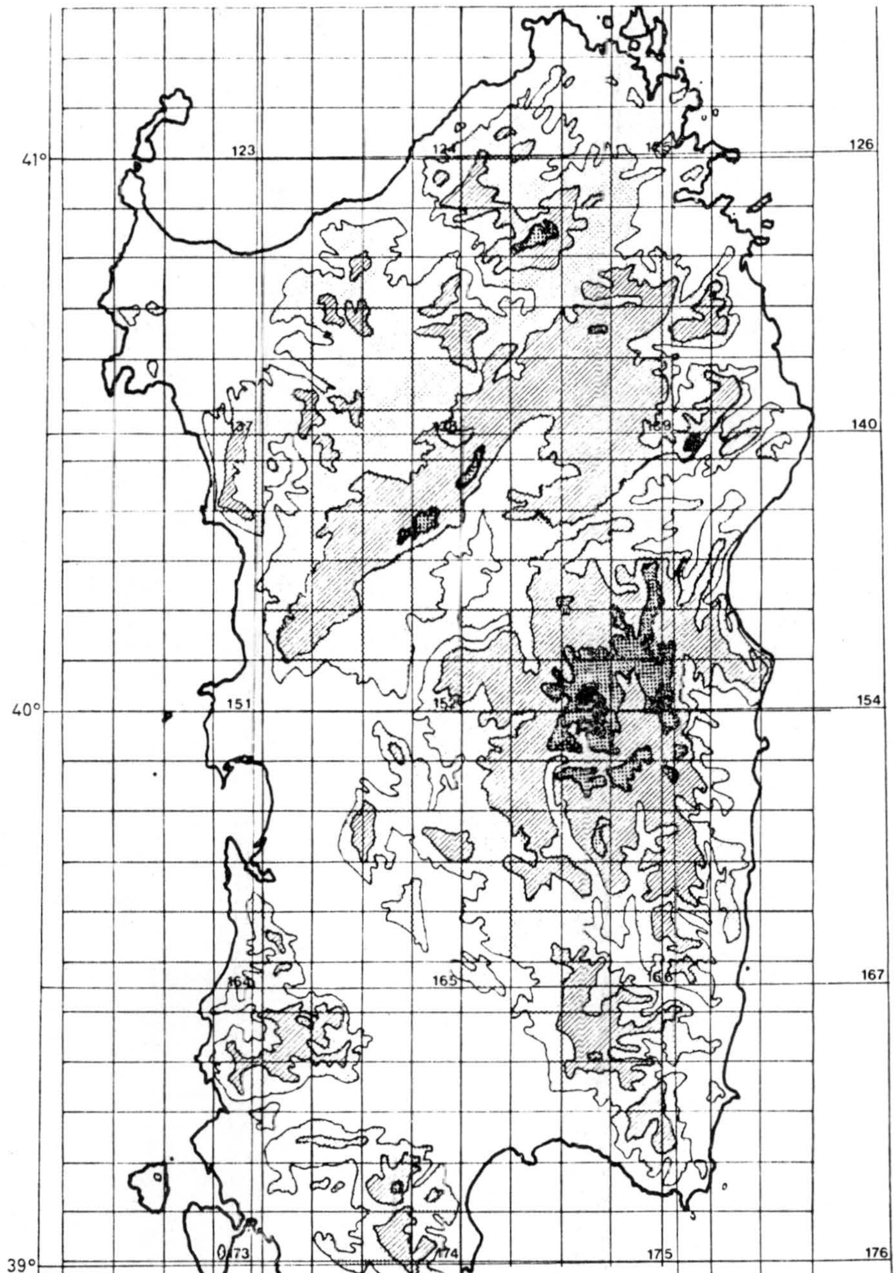

Figure 9. The island of Sardinia and its orography.

How the numerical scheme treats the propagation of waves can be described through the integration of the "shallow water" equations:

$$\frac{\partial u}{\partial t} + U \frac{\partial u}{\partial x} + g \frac{\partial h}{\partial x} = 0 \qquad \frac{\partial h}{\partial t} + U \frac{\partial h}{\partial x} + \frac{c^2}{g} \frac{\partial u}{\partial x} = 0 \tag{36}$$

where the depth h should be identified with the temperature in our set. The integration scheme uses h(i.e. ϕ) at the current time to forecast u, and the new updated value of u to calculate h. This implicit character of the scheme causes wave propagation to be stable, and not heavily damped when the C.F.L. condition $c\Delta t < \Delta x$ is satisfied. Thus the solution can be written $h = \alpha^{t/\Delta t} \exp(ikx)$.

The amplitude α depends on the three parameters $\mu = U \Delta t/\Delta x$, $m = c\Delta t/\Delta x$ and $\theta = k\Delta x$ and is given by

$$[\alpha - 1 + \mu\{1 - \exp(-i\theta)\}]^2 + \alpha(2m \sin 1/2\ \theta)^2 = 0. \tag{37}$$

The limit $\theta^2 << 1$, gives

$$|\alpha|^{t/\Delta t} = \exp - 1/2(1 - \mu \pm m)\theta^2\ Ut/\Delta x. \tag{38}$$

for the amplitude. We notice that dependency on the length of the time-step has nearly disappeared, leaving the advection distance $Ut/\Delta x$ as the dominant term. Stability is ensured if $\mu + m < 1$ and good energy conservation demands $\theta^2 << 1$ at about 60 points per wave-length. Dissipation is a quadratic function f wave length: dissipation for the model with 30 points per wave-length is 4 times that for 60 points (Dalu and Green, 1980).

REFERENCES

Dalu, G. A. and J. S. A. Green, 1980: Energetics of diabatic mesoscale circulation: A numerical study. Quart. J. R. Mat. Soc., 106, pp. 727-734.

Dalu, G. A. and R. Purini, 1980: Un modello numerico della dinamica delle acque costiere accoppiato con un modello numerico di circolazione di brezza. Rivista di Ing. Sanitaria., 5, pp. 235-238.

Dalu, G. A. and R. Purini, 1981: A numerical study of the marine surface layer in a sea breeze regime. Ocean Management, 6, pp. 111-116.

Dalu, G. A. and J. S. A. Green, 1982: An energy theory for the propagation of gravity currents. Mesoscale Meteorology Theories, Observations and Models, 114, NATO-ASI, pp. 211-217.

Dalu, G. A. and A. Cima, 1983: Three-dimensional airflow over Sardinia. Il Nuovo Cimento, 6, pp. 453-472.

Dalu, G. A., M. Baldi and C. Lavalle, 1988: Upwelling induced by periodic wind stress. Il Nuovo Cimento.

Green, J. S. A. and G. A. Dalu, 1980: Mesoscale energy generated in the boundary layer. Quart. J. R. Met. Soc., 106, pp. 721-726.

Mahrer, Y. and R. A. Pielke, 1976: The numerical simulation of the air flow over Barbados. Mon. Weather Rev., 104, pp. 1392-1402.

Mahrer, Y. and R. A. Pielke, 1977: The effect of topography on sea and land breezes in a two dimensional numerical model. Mon. Weather Rev., 105, pp. 1151-1162.

Mahrer, Y. and R. A. Pielke, 1977: Beitr. Phys. Rev., 105, p. 98.

Mizzi, A. P. and R. A. Pielke, 1984: A numerical study of the atmospheric circulation observed during a coastal upwelling event on 23 August 1972. Part I: Sensitivity studies. Mon. Wea. Rev., 112, pp. 76-90.

Rotunno, R., 1983: On the linear theory of the land and sea breeze. J. Atm. Sci., 40, pp. 1999-2889.

Pielke, R. A. and Y. Mahrer, 1975: Representation of the heated-planetary boundary layer in mesoscale models with coarse vertical resolution. J. Atm. Sci., 32, pp. 2288-2308.

Simpson, J. E., D. A. Mansfield and J. R. Milford, 1977: Inland penetration of sea breeze fronts. Quart. J. R. Met. Soc., 103, pp. 47-76.

Smith, R. K., N. Crook and G. Roff, 1982: The morning glory: an straordinary atmospheric undular bore. Quart. J. R. Met. Soc., 108, pp. 937-952.