Air-Sea Interactions

The interactions between the atmosphere and the ocean consist mainly of the exhcange of mechanical energy, heat, moisture, dissolved and suspended materials between and within the lower layer of the atmosphere and the upper layer of the ocean. They govern in great part the weather changes as well as the major ocean currents. Such a global aspect of the air-sea interaction problem is not of direct interest in the study of a relatively small region, as in the scope of the Belgian National Program, where attention should be focused on smaller scale phenomena such as the local effect of the wind and the exchange of gases and chemical constituents.

In the context of the present mathematical model [Nihoul (1970, 1971)] the air-sea interaction problem consists of establishing the value of the « boundary interaction terms » and the eddy factors which appear in the various equations to be solved. In the vertically integrated equation for the horizontal velocity $\overline{\mathbf{v}}_h$, there is a boundary interaction term due to the surface stress $\boldsymbol{\tau}_s$ and the bottom stress $\boldsymbol{\tau}_b$ per unit water mass

$$(1) \qquad \boldsymbol{\tau}_s - \boldsymbol{\tau}_b = - [\overline{\hat{v}_3 \hat{\mathbf{v}}_h}]_{x_3=-h}^{x_3=\zeta}$$

where v_3 denotes the vertical velocity, x_3 the vertical coordinate, ζ and h the elevation of the mean free surface and of the bottom. An overline refers to mean value (in the K.B.M. sense) and a circumflex to fluctuating quantities. The vertically integrated equation for the concentration of a pollutant also contains boundary interaction terms

$$(2) \qquad [\lambda_3 \frac{\partial \rho_\alpha}{\partial x_3} - m_{\alpha,3}]_{x_3=-h}^{x_3=\zeta}$$

where ρ_α denotes the mean (in the K.B.M. sense) of any state variable measuring the concentration of a pollutant, $m_{\alpha,3}$ the vertical component of the « migration » flux of α with respect to the water masses and λ_3 a coefficient of vertical eddy diffusion.

The local winds govern in great part the state of the sea by generating wind (or sea) waves in the storm area. Hence, the air-sea interaction plays an important role in the determination of the eddy viscosities ν_1, ν_2 and the eddy diffusivities λ_1, λ_2 and λ_3 which account for the mixing of momentum and pollutants. Swell waves (or waves that have traveled out of the generating area) also affect the determination of the eddy factors : their influence should be considered as external data to be introduced into the model.

In the present report, the dynamical problems will be reviewed, with emphasis on the actual state of knowledge on wind stress determination. Preliminary information on the practical problem of collecting the required meteorological data which are available will also be given.

1.- Description of the air-sea interaction

The wind stress acts on the water masses at the air-sea interface in a very complex way. It transfers momentum by generating surface waves, drift current, water surface setup and storm tides, and has an important influence on the transfer of heat and mass (moisture, gases, pollutants) through the air-sea boundary. Many investigations on the subject have been published but there is considerable disagreement in the numerous wind-stress data accumulated so far from field observations and laboratory studies. From the theoretical point of view, most of the mechanisms which govern wind action and wave generation remains to be found. A review of the recent theories will not be attempted here, but a description of the basic concepts and observational evidence will be given.

Measurements show that the structure of the atmospheric boundary layer — the thick layer close to the surface of the earth resulting from the combined action of turbulent friction (affected by the density stratification of the air) and Coriolis force — and, in particular, the distribution of wind with height, is often very irregular, due to a number of complicating factors (for more details, see e.g. Monin, 1970). The lower part of the atmospheric boundary layer is called the surface layer of air. Under simplifying conditions (statistically steady and horizontally homogeneous wind), it is characterized by the fact that the vertical momentum flux and the vertical heat and humidity fluxes remain practically constant with height; the action of the Coriolis force can be neglected. The thickness of the surface layer is tens of meters.

If x_1, x_2, x_3 are the cartesian coordinates and if the x_1-axis direction coincides with that of the surface wind, the time-averaged equations of motion have the form

$$(3) \qquad \frac{\partial}{\partial x_3}\left(\tau_{x_1x_3} + \nu_a \frac{\partial \bar{u}}{\partial x_3}\right) = 0 ,$$

$$(4) \qquad \frac{\partial}{\partial x_3}\left(\tau_{x_2x_3} + \nu_a \frac{\partial \bar{v}}{\partial x_3}\right) = 0 .$$

Here $\bar{u}$ and $\bar{v}$ are the components of the mean wind velocity, ν_a is the coefficient of molecular viscosity of air and $\tau_{x_1x_3} = - \overline{\hat{u}\,\hat{w}}$ and $\tau_{x_2x_3} = - \overline{\hat{v}\,\hat{w}}$ are the components of vertical turbulent momentum flux (Reynold stresses) divided by the air density ρ_a. w is the fluctuation in vertical velocity.

Due to the choice of the x_1-axis, as the underlying surface is approached one has

$$(5) \qquad \lim_{x_3 \to 0} \left(\tau_{x_1x_3} + \nu \frac{\partial \bar{u}}{\partial x_3}\right) = u*^2 = \frac{\tau_s \rho}{\rho_\alpha}$$

$$(6) \qquad \lim_{x_3 \to 0} \left(\tau_{x_2x_3} + \nu \frac{\partial \bar{v}}{\partial x_3}\right) = 0$$

where $u*^2$ is some positive limit. The value $u*$ is called the friction velocity, τ_s is the tangential wind stress at the underlying surface. From the remarks above, it follows that the relations (5) and (6) hold in the whole surface layer.

The effect of density stratification on turbulence decreases as the underlying surface is approached, and there is a layer, the dynamic sublayer, in which the influence of stratification can be neglected. Its thickness changes from several meters in the case of very strong hydrostatic stability or instability to very large value under neutral stratification. In the latter case, the whole surface layer can be « dynamic ». Only a limited number of parameters determine the dynamic properties of the dynamic sublayer : $u*$, ν and the roughness parameters of the underlying surface, first of all the mean height of roughness h_s which, above the sea, depends on the wave heights at the sea surface. For $h_s \leq \frac{\nu}{u*}$, the roughness does not affect the structure of the sublayer and the underlying surface is called dynamically smooth. In this case which only occurs with very light wind, viscous forces dominate in (5), (6), very close to the surface, in the so-called viscous sublayer. For $h_s \gg \frac{\nu}{u*}$, the turbulent stresses dominate and the underlying surface is called dynamically rough (the land surface is always such).

For $x_3 \gg \frac{\nu}{u*}$, h_s , the dynamic parameters of the sublayer are determined by moderately small scale turbulence and depend only upon the friction velocity $u*$. Dimensional arguments show that

$$\frac{\partial \bar{u}}{\partial x_3} = \frac{u*}{\kappa \; x_3} \tag{7}$$

where κ is the von Karman constant for which measurements in this layer give a value about 0.4 . By integration, one gets the well-known logarithmic profile

$$\bar{u}(x_3) = \frac{u*}{\kappa} \; \ln \frac{x_3}{\eta} \tag{8}$$

where the constant η is the « roughness length » or « dynamics roughness » which defines the virtual origin of the logarithmic velocity profile and does not depend upon x_3 (nor upon the stratification). The law (8) has been verified in most oceanic observations near (but not very close to) the water surface. Above the sea, η depends on a number of factors such as the average wave height, but mainly on the local wind field. In coastal regions, the transition in wind regime (above land and above sea) and the modification of the wave properties by important depth effects can introduce additional complications. The determination of the roughness length η is intimately connected with the estimation of the drag coefficient C_{x_3} (dimensionless), which is defined as the ratio of the turbulent stresses $\rho_a u*^2$ to $\rho_a \overline{u}^2$. One has

$$C_{x_3} = \frac{\tau_s \rho}{\rho_a \quad \overline{u(x_3)}^2} = \frac{\kappa^2}{(\ln \frac{x_3}{\eta})^2} \tag{9}$$

x_3 being a convenient height which is usually taken to be 10 meters above the mean water level. As τ_s occurs in equation (1) of the mathematical model, the determination of C_{10} is of considerable importance.

A related problem is to establish which portion of the surface wave spectrum supports the bulk of the wind stress, under given wind conditions and hence characterizes the roughness of the surface. The importance of the short gravity waves and the capillary waves (or ripples) on the transfer of momentum from the wind to the waves has been demonstrated clearly. According to Phillips (1966), the momentum flux to the longer waves (whose phase velocity c is larger than $5\ u*$) is always a small fraction ($\simeq 10\ \%$ at most) of the total momentum transfer to the water surface. Shorter waves with steeper faces, rather than longer waves with flatter shapes cause the drag resistance to air flow.

Many authors distinguish three boundary layer flows : aerodynamically smooth flow, transition flow and aerodynamically rough flow. For a wind speed U_{10} less than 3 m/s , the flow is aerodynamically smooth. There is a thin viscous sublayer near the sea surface and the excitation of very short gravity waves is the dominant mechanism at very light wind. As the steepness of the short gravity wave increases, the crests become sharper, which gives rise to capillary waves ahead of the disturbance. The surface roughness is then constituted by ripples, with steeper faces than the gravity waves. For wind velocities $3\ m/s < U_{10} < 10\ m/s$, the flow is in the transition region. White caps appear near the upper wind velocity limit of this region. For $U_{10} > 10\ m/s$, the flow is aerodynamically rough. The viscous sublayer is disrupted by surface roughness (which depends mainly on the short waves) and flow separation from the roughness elements (the short waves) occurs. After the occurrence of wave breaking, the surface roughness is constituted by the basic short gravity waves which receive momentum from the wind stress. Recent studies (Mollo-Christiensen, 1970) suggest that wave generation is intermittent and takes place as a hierarchy of strong non-linear interactions between capillary and gravity waves of different wave length. As shown theoretically by Longuet-Higgins (1969), the ripples still play an important role in supporting the wind stress and transmitting it to the larger dominant waves in the form of a tangential stress unevenly distributed at their crest. At very high wind velocity, wave breaking and whitecaps are very frequent. Momentum is transferred from high frequency components to low frequency waves so that capillary and short waves still play an important role. The effect of the shorter waves at all wind velocities explains in particular the modification of the air-sea exchange when a slick covers the water surface (see below).

2.- <u>Wind stress evaluation</u>

On the basis of a compilation of many recent studies, we suggest the following approximation formulae for the drag coefficient

breeze : $U_{10} \leq 1$ m/s ,

$$C_{10} = 1.25 \; U_{10}^{-\frac{1}{5}} \; 10^{-3} \; ; \tag{10}$$

light wind : $1 \text{ m/s} < U_{10} < 16$ m/s ,

$$C_{10} = 1.4 \; 10^{-3} \; ; \tag{11}$$

strong wind : $U_{10} \geq 16$ m/s ,

$$C_{10} = 2.6 \; 10^{-3} \; . \tag{12}$$

For U_{10} larger than 30 m/s , there is little available data. In hurricanes, values ten times larger than (11) have been reported, but the observations lack accuracy. The empirical formulae (10), (11), (12) are rather approximate and should be used with much care. In particular, in all experiments value (11) has a very large standard deviation, due mainly to the strong influence of the stability conditions. Indeed, under very stable or unstable thermal stratification, the thickness of the dynamic sublayer is smaller than 10 meters so that the logarithmic velocity profile (8) entering in (9) should be corrected. Also, the wind fetch (length of the wind field) and duration add to the dispersion of the drag coefficient values. Many investigators (e.g. Wu, 1969) have suggested that C_{10} increases with wind velocity (as $U_{10}^{\frac{1}{2}}$) in the range (11). However, this law was established by comparing a great number of experimental investigations without attempting to examine their respective accuracy. More careful experiments using only thrust, sonic or hot-wire anemometers on stable plateforms or stabilized buoys (e.g. Smith, 1970) suggest that the simpler law (11) is more relevant. In the range (12), the stability conditions have less influence and, as the standard deviation is smaller, the estimated value (12) is more reliable. In conclusion, the values (10)-(12) of the drag coefficient have relatively good accuracy, but could be slightly modified in the course of the elaboration of the model. They are, anyhow, more realistic than in most mathematical models used so far (e.g. Hansen, 1966; who used $C_{10} = 2.6 \; 10^{-3}$ for all wind speeds).

Very few studies have been made to take into account the effect of rainfall on the drag coefficient. Acting similarly to sea spray, a heavy rainfall appears capable of increasing considerably the wind stress on the sea surface.

Caldwell and Elliott (1971) have shown recently that the additional stress produced by rainfall τ_R obeys approximately the relation $\rho_R = 1.6\ R\ U_{10}^{-1}\ \rho_s$, if the rainfall rate R is in cm/h and U_{10} in m/s and if $C_{10} = 1.2\ 10^{-3}$. Thus, a rain of several centimeters per hour in winds of several meters per second could produce stresses comparable in magnitude to the wind stress.

Emphasis was given above on the effect of capillary waves on the wind wave generation and thereby on the wind stress τ_s in order to reveal more clearly the effect of the presence of an oil slick on the water surface. Such an artificial slick (e.g. soap, detergent or oil) reinforces the damping effect of viscosity on the waves and modifies the surface tension of the fluid. As a consequence, capillary and very short waves cannot develop so easily, which explains the smooth aspect of oil slicks on the sea surface. A much stronger wind is needed to induce the appearance of wind waves. As the surface remains aerodynamically smooth for larger wind velocity, the validity of the drag coefficient estimation (10) should be extended to a much large range, up to about $U_{10} = 6$ m/s . Under moderate wind, the drag coefficient should be approximated by the formula

moderate wind : $6 \text{ m/s} < U_{10} < 14 \text{ m/s}$,

(13) $$C_{10} = 1.1\ \ 10^{-3} .$$

Again, this is only a rough approximation and many factors, such as the extension of the surface slick, have a considerable influence (e.g. Van Dorn, 1953). No data have been found so far for the case of higher wind velocity.

Wind action on the sea produces, via friction, a so-called drift current in the upper layer of the sea. Wind generated currents have been studied theoretically for a long time, particularly since Ekman's work on the subject. However, it appears that the observational evidence is conflicting so that, again, no definitive theory can be used. The direction of the drift currents differs generally from the direction of the tangential wind stress by an angle of $\sim O(15°)$ to the right of it, depending on many factors such as the water depth, the presence of coasts, the wind speed, ... The velocity of this current is generally one order of magnitude less than the friction velocity $u*$. From observations made at five lightvessels in the southern North Sea and eastern English Channel, Veley (1960) found in fact that the angle of deflection between wind and current was usually less than $20°$, though it changed with the direction of the wind owing to the presence of nearly coasts and submarine topography which is very important in such shallow water. He also found that the speed of the drift currents range from 0.8 to 2.4 % of the wind speed. In the scope of the present mathematical model, however, such observations can only be used to adjust and check the different parameters of the model. Wind drift currents will be calculated numerically, using the wind stress value as input data.

3.- Geostrophic wind

In principle, it is possible to deduce the surface wind from the geostrophic wind which is calculated from the field of atmospheric pressure. Under steady conditions, the wind velocity at the ground or sea surface (say $\mathbf{U}_{10}$) can be deduced from the geostrophic wind $\mathbf{U}_g$, knowing the latitude and the frictional forces due to turbulence in the atmospheric boundary layer, depending upon surface roughness and stability conditions. A few empirical relations has been suggested but the scatter of the observations is very large. Over the oceans at our latitude, the mean angle between $\mathbf{U}_g$ and $\mathbf{U}_{10}$ is approximately 15 degrees; the value of the ratio of their magnitude is about 0.65 (e.g. Roll, 1965). However, the spread about these mean values is very large (the presence of lands has also some importance). A convenient practical procedure is to fit best a polynomial relation between $\mathbf{U}_{10}$ and $\mathbf{U}_g$ at a given site, using as much data as possible. As an example, let us mention the work of Smith (1970) who used this method for a coastal site of the East Coast of Canada (latitude $44°$). He found that the average deviation of U_{10} was $28°$ to the left of the geostrophic wind, with standard deviation $8°$. Also,

$$U_{10} = 4.8 \text{ m/s} + 0.4\, U_g \tag{14}$$

with standard deviation 2.1 m/s . However, the conclusion of his computation is that direct measurements of U_{10} are much more reliable, in particular because the geostrophic wind can hardly be determined with the necessary accuracy. When the isobars are strongly curved, gradient winds are used instead of geostrophic winds, which brings in additional uncertainties. Some work has also been done to take into account the time variation of the pressure gradients.

The calculation of the surface wind from the geostrophic wind is mainly of interest when there is no direct information on the wind direction and strength over the sea surface, but this is not the case of the region under study in the Belgian National Program (see below). In some cases, however, this procedure could be useful. The best example is when, for some reason (e.g. accidental oil slick on the sea surface), a forecasting is needed. As the recorded wind data provide no informations of the expected meteorological evolution, the use of the weather forecasts (made up to 72 hours in advance) could provide an at least rough indication of the evolution of the pressure field and thereby of the expected geostrophic and surface wind. Though very unusual, such circonstances might be important and the possibility of using the meteorological prediction charts should be kept in mind. Wind and pressure data for the test region will be accumulated in order to establish an empirical relation analogous to (14). Such a procedure will not, however, be used currently (even in a latter stage of the model), but only when forecasting is needed.

4.- Bottom stress evaluation

Though the bottom stress evaluation is not directly related to the air-sea interaction problem, there is a parallelism between wind and bottom effect, both in the physical aspect (turbulent boundary layer) and in the way they enter into the mathematical model [see equation (1)]. Thereby, the subject will be mentionned briefly.

The frictional force on the sea bed is often given in the form, similar to (9),

$$(15) \qquad \tau_b = k \; \mathbf{v}_b \; |\mathbf{v}_b|$$

where $\mathbf{v}_b$ is the mean current measured at a specified height above the bottom, ρ the water density and k the friction coefficient. Dimensional analysis can be made in a similar way as for the atmospheric boundary layer, and a region of constant Reynolds stress can be defined; expression (15) follows.

Comparison with observations leads to an empirical value for the coefficient k . Again, there is a dependence on the bottom roughness, the bottom material, the degree of turbulence, the presence of suspended sediments, the depth, ... In the present mathematical model, however, only the vertically integrated mean velocity $\mathbf{V}$ is considered. Still, one usually assumes that the bottom stress is proportional to the squared integrated velocity. It also depends upon the many factors listed above so that only empirical formula can be used. It is generally assumed that

$$\tau_b^{(1)} = r \, \frac{\mathbf{V} \, |\mathbf{V}|}{(h + \zeta)^2} \, . \tag{16}$$

According to Hansen (1962), a value $r = 3 \; 10^{-3}$ is applicable in estuaries and in open seas as well as in the ocean. Such a value will be used first and eventually adjusted in the elaboration of the model. It has been noted indeed by Leendertse (1967) that, as the exact expression of the friction coefficient must be established in an iterative manner by comparing computed results with actual field measurements, the coefficients used were in fact influenced by the grid size, the time step and the approximations in the numerical model. This problem becomes a problem of numerical analysis and goes beyond the scope of the present report.

It should be mentioned that in case of zero or negligible volume transport of water (e.g. in case of opposite action of wind and tides), there is still a stress exerted by the bottom. It has been suggested (e.g. Groen and Green, 1962) to add to formula (16) a term proportional to the wind stress, of the form

$$\tau_b^{(2)} = - m \, \tau_s \tag{17}$$

where m is a factor much smaller than unity to be determined experimentally and numerically. This allows to rewrite (1) in the form

$$\tau_s - \tau_b = \tau_s (1 + m) - \tau_b^{(1)} \, . \tag{18}$$

It is easier not to consider $\tau_b^{(2)}$ or m (which is very small) but to allow for an eventual adjustement of the drag coefficient included in the expression for τ_s .

5.- Available meteorological informations

A preliminary list of the existing meteorological stations in or near the border of the region under study has been established. Most available informations are taken from anchored ships (lightvessels) or coastal stations. The approximate location of the stations which are operational now and whose informations are regularly received by the *Régie des Voies aériennes* is represented in figure 9.

fig. 9.- Location of a few existing meteorological stations (☆ refers to light-vessels, © to coastal stations).

Most of these meteorological observations at sea are taken from lightvessels. This has many disadvantages as compared with measurement taken from platforms or moored buoys. First of all, the ships are exposed to the action of the waves. The resulting pitching and rolling motion introduces large uncertainties in the measurements. In particular the readings of cup anemometers are increased so that the measured wind is larger than the real wind. Secondly, the body of the ship causes a considerable disturbance of the air and water flow, and forms a source of convective and radiative heat. The errors due to these factors are very difficult to estimate as they depend upon the ship, the location, the wind speed and the sea state. As an example, it has been reported by Roll (1965) that the mean wind speed (average over 10 minutes) at sea under fair conditions is estimated to be about $\pm 5^\circ$ in direction and ± 0.5 m/s in speed. Under bad conditions, the maximum error may reach $\pm 15^\circ$ and ± 2.5 m/s . To get more accurate informations, fixed constructions or stabilized buoys should be used. Measurements made from lightvessels or anchored ships are however much more reliable than those made from ships under way, as the disturbance caused by a moving ship is more important. Another factor to be considered in handling data is the reliability of the instruments and the quality of the people who make the experiments. Clearly, informations emitted by ships which are not especially equiped for meteorological purposes (as merchant ships) are only of little interest. Among the possible sources of data, the ships of the Belgian Navy should also be mentioned. In particular, the research vessel *Mechelem* is equiped for meteorological purpose and, when anchored, can play the role of a supplementary (but itinerant) lightvessel.

As emphasized above, observations made from platforms, tower or moored buoys are more reliable. A few Belgian buoys will be operated on rather soon as three buoys will be used in the pollution program. Another one is being built by meteorologists. The latter one will possibly be set near the West Hinder lightvessel. Other stations are being built by the neighbouring countries.

No report will be made here on the parameters which are being measured in all these stations nor on the frequency of observations, but the situation seems satisfactory. All informations will be transmitted *via* the *Régie des Voies aériennes* and used as input data of the mathematical model. Preliminary analysis of a few meteorological data indicates a large variability of the wind speed and direction within the test region so that the maximum number of available observations should be used. On the other hand, the atmospheric pressure seems rather uniform (only a few mb of difference). More information is needed before drawing any conclusion on eventual simplification of the model (e.g. by neglecting the atmospheric pressure effects).

6.- Conclusion

In this report, the boundary interaction terms entering into the purely dynamical equations of motion of the mathematical model have been reviewed. Preliminary formulation based on recent work is suggested as a basis for the use in the numerical model. Fitting of the values of the parameters will be made in a later stage, by comparing observations and numerical simulation. The amount of available meteorological data appears already promising for handling the dynamical air-sea interaction problem, and will increase in the following year.

No mention has been made of the chemical aspect of the air-sea interaction problem which is important in the modelling of the pollution problem. A review of the actual state of knowledge will be attempted in a forthcoming paper.

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