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INHOUD

Interactions of Sea and Atmosphere.

by J.A.Businger.

1. Introduction.

Since the previous review of this subject by Pond (1971) a lively activity in air-sea interaction research has been displayed. A number of large scale experiments have and will be carried out since the BOMEX (Barbados Oceanographic and Meteorological Experiment, 1968), notably: JONSWAP (Joint North Sea Wave Project, 1968, 1969, 1973 and probably 1975), ATEX (Atlantic Trade-Wind Experiment, 1969), IFYGL (International Field Year of the Great Lakes, 1972), JASIN (Joint Air-Sea Interaction, 1972 and probably 1977), GATE (GARP Atlantic Tropical Experiment, 1974) and AMTEX (Air Mass Transformation Experiment, 1974, 1975), and others. It will be some time before all the information from these field experiments will have been digested. so we may look forward to continued active reporting in this area for the coming years.

Numerous publications carrying articles in this area have appeared recently. The new Journal of Physical Oceanography has given ample space to air-sea interaction topics as well as the almost equally new Journals of Geophysical Fluid Dynamics and Boundary Layer Meteorology. In 1974 this last journal devoted a special double issue to "air-sea interactions" in memory of the German scientist Karl Brocks. Furthermore the first textbook in the English language has appeared (Kraus, 1972).

No serious attempt will be made here to define what encompasses "air-sea interaction research". It is assumed that the physical processes governing the interface and the boundary layers on both sides of it constitute the central area of interest.
Usually the boundary layers are capped by an inversion in the atmosphere and a pycnocline in the ocean. However, the turbulent boundary layers are not always well defined, particularly when deep convection takes place. In some rare but important cases it is possible that te convection extends from the bottom of the sea to the tropopause.

Because the structure of the boundary layers is an essential part of air-sea interaction some overlap with John Wyngaard's review on "Boundary Layers and Atmospheric Turbulence" will be inevitable. The area of atmosphere - sea ice - ocean - interaction has not been included in this review in the expectation that it will be discussed by Gunther Weller in his report on "Polar Meteorology".

The international experiments and the exchange of scientists have led to substantial international co-operation between scientist of many countries resulting in a number of joint multinational papers. Prime examples of this type of reporting can be found in the results of the JONSWAP experiments by Hasselmann et al. (1973) and in the results of ATEX by Augstein et al. (1973, 1974). Examples of more incidental co-operation are given by Zilitinkevich and Deardorff (1974) and by Businger and Yaglom (1972). This United States report must therefore inevitably have an international flavor.

The bibliography covers essentially the literature of the last four years. References to earlier work can be obtained from the reviews by Pond (1971) and Hidy (1972) and the text by Kraus (1972) as well as from other papers in the bibliography.

In organizing the report an effort has been made to start with the interface itself and the molecular layers adjacent to it, and then gradually to expand the scale of the interactions until the entire ocean-atmosphere system is included. This seemingly logical approach still has shortcomings, and several important contributions got lost in the organization but have been included in the reference list.

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2. The structure of the air-sea interface and its immediate environment.

The variety and complexity of the physical and chemical processes near the air-sea interface has been well illustrated by Mc Intire's (1974) discussion of the top millimetre of the ocean. In his logarithmic display of the ocean the top millimetre represented half of it. We shall restrict ourselves to the physical processes and refer the reader for a discussion of the surface chemistry of the interface to Hidy's (1972) review as well as to discussions by Mallinger and Michelson (1973), Mansfield (1972), Palmer and Berg (1972), Garrett (1971), and Wu (1971a) on monolayers and surfactants.

a. The molecular (diffusion) sublayers near the interface.

Because the sea-surface is largely a continuous material surface, turbulent transfer at the surface itself is being suppressed and molecular processes must take over the transfer of momentum, heat, water vapor, and other quantities. Since this is very slow compared to the turbulent transport farther away from the suface, the sub-layers form the bottle neck in the transfer of properties from sea to air and vice versa. The equivalent thickness of the layers defined by $\delta = D\Delta s/F_s$ (where D is the diffusion coefficient, Δs the difference in concentration across the layer, and F_s the flux-density of property s) and the concentration difference, Δs, are the quantities one would like to know. It is tempting to apply results of high Reynolds number channel flow in wind tunnels to this problem. It is clear that when forced convection dominates $\delta \propto \nu/u_*$, where u_* is the friction velocity and ν the kinematic viscosity (Wu, 1971b). This leads to the expression for the heat flux from below to the surface $F_H = -\gamma\ c\rho_w\ v_*\ \Delta T$, where c is the specific heat of water, ρ_w the density of water, ΔT the temperature difference across the layer, and v_* the friction velocity in the water ($v_* = (\rho_a/\rho_w)^{\frac{1}{2}}\ u_*$), and γ is a constant of proportionality (Saunders, 1973a). Hasse (1971) successfully determined the heat flux using the channel flow analogy. Saunders argued that the analogy between the sea surface and a rigid smooth wall may be valid for the air above the surface but is unlikely to be valid for the water below the surface because of the effect of surface tension. He considered Hass's success to be fortuitous. In a similar vain Omholt (1973) criticised Wu's results. Measurements by Schooley (1971) seem to agree with Wu, but more recent measurements by Mangarella, et al. (1973) disagree with these results.

More detailed studies on the molecular layer below the water surface have been carried out in the laboratory by Hill (1974), Paulson and Parker (1972), and Liu (1974).

Although intuitively important the effect of surface tension has not been clearly demonstrated. Katsaros and Liu were able to measure the temperature profile inside the molecular layer under conditions of free convection driven by an evaporating water surface, (Liu, 1974). The profile fitted the model presented by Howard (1964) accurately. They also found the free convection similarity condition for the heat flux for large Rayleigh numbers to be valid in this case ($Nu = a\ Ra^{1/3}$, where $Nu = F_a\ d\ (k \Delta T)^{-1}$. a is a constant of proportionality, $Ra = g\ \alpha\ \Delta T\ d^3/\kappa\nu$, d is the thickness of the convective layer, k is the thermal conductivity, g is acceleration due to gravity, α is coefficient of thermal expansion, and κ thermal diffusivity).

b. The surface temperature.

From the preceding section it is clear that the surface temperature (i.e. the temperature of the interface itself) is an important parameter to know. Fortunately this quantity can be measured remotely with infrared or microwave radiometers from ship, aircraft and satellite. With care an accuracy of 0.05°C may be obtained with infrared (Saunders, 1973a). This is enough to detect the temperature difference across the molecular layer. It is also possible to map the surface temperature and detect small to mesoscale horizontal variations in it (Saunders, 1972, 1973b). By repeated mapping of the surface temperature a synoptic picture of disturbances may emerge and something can be said about the surface flow. Holladay and O'Brien (1975) report an extensive series of surface temperature mappings off the Oregon coast which enabled them to analyze the effect of the wind on the upwelling in that area. Valuable global information about the sea surface temperature may be obtained from satellites (Shenk and Salomonson, 1972), although the accuracy and certainly the horizontal resolution are less than that obtainable from aircraft.

Since the first successful observations of the heat flux through the ocean surface were reported with the two wavelength infrared radiometer (Mc Alister et al., 1971) no further observations have been mentioned. It is desirable that this technique, although it has a limited accuracy (Katsaros and Businger, 1973), be further developed. Possibly another means of obtaining the heat flux is by measuring the small scale temperature fluctuations of the sea surface. It is likely that the small scale horizontal temperature fluctuations are a measure of ΔT across the molecular sublayer, thus together with an estimate of u_* the heat flux may be obtained (Paulson and Leavitt, 1975).
Supercooling of the surface was reported by Katsaros (1973) of an arctic lead near Point Barrow, Alaska. This observation was followed up by a laboratory study (Katsaros and Liu, 1974), where supercooling of the order of $1^{\circ}C$ was observed prior to freezing. This interesting piece of information may have some relevance to deep convection.

c. Drift, capillary waves and small gravity waves.

When air flows over a smooth water surface the first effect of the shear near the surface is to induce a drift current in the water. This drift is in the range of 2.5 - 4% of the wind speed under turbulent conditons (Wright and Keller, 1971). With increased shear near the surface ripples will form. Stern and Adam (1973) give a theoretical analysis of the minimum shear that generates capillary waves. The phase speed of wind-generated waves is greater than the linear wave theory predicts. Shemdin (1972) found that the phase speed was strongly influenced by the coupled shear flows in air and water. This first order perturbation analysis of the problem gave adequate results, by using simple logarithmic wind- and drift-profiles as measured in a wind wave facility (Shemdin and Lai, 1971). When the waves become more substantial there is an effect on the structure of the turbulent shear flow. Takeuchi et al. (1974) find a pronounced effect of the waves on the variances and covariances of the horizontal and vertical velocity components of the wind above the waves. Other studies on the interactions between the air flow and small scale waves are reported by Valenzuela and Laing (1972) and Chang et al. (1971).

The formation of ripples on the surface modifies the structure of the molecular sublayers. Witting (1971, 1972) has analyzed this problem theoretically and found that capillary waves especially may reduce ΔT of the sublayer substantially and consequently raise the surface temperature. It should be possible to verify this with radiometric observations.

d. Wave generation and dissipation.

The work on wave generation has continued at a high level of activity. A major contribution was made by Hasselmann et al. (1973) describing the JONSWAP experiment. The non-linear weak interaction theory of Hasselmann could be verified for wave-to-wave interactions. This theory describes the rapid growth of the long wave energy with increasing fetch adequately and provides the much needed extension of the existing linear theories. However, a number of questions remain to be answered. It is not quite clear how the momentum from the air is transferred to the waves. The observations by Dobson (1971) of pressure fluctuations at the wave surface itself indicate that a large fraction of the momentum transferred to the water is through the waves. Pressure fluctuations just above the wave crests were measured by Elliott (1972). His results seem to agree with Dobson. On the other hand, more recent measurements by Snyder (1974) (see also Snyder et al., 1974), also above the wave crests, suggest a much weaker transfer of energy by pressure fluctuations to the wave. However, the extrapolation from observations above the wave crests to the wave surface requires assumptions which have not yet been verified. Although there is considerable evidence that most of the momentum is transferred from the atmosphere to the waves, especially in a growing wave-field, there is not yet a model describing how the atmosphere does it. Stewart (1974) sketched a three-dimensional flow over water which would be very effective in transferring the momentum to the waves. He also argues that the momentum must first be transferred to the waves before it is absorbed in the current. In a more theoretical analysis Lee (1973) shows that wind profile distortion as a result of non-linear wind-wave interaction may enhance the momentum transfer from wind to waves. Experimental evidence of wave-induced fluctuations in the air-flow have been reported by Davidson and Frank (1973), and Takeuchi et al. (1974).

On the other hand, the variability of the wind (Manton, 1972) and the intermittency of the stress (Davis, 1972, and Dorman and Mollo-Christensen, 1973) may contribute significantly to wave generation and wave growth. Much remains to be resolved in the area of interaction between waves and wind.
Considerable interest has been displayed for breaking waves and white caps. Monahan (1971), but expecially Kondo et al. (1973), and Ross and Cardone (1974), showed that the percentage of the surface area covered by white caps is considerably less than that which was reported by Blanchard (see Kraus, 1972).
The wave generation mechanism proposed by Longuet-Higgins (1969), that small breaking waves add their momentum to the larger waves underneath and thus transfer energy from small to large waves ("maser" mechanism), has been refuted by Hasselmann (1971). The work done on the long waves by the interaction stresses is almost exactly balanced by the loss of potential energy arising from the mass transfer. In a recent paper Hasselmann (1974) gives a more complete account of the non-linear effects of breaking waves. Other recent studies by Banner an Phillips (1974) and Phillips and Banner (1974) deal with the maximum amplitude a wave can attain when it is at the point of breaking, and by Lemmin et al. (1974) with the structure of turbulence generated by breaking waves.

e. Miscellaneous studies of and near the interface.

A number of remote sensing techniques have been applied to gain information about the shape of the water surface. Tober et al. (1973) used a laser instrument for the detection of water ripple slopes with good success in the laboratory. Schule et al. (1971) determined wave spectra with an airborne laser. Levanon (1971) determined the sea surface slope distribution and wind velocity using sun glitter viewed from a synchronous satellite. Pierson et al. (1971) investigated the problem of radar return from a wind-roughened sea. Capillary waves are apparently important. Mc Grath and Osborne (1973) looked at the effect of wind drag on infrared images of the sea surface. A promising new technique of measuring the shape of the sea surface has been developed by Stillwell (1974). He determines the slope of the sea surface from the reflectivity of the surface.

The effects of precipitation on and near the interface are numerous and some of them have been studied. Caldwell and Elliott (1971, 1972) studied both the effect of rainfall on the surface stress and the wind in the surface layer. These effects may become significant in heavy rainfall. The rainfall has its effect on the salinity of the upper ocean layer and therefore leaves a "footprint" on the ocean (Ostapoff et al., 1973), which modifies the heat flux to the surface. Katsaros (1974) and Manton (1974) in theoretical studies have analyzed the effect of precipitation on the stability and turbulent structure below the surface. Although the temperature and salinity at the surface are only weakly affected the stability changes considerably, and this strongly damps the eddy exchange generated by shear in the drift current. Wave-induced turbulence is much less affected. Observations are needed to verify these effects.

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3. Turbulent transfer.

The turbulent transfer of properties in the mixed layers on both sides of the interface is still a central problem in air-sea interaction. Research in this area extends from the fine scale structure of turbulence (Schedvin et al., 1974) to bulk parameterization (e.g. Pond et al., 1974, Hicks et al., 1974). The bulk of the effort has been carried out in the atmospheric surface layer, only few observations in the mixed layer of the ocean have been reported.

a. Fine scale structure of turbulence.

The turbulent structure of the surface layer has been investigated mainly with the aim of obtaining the Kolmogorov constants for velocity, temperature, and humidity fluctuations (Paquin and Pond, 1971, Boston and Burling, 1972). The one-dimensional coefficient for velocity fluctuations, α , still shows considerable scatter (from 0.57 (Paquin and Pond, l.c.) and 0.55 (Wyngaard and Pao, 1971, and Mc Bean et al., 1971) to 0.48 (Schedvin et al. 1974)). On the other hand, the value of 0.52 obtained by Wyngaard and Coté (1971) over land compares well with 0.51 by Boston and Burling over water.

The uncertainty in the Kolmogorov constants for temperature, β_θ , and humidity, β_q , is still very large. Although the observations over land by Wyngaard and Coté ($\beta_\theta = 0.8$) agree rather well with those over water by Paquin and Pond ($\beta_\theta = \beta_q = 0.8$), Boston and Burling using the more basic technique of measuring the entire dissipation spectrum find the much larger value of $\beta_\theta = 1.6$. An earlier report by Gibson et al. (1970) using a similar technique mentions an even higher value ($\beta_\theta = 2.0$). So far no determinations of β_q using the dissipation range have been reported.

The Kolmogorov constant, α , may be related to the v. Kàrmàn constant, κ , by using the turbulent energy equation under neutral conditions. By measuring the wind shear and, from the inertial subrange, the dissipation (assuming $\alpha = 0.55$), Frenzen (1973a) and more recently Högström (1974) arrived at $k \approx 0.35$, which is in agreement with an independent estimate over land by Businger et al. (1971), but lower than the customary value of 0.4. However, the last word has

not been said about the constants, as is clear from the discussion between Garratt (1973) and Frenzen (1973b) and from the suggestion by Tennekes (1973) that the v. Kàrmàn constant may not be constant but varies with the roughness of the terrain.
It is clear that more experimental work is needed on the fine structure of turbulence "under conditions of site, stratification and steadiness adequately approximating those ideal circumstances specified for the theoretical analysis" (Frenzen, 1973b). The major obstacles are still the development of instrumentation to measure high frequency humidity fluctuations and the requirement of the condition of local isotropy of turbulence.

b. Fluxes and profiles.

Numerous direct and indirect observations of turbulent fluxes over water surfaces have been reported. The eddy correlation technique has been used successfully from fixed platforms (Smith, 1974, Müller-Glewe and Hinzpeter, 1974, Wieringa, 1974), buoys (Dunckel et al., 1974, Phelps and Pond, 1971), and aircraft (Donelan and Miyake, 1973, Grossman and Bean, 1973, Lenschow, 1973).
The accuracy of the fluxes is in general within ± 20%, especially when the cospectra are fully measured. Profile measurements from buoys or FLIP were discussed by Badgley et al. (1972), Paulson et al. (1973), and Dunckel et al. (1974). Donelan et al. (1975) give an extensive analysis of profile measurements from a fixed platform over Lake Ontario. The major goal of these experiments is usually to arrive at bulk transfer coefficients, which will be discussed below, but also of interest is the average structure of the fluctuations, i.e., the variances, covariances and the related spectra (Kaimal et al., 1972). Especially intriguing is the dissimilarity between temperature and humidity fluctuations over the open tropical ocean (Phelps and Pond, 1971, Holland, 1972, Donelan and Miyake, 1973 and Leavitt, 1974). The dissimilarity is pronounced in the low frequency range of the spectrum, where the humidity has considerably more variance than the temperature. At the high frequence end of the spectra the cold spikes in the temperature cause significant dissimilarity between the two types of fluctuations (Leavitt, 1974). The physical explanation of the cold spikes is still subject to speculation.
A reasonable suggestion is that the divergence of long wave radiation

along the upwind surface of moist convective plumes cools a thin layer of air which is then carried down in the general downdraft following the plume.

Few measurements have been reported of fluxes of momentum and heat in the water. Kaiser and Williams (1974) determined, the heat flux towards the surface by measuring the time rate of change of temperature of a 65m column of sea water, whereas Hoeber (1972) determined the eddy thermal conductivity (turbulent transfer coefficient for heat) in the upper 12m of the tropical Atlantic also using an indirect technique.

c. Bulk parameterization.

A major effort continues to be devoted to determining a suitable surface drag coefficient, $C_D \equiv u_*^2 / u_{10}^2$. Under neutral conditions one can relate this to the surface roughness-length, z_o. Several attempts have been made to replace or generalize Charnock's simple relation ($z_o = a\, u_*^2/g$). Manton (1971) suggested that $z_o \propto H/L$, where H is the wave height and L the wave length. Hsu (1974) combined Charnock's and Manton's relations and arrived at $z_o = (2\pi)^{-1}\, H\, (C/u_*)^{-2}$ (where C is the phase speed of the deep water waves), which seems more satisfactory physically. Hsu then compared his relation with a large number of observations and obtained reasonable agreement, although the scatter is still substantial. Stewart (1974) sketched a generalization of Charnock's relation in a similar vein, but he did not work out the details. The observational evidence supports a constant z_o as well as Charnock's formula. It would be gratifying if the scatter of the data could be reduced by properly accounting for the wave characteristics of the surface. A theoretical analysis of the drag coefficient over shallow water waves by Baines (1974) indicates that this coefficient is smaller in this case than over deep water waves. Observations by Hicks et al. (1974) seem to support this result. Also, there is the suggestion that the drag coefficient is smaller under conditions of a limited fetch than is the case of an open fetch (Smith, 1973).

Besides using measured profiles and fluxes, averaged over an hour or slightly less, the drag coefficient may be determined both on a smaller scale using structure functions or dissipation analysis, and on a larger scale from wind wet-up or from the momentum budget over a large area.

The structure function approach has been used by Franceschini and Cain (1971) yielding a C_D varying from 1.7 to 0.7, with a trend to lower values with increasing wind speed. The averaging time of 5 min. was probably too short, because the dissipation of turbulent energy is a rather intermittent phenomenon. Stegen et al. (1973) used the dissipation technique by measuring the dissipation spectra of both the wind and temperature. A correction for stability was applied and the stress results agreed well with the profile measurements of Paulson et al. (1972).

New measurements of wind set-up were recently published (Wieringa, 1974, Donelan et al., 1975) after more than a decade of neglect of this large-scale method of drag determination. The older results, mainly pertaining to extremely high wind speeds, yielded larger values of C_d than the eddy correlation technique. Wieringa also finds this ($C_d \approx 2.4 \times 10^{-3}$) for shallow water, in contrast with Donelan et al. who find $C_d \approx 1.5 \times 10^{-3}$ over the much larger and deeper Lake Ontario. Some uncertainties in deriving drag coefficients from the wind set-up may stem from a lack of proper consideration of the geometry of the water body and of the bottom stress; also the Coriolis force may play a significant role in the deep water case. A further methodological study seems desirable, particularly since the applicability of bulk drag calculations to coastal engineering problems is not proved until any discrepancies between set-up-derived stresses and eddy correlation stresses are properly explained. Attemps to derive the wind drag from numerical grid point analyses of storm surges suffer from additional statistical uncertainties.

The budget method is also still quite uncertain. Holland (1972) quotes a measurement of $C_d \approx 2.1 \times 10^{-3}$, whereas in a later study Holland and Rasmussen (1973) find $C_d \approx 1 \times 10^{-3}$, using the same BOMEX data with the application of some wind corrections. Other large-scale observations from an aircraft by Grossman and Bean (1973) yielded the relatively large value of $C_d \approx (2.6 \pm 1.4) \times 10^{-3}$. Grossman and Bean argue that the aircraft obtains a more complete sample of the stress than can be obtained at a fixed point, and consequently the full cospectrum of the u-w-components.

Experimental determinations of the bulk transfer coefficients for heat, C_T, and water vapor, C_q, are much less numerous. Recent observations by Müller-Glewe and Hinzpeter (1974) and by Dunckel et al. (1974) suggest that the overall value of 1.5×10^{-3}, recommended by Pond et al. (1974) is somewhat too large in most cases. Nevertheless, in view of the large scatter of the results it seems that $C_D = C_T = C_q = 1.5 \times 10^{-3}$ is a reasonable approximation that may be used for practical applications, provided the surface layer is slightly unstable, which is a very common condition over the oceans. The large experiments over tropical waters (BOMEX, ATEX, GATE) all indicate small Bowen ratios of 0.1 or less. Consequently the difference in temperature between air and water (usually the bucket temperature is used) is rather small and quite inaccurate for use in the bulk transfer equation. The correlation between the heat flux and $u \Delta T$ therefore leaves much to be desired. Holland (1972) plotted the heat flux against $u \Delta q$ and found a much better correlation, which suggests that the Bowen ratio tends to be constant. The determination of the bulk transfer coefficients using the budget method, although in principle possible, has not been attempted by Holland and Rasmussen (1973) or Augstein et al. (1973). The vapor budget is the most accurate and allows an estimate of the surface vapor flux with an accuracy of about 10%. The heat budget is not accurate enough to determine the very small surface heat flux (accuracy $\pm$ 100%).

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4. The structure of the boundary layers.

a. A general description.

The quasi steady state atmospheric boundary layer without condensation has been fairly well described and parameterized (see Wyngaard's review in this report). Over the ocean condensation is often an essential feature of the boundary layer, especially in the tropical regions. This adds a dimension of complexity which has not yet been fully comprehended. Nevertheless a general picture emerges from the observations by Augstein et al. (1974), Pennell and Le Mone (1974), the model by Betts (1973), and the reviews by Garstang and Betts (1974) and Businger (1974).

The slightly unstable stratification due primarily to evaporation at the surface creates a well mixed region below the cloud base which typically has a height of 500-800m over the tropical oceans. Above this region extends the cloud layer. The release of latent heat in the cloud layer drives the convection in this layer, and consequently the potential temperature there increases with height and the descending air between the clouds has a somewhat higher potential temperature than the mixed layer below, which prevents the convection in the mixed layer from penetrating into the cloud layer except where and when clouds are initiated. Thus the cloud action decouples the interaction between the lower mixed layer and the cloud layer, resulting in a small inversion between the two layers. The interaction is limited to the initiation of clouds when moist plumes penetrate the transitional inversion and reach the condensation level, and possibly also to the dissolution of cloud parts when evaporation and mixing with dry air create locally potential temperatures below the one of the mixed layer. This may develop downdrafts which reach the surface with surprisingly cool temperatures (Ulanski et al., 1974, Garstang and Betts, 1974). The top of the tropical cloud layer is characterized by the trade wind inversion, which is well developed in undisturbed conditions but almost disappears in disturbed conditons when deep convection occurs (Augstein et al., 1974).

The inversion which tops the atmospheric boundary layer and the pycnocline which limits the mixed oceanic layer have certain features in common, especially the entrainment of the turbulent mixed medium into the stable almost laminar layer beyond. This aspect of the boundary layers has received considerable attention (e.g. by Tennekes, 1973, Stull, 1973, Carson, 1973, Betts, 1973, and Deardorff, 1974, for the atmosphere, and by Phillips, 1972, for the ocean). More detailed observations of the entrainment process are needed.

The study of mesoscale phenomena within the boundary layers, such as Langmuir circulations in the ocean, and in the atmosphere helical rolls (Le Mone, 1973) and convective cells (Agee et al., 1973, Agee and Dowell, 1974) is still in its beginning stages. These phenomena, which are sometimes spectacular in satellite pictures (cloud streets, convective cells), require for their detailed description a denser network of observation stations than have been deployed so far.

Air mass modification in the atmospheric boundary layer depends on the transfer of heat and water vapor at the surface. The budget studies from BOMEX and ATEX, generally involving modest air-sea temperature differences, suggest rather gradual modification of the boundary layer air over a relatively large distance. Much more spectacular is the modification when cold arctic air flows over relative warm surfaces, e.g. the Great Lakes or the East China Sea. Such situations have been analyzed by Lenschow (1973). In two case studies he describes how the boundary layer is modified over Lake Michigan and Lake Huron and demonstrates the value of a well equipped aircraft for such experiments. Also striking is the substantial modification of the lake surface temperature, suggesting equally dramatic modification of the mixed layer below the lake surface. Preliminary results have been mentioned of the larger and more ambitious AMTEX (Lenschow and Agee, 1974) indicating a valuable set of data which may improve our understanding of the boundary layers during periods of important transformation.

b. Modelling of the boundary layers.

A variety of models have been proposed for the boundary layers varying from the relatively simple slab models, e.g. Pollard et al. (1973) (for the ocean) and Brown (1974) (for the atmosphere) to the sophisticated three-dimensional numerical model of Deardorff (1974 a,b) with a higher order subgrid closure scheme (for the atmosphere only). The slab models are well suited for the oceanic boundary layer because this layer often is well mixed and simple parameterization

seems feasible. No eddy viscosity is introduced into the problem and the depth of the layer and its mean drift are the main variables. The models which use the momentum equations (Pollard et al., 1973, Thompson, 1973) describe the drift including inertial oscillations but must parameterize the depth with a bulk Richarson number. The models which use the turbulent kinetic energy equation for the mixing of the upper layer (Denman, 1973, Kraus and Hanson, 1974) only predict the depth of the mixed layer. Both models describe the depth of the mixed layer adequately (Pollard et al., Denman and Miyake, 1973). It seems that these models are not so much contra-dictory as complementary, and that an effort to combine them should be worth while.
The K-models which use parameterized turbulent transfer coefficients (K) give more detail of the flow near the interface and have been used to describe the interaction between the two boundary layers (Pandolfo, 1971, Pandolfo and Jacobs, 1972, Yeh, 1974). Pandolfo and Jacobs are successful in simulating BOMEX observations of the atmospheric boundary layer whereas Yeh presents a model that includes both the oceanic and atmospheric boundary layers. These one-dimensional models have significant diagnostic value but are not sophisticated enough to describe the structure of turbulence nor complete enough to describe three-dimensional mesoscale phenomena.
Attempts to use higher order closure models in the air-sea interaction problem have not yet been reported. The success of Wyngaard et al. (1974a,b) in describing the convective atmospheric boundary layer over land makes it desirable that this technique be applied to the air-sea boundary layers and their interaction. The same can be said for Deardorff's (1974a,b) numerical model. Although an important objective of modelling the boundary layers is to arrive at a simple parameterization, the sophisticated models are needed to gain the necessary insight in the physical processes and may be essential in describing mesoscale phenomena.

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5. <u>Larger scale interactions</u>.

The oceanic response to large scale wind stresses and the atmospheric response to the large scale oceanic temperature fluctuations remain important problems and may lead to interesting feed-back mechanisms (White and Barnett, 1972, Bjerknes, 1972, Namias, 1971, Namias and Born, 1974). The various scales of interaction have been analyzed by the JOC-GARP Working Group on "Ocean coupling and response times" (Phillips, 1972b) and a number of recommendations made. Concerning atmospheric-oceanic models it is suggested that the following processes be simulated:

a. general circulation of the world oceans, b. seasonal variations of the upper (about 0.5km) ocean layer characteristics, and c. two-week variations of sea surface temperature; but that it is probably too early to attempt models that deal with the scales of decades and centuries.

The importance of coastal upwelling for fisheries has led to an intensive investigation of this phenomenon in its relation to the wind field. The Coastal Upwelling Experiments (CUE) have resulted in a number of papers both theoretical (O'Brien and Hurlburt, 1972, Thompson and O'Brien, 1973, Mcnider and O'Brien, 1973, Kindle and O'Brien, 1974, Hurlburt and Thompson, 1973) and experimental (Burt et al., 1972, 1973, Elliott and O'Brien, 1975, Holladay and O'Brien, 1975). The theoretical models of O'Brien et al. describe the major features of the upwelling phenomenon, such as in the area off the coast to which upwelling is confined, the time scale for the pycnocline to penetrate the surface; an alongshore baroclinic surface jet; a β-effect, which produces a poleward undercurrent on an eastern ocean circulation but not along a zonally oriented coastline.

The ultimate atmosphere-ocean interaction can only be described in a global numerical model. The numerical experiments of the scientists at NOAA's Geophysical Fluid Dynamics Laboratory are serious efforts in this direction (e.g. Wetherals and Manabe, 1972, Bryan and Cox, 1972, Manabe et al., 1975, Bryan et al., 1975).

The recent experiments by Manabe and Bryan et al. illustrate the effect of the atmosphere-ocean coupling on the climate. Although the improvement over uncoupled models is significant, there are still severe shortcomings of the model. The most disturbing aspect is the effect of the

grid size on the lowest wave numbers in the atmosphere.
Although the basic physical processes are introduced into the model there are still scale interactions (not necessarily air-sea interactions) which escape it. It is clearly desirable to continue the air-sea interaction research on all scales of interaction.

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