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AIR—SEA INTERACTION AND CLIMATE

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Air—sea interaction (ASI) is a large-scale natural phenomenon of paramount importance both in the very existence of the ocean and in many of its processes. ASI also includes the processes of heat and moisture supply to the atmosphere (mainly in the form of latent heat of evaporation and evaporated moisture) which create the atmospheric phenomena having the greatest energy concentration — hurricanes and typhoons — and long-term weather and climate anomalies.

It is usual to distinguish between small-scale (local) and large-scale (global) ASI processes. The local ASI includes in the first place the exchanges of momentum, heat and moisture through the ocean surface; important but somewhat lesser components are the exchange of gases (primarily carbon dioxide and oxygen), the transfer of sea salt from the ocean to the atmosphere, and the precipitation of aerosol from the atmosphere into the ocean. Quantitative descriptions of these processes are given by the coefficients of momentum transfer C_τ, heat exchange C_q and evaporation C_E. These were introduced for the first time by Shuleikin (1928) and are defined by the equations:

$$C_\tau = \tau/(\rho u^2); \qquad C_q = q/(C_p \rho u \delta T); \qquad C_E = E/(\rho u \delta Q) \tag{1}$$

where τ, q and E are the vertical momentum, heat and moisture fluxes at the ocean surface; τ is also called "wind stress", and E "evaporation rate" when moisture is transferred from the ocean to the atmosphere. C_p and ρ are the specific heat capacity at constant pressure and air density, respectively; u is the wind velocity (at 10 m); δT is the difference in temperature between the water surface and the air at 10 m; δQ is the difference between the saturation specific humidity of the air at the ocean surface temperature and the actual specific humidity of the air (at 10 m). Standard values of the local ASI coefficients (1) are taken to be 2×10^{-3} cm^2 s^{-1}. However they are likely to increase with the wind velocity. According to some data, the drag coefficient C_τ increases linearly, i.e. according to $C_\tau = C_{\tau_0}(1 + lu)$ with $l \approx (10 \text{ m s}^{-1})^{-1}$ and C_E grow still faster with increasing u.

If account is also taken of the fact that strong winds occur much more often than would correspond to a Gaussian probability distribution of the wind velocity vector, it becomes clear that storm regions are responsible for the basic contributions to the momentum, heat and moisture exchanges between the atmosphere and the ocean. This is corroborated by a small number of q and E flux measurements, for instance Garstang's (1965) data. The situation here appears to be similar to that for bottom relief changes in the near-shore zone of the sea, which are negligible during calm weather and become considerable only during several of the strongest storms of the year. This point of view provides a better understanding of the interesting result of Marchuk (Marchuk and Skiba, 1977) who has established, by numerical integration of the so-called conjugate equations of fluid dynamics, that the regions producing a long-term influence on weather over the territory of the U.S.S.R. include the areas of tropical hurricanes of the Caribbean and western tropical typhoons of the Pacific Ocean.

The vertical momentum, heat and moisture fluxes formed in the local ASI processes, together with the quasi-constant "buoyancy parameter" (i.e. the product of acceleration due to gravity and the air expansion coefficient characterizing buoyancy), determine the structure of the atmospheric layer near the ocean surface. The similarity theory for the surface layer of the atmosphere was developed from this assumption by Oboukhov and the author (Monin and Oboukhov, 1953) and then became a basis for the interpretation of meteorological data from the surface layer of the atmosphere in different temperature stratifications. From the point of view of similarity theory the near-water atmospheric layer differs only very slightly from the surface layer over land: as regards the possibility of a formation of drift currents (and hence in different boundary conditions for the wind velocity, instead of the zero condition which applies on solid walls) and the feedback of surface waves on the air motions above them. The similarity theory makes possible the calculation of the major parameters of the local ASI from standard meteorological measurements of wind velocity and vertical differences of temperature and humidity in the near-water atmosphere layer; special nomograms have been constructed for such calculations.

The similarity theory for the near-water atmospheric layer can be expanded to cover the entire atmospheric boundary layer (ABL). For this purpose it is necessary to add to the set of determining parameters of the similarity theory the ABL thickness, or in case of a stationary and horizontally homogeneous ABL (called the "Ekman" boundary layer — EBL), the Coriolis parameter f which determines the thickness of the Ekman layer through $h \sim (\tau/\rho)^{1/2} f^{-1}$. Such a similarity theory for the EBL was suggested for the first time by the author (Monin, 1950) and was later developed in a number of his co-authored papers (Kazansky and Monin, 1960, 1961; Zilitinkevich and Monin, 1974, 1976).

The local ASI effects determine both the ABL structure and the structure of the upper mixed layer (UML) of the ocean. The determining parameters of the similarity theory for the UML are the vertical momentum, heat and salt fluxes, together with the buoyancy parameter and the Coriolis parameter; by contrast in the dynamical theory one of the most important additional quantities is the vertical turbulent energy flux. Similarity of the temperature profiles in the upper thermocline was established by Kitaigorodsky (1960). The semi-empirical dynamical theories of the UML deal both with its structure and with the synoptic and seasonal variations of its thickness, as well as density jump values in the underlying pycnocline layer.

One of the most significant aspects of the UML theory is the description of wind-driven waves — their generation, the distribution of momentum and kinetic-energy fluxes from the atmosphere among the wind-driven waves, internal waves, drift currents, surface waves breaking, and the generation of dynamical turbulence causing (together with thermal turbulence, i.e. convection) UML mixing and thus determining the UML thickness and its synoptic and seasonal variations. All these problems are far from completely solved. In particular, accurate experiments have shown that the elegant theories of wind wave generation of Phillips (1957) and Miles (1957) are still inadequate since they underestimate (by an order of magnitude) rates of growth of wind-driven waves; Zaslavsky's suggestions to make the theory more precise, by taking into account the random character of turbulent wind velocity profiles in the near-water atmospheric layer, will still not help here. Among new approaches Chalikov's (1976) numerical experiments on wind-wave generation should be mentioned.

When integrated over area and time the local ASI effects lead to a number of global processes in the ocean and the atmosphere. The most important of these, from a practical point of view, are long-term weather anomalies. These include, first of all, the anomalies generated by processes of the thermobaric seiche type in the atmosphere that were the subject of a number of Shuleikin's (1939, 1942) papers. Next the present author has argued (Monin, 1963) that the water surface temperature in the World Ocean is the most important initial field for long-term weather forecasting; even small anomalies of the water surface temperature field may correspond to rather considerable anomalies of the heat content of the upper ocean layers, as was shown by Kort (1970) from multi-year measurements on a section through the Kuroshio Current.

A vivid example of a large-scale ASI process is given by the El Niño phenomenon studied by Bjerknes (1966). It consists of an attenuation of the easterlies in the eastern part of the equatorial Pacific (on the south-eastern periphery of the Hawaii anticyclone); this leads to weakening of the equatorial upwelling, heating of the upper layer of the ocean and then of the atmosphere above it, amplification of the trade-wind circulation and subsequently of the easterlies in temperate latitudes, and deepening of the Aleutian cyclone. This phenomenon can be described as a natural calamity because attenuations of upwelling in that region lead to mass fish-kills and sharp decreases of anchovy catches. In the Atlantic Ocean during such periods the easterlies in temperate latitudes weaken, the Icelandic pressure minimum becomes shallow, the easterly winds north of Iceland slacken, and the Arctic Ocean appears to be affected by an anticyclone in northern Alaska. Thus this process embraces the entire Northern Hemisphere.

The global ASI is one of the most important factors in climate formation. Climate in this context is defined as the statistical ensemble of the states through which the atmosphere—ocean—land system passes during time periods of the order of several decades. Not so long ago the definition of climate referred only to the state of the surface layer of the atmosphere; it is now clear that the entire atmosphere, ocean and their interaction should be taken into account, as well as the active layer of the underlying land surface. Note that representations of climate types are not strictly zonal but reflect differences between the continents and the oceans. The significance criterion of these differences is the rotational Mach number $Ma = \omega l/c$, where ω is the angular velocity of the planet's rotation, l the planet's radius, and c the velocity of sound in the planet's atmosphere; at large Ma zonality effects predominate, and at small Ma the difference between the daytime and night-time sides of the planet. For the Earth $Ma \approx 1.4$ and the non-zonal effects of the differences between the continents and the oceans are comparable with the latitudinal effects.

In regard to effects produced on the atmosphere the oceans differ from the continents; firstly in their thermal properties, their much greater turbulent heat conductivity and heat capacity produces a "heat inertia" smoothing short-period diurnal and seasonal temperature fluctuations. For this reason the oceans are cooled in winter and heated in summer to a lesser extent than are the continents and hence appear to be warmer in winter and colder in summer than the continents. The mean annual temperature contrasts between the equator and the poles produce atmospheric and oceanic circulations which serve to smooth out these contrasts; in addition in the lower atmosphere there occur seasonal temperature contrasts (changing sign from winter to summer) between the continents and the oceans, and general seasonal circulations which reduce those contrasts and are called monsoon circulations. These seasonal oscillations represent the most vivid manifestation of the difference between the continents, where the amplitudes of the seasonal temperature changes in the surface layer of the atmosphere are very great, and the oceans, where the seasonal amplitudes are small. As a result the map of the seasonal temperature amplitudes (see Fig. 1) clearly shows the distribution of land and sea, without any indication of shorelines.

Apart from their differences in heat conductivity and heat capacity, oceans and continents differ, on the average, in their ability to reflect short-wave solar radiation. Satellite albedo maps show, alongside the general growth of albedo from the equator towards the poles, an increase of albedo from the oceans towards the continents in the same latitudes. For this reason the continents, on the annual average, should be somewhat colder than the oceans. The planetary albedo map also exhibits a certain year-to-year variability; this

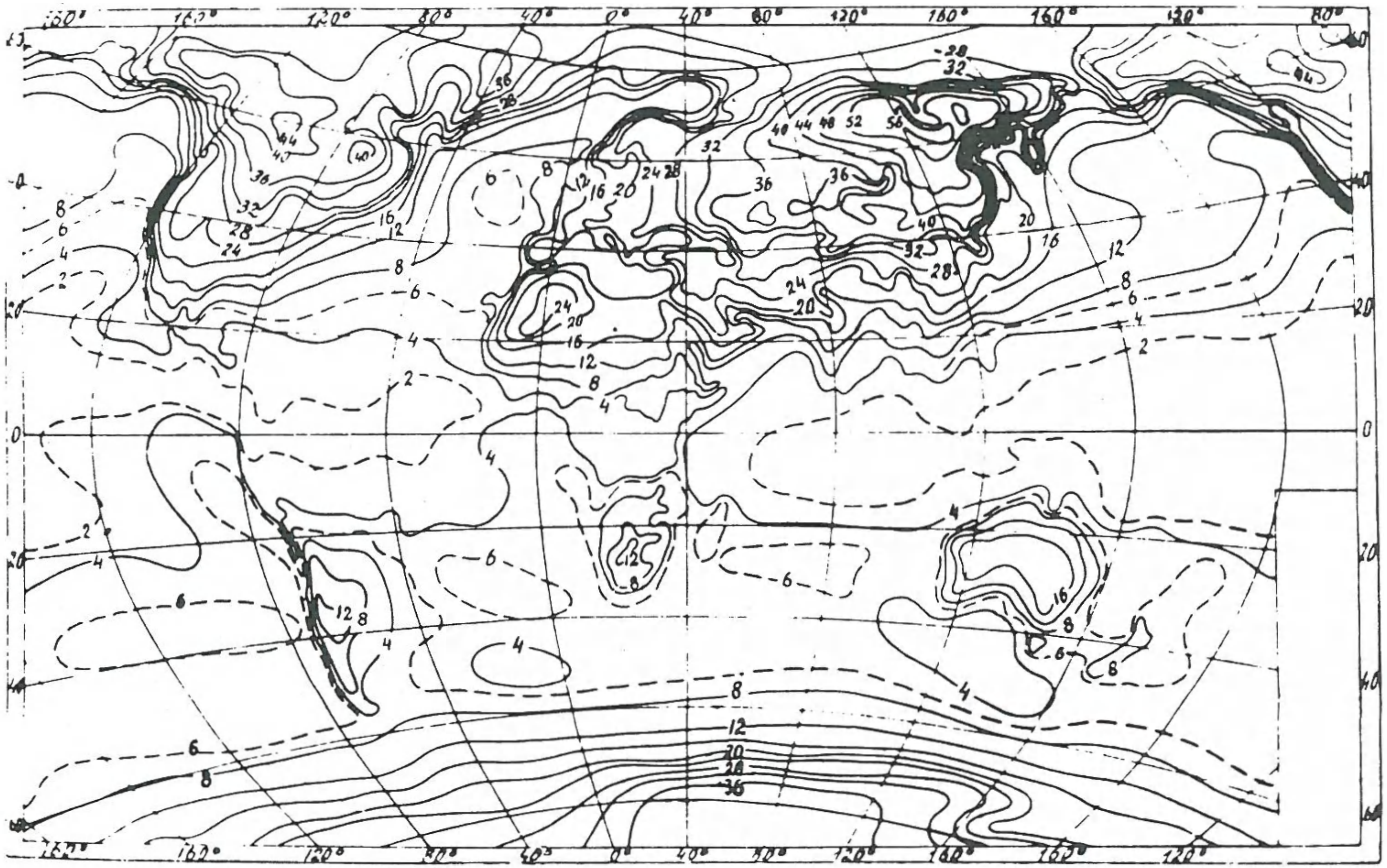

Fig. 1. Map of seasonal temperature-fluctuation amplitudes.

is one of the sources of long-term changes in weather and possibly in climate.

The seasonal temperature differences between the oceans and the continents are demonstrated by maps of mean monthly sea-level temperatures. In the winter maps, the coldest regions are found in Antarctica, Yakutiya, North Canada and Greenland, whereas in the summer maps, the highest temperature appears in the subtropical deserts of Africa, Southern Asia and Mexico. Monsoon effects are clearly demonstrated by mean monthly maps of the sea-level atmospheric pressure. These maps show quasi-permanent subtropical regions of high pressure amplifying from winter to summer (the Azores and Honolulu anticyclones in the Northern Hemisphere, the Saint Helena, Mauritius and South Pacific anticyclones in the Southern Hemisphere) and low-pressure regions closer to the poles amplifying from summer to winter (the Icelandic and Aleutian cyclones in the Northern Hemisphere and the circum-polar trough in the Southern Hemisphere). On the continents, winter regions of high pressure (in Siberia, Canada, South Africa and Australia) are replaced by low-pressure regions in summer. The global map of annual total precipitation gives a good presentation of the humid intertropical convergence zone and the arid zones of the subtropical deserts. An increasing precipitation trend is evident from subtropical to temperate latitudes, from the continents to the oceans, as well as in the near-shore areas of the monsoon regions and on windward (mainly western) mountain slopes.

Stepanov (1974) has made considerable progress in the description of the ocean climate. Here are some zonal characteristics of the ocean climate he has obtained. The heat budget of the ocean is positive (the ocean is heated) in the tropical zone between 30° N and 15° S and negative (the ocean is cooled) beyond this zone; the greatest positive budget, up to 80—100 kcal cm^{-2} y^{-1}, is recorded in the equatorial Pacific, the greatest negative budget, up to 75—100 kcal cm^{-2} y^{-1}, in the Gulf Stream and Kuroshio zones. The mean annual zonal temperature of the ocean surface in the tropical zone exceed 25° C, reaching a maximum of 27.4° C slightly north of the equator; in temperate latitudes they decrease rapidly towards the poles, dropping below zero in the zones 60—65° S and 70—75° N. The mean temperature of the whole water column in the World Ocean (without the Arctic Basin) is 5.7° C.

The water budget of the ocean is positive (precipitation exceeds evaporation) in the equatorial zone between 10° N and 5° S and in temperate latitudes; it is negative (evaporation exceeds precipitation) in the tropical and subtropical zones. The greatest positive budget, up to 150—200 g cm^{-2} y^{-1}, is recorded in the western part of the equatorial Pacific, the greatest negative budget, up to 150 g cm^{-2} y^{-1}, in the subtropical regions, especially of the Atlantic Ocean. The surface water salinity reaches its maximum in the subtropical ocean (35.75‰ in the zones from 30—25° N and 20° S); it has a secondary minimum in the equatorial zone (34.43‰ in the zone 10—5° N) and decreases towards the poles in temperate latitudes falling below 35‰ in about ±40° latitude to 35.5‰ and lower in Antarctica, and to 31—30‰ in the Arctic Ocean. The average salinity of the whole water column of the World Ocean is 34.71‰.

The sea water density, conveniently stated in units of $\sigma_t = 1000\ (\rho - 1)$, has its minimum at the ocean surface in the equatorial zone (22.18 in the zone 10—5°N); it increases towards the poles, reaching 27.30 in Antarctica and 26.19 in 55—60°N latitudes. Further into the Arctic Ocean σ_t decreases again to 24.55.

The quasi-stationary currents at the ocean surface are evidently of wind origin; their mean velocities are about 10 cm s^{-1} and they are responsible for the piling up of water which, together with thermohaline expansion and compression of the water, results in deviations of the ocean surface level from the geoid equilibrium level, of the order of decimeters. The greatest upward deviations are observed along the western peripheries of the oceans, particularly in the subtropical zone, the greatest downward deviations in the near-polar regions. A contribution to the motion of the ocean water, comparable to those quasi-stationary currents, is made by synoptic eddies with horizontal scales of the order of 10^2 km and time scales of the order of months.

The global ASI contributes both to climate formation and to the generation of a number of processes producing climate oscillations. This is particularly well illustrated by climate oscillations with periods of tens or hundreds of years, observed during historic times. For instance, the climatic warming of the first half of the 20th century, according to Mitchell's (1963) data, occurred in the oceans; on the contrary, a slight cooling took place during that time on the continents. The "Little Ice Age" of the 17th—19th centuries, according to Bjerknes's (1965) assumptions, can be explained by a positive feedback between negative anomalous temperatures in the Atlantic waters in the vicinity of Iceland and positive anomalies in the Sargasso Sea, on the one hand, and the attenuation of the winter atmospheric circulation in temperate latitudes over the Atlantic due to ASI attenuation, on the other hand. The development of this process stops at some level because of the appearance of a negative feedback with the meridional heat transport by the ocean currents.

An alternation of glacial ages with immense ice sheets and interglacial periods when these ice sheets melted away almost completely, occurred during the Pleistocene, with periods of the order of 20—100 thousand years; similar alternations apparently took place in the Permo-Carboniferous and in earlier glacial epochs, such as the Vendian, Upper Riphean and Proterozoic. According to Milankovich (1930) these alternations can be explained as reasonably amplified forced oscillations caused by astronomic oscillations of the Earth's orbital elements and equatorial inclination to the ecliptic. Direct evidence in favor of such an explanation (astronomic periods in the spectra of climatic indicators) are given in the paper by Hays et al. (1976).

During the warm geological eras no resonant amplification of astronomically forced climate oscillations occurred and these oscillations were negligible because in their minima the climatic background remained so warm that no continental ice sheets were formed. In this scheme, the climatic background variations (coolings and warmings) of geological eras with time scales of the order of 10^8 years are still awaiting explanation. Such an explanation may be found in qualitative variations of the global ASI caused by changes in the configuration of oceans and continents (and the poles) due to continental drift.

Quantitative theories of the global ASI can be constructed on the basis of global physical—mathematical models of the atmosphere and the ocean in their interaction (or, more completely, of the whole atmosphere—ocean—land system). Of different possible classification of such models we shall point out here their division into few-parameter models (non-hydrodynamic models with lumped parameters) and multiparameter or comprehensive (hydrodynamic) models. The first group cannot lay a claim to quantitative details; however, even before super-power computers are created (which will make it possible to perform statistical-hydrodynamic numerical experiments with the multiparameter models) these models can yield a number of constructive results, even though a really provable theory will be provided only by the multiparameter models.

The minimum number of parameters of the models of the first type should include the mean global temperature of the surface layer of the atmosphere and the typical temperature difference between the equator and the poles. Such models can be constructed in the pattern of the similarity theory for the circulation of the planetary atmospheres of Golitsyn (1973) by adding the parameters that characterize the role of the oceans and the continents. The model constructed by Zilitinkevich and Monin (1976) may serve as an example. A model with lumped parameters, due to Sergin and Sergin (1969) should also be noted; this describes the intermittence of glacial periods.

By now several tens of multiparameter models of the atmosphere have been constructed in different countries; however, only two genuine multiparameter models of the atmosphere—ocean—land system are in existence. The first was constructed by Manabe and Bryan (1969) and has already yielded a number of promising results. Some of its shortcomings include preassigned cloudiness (from climatic data) and artificial matching of large time intervals in the ocean with small intervals in the atmosphere. These shortcomings are eliminated in a model which was completed in 1976 in the Leningrad Department of the P.P. Shirshov Institute of Oceanology, U.S.S.R. Academy of Sciences, by Chalikov with the participation of Turikov, Zilitinkevich and the author (Zilitinkevich et al., 1976). Further improvements of the models of this type should lay a basis both for long-term weather forecasting and for the theory of climate.

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