New temperature–humidity criterion for estimation of the possibility of tropical cyclone generation

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Received March 29, 2004

Typical conditions for tropical cyclones generation have been analyzed. The complex criterion Ω is proposed to characterize a probability of cyclone generation and its power. The criterion can be calculated based on three parameters: the ocean surface temperature, the above-surface relative humidity of air, and the geographic latitude. The estimates of probabilities of the tropical cyclones generation based on the criterion suggested and on the frequency of cyclone generation during the 16-year period have been compared. Besides, the average values of the total energy and kinetic energy of a cyclone were calculated using the method developed and compared to available estimates. A good agreement of these estimates proves the usefulness of the criterion suggested and indicates a need of the further investigation of physical phenomena, which enabled us to formulate the criterion.

Any tropical cyclone (TC) is conducted by a process of fast release of a huge energy. The energy of tropical cyclones is on the order of 10^{20} J, the wind sometimes achieves 80 m/s, causing a severe damage of buildings and many casualties, and tropical showers cause no less destructive floods.[1] In Russia, the territory subject to cyclones is Sakhalin and neighboring territories. The lifetime of a tropical cyclone is from several days to 1–2 weeks. The release of huge energy for a short time suggests that in the pre-cyclone period the energy is gradually accumulated up to a certain level. Then, as a result of some "trigger" process, the energy is released in the form of a tropical cyclone. Monitoring of the Global Ocean for revealing the cyclone-risky regions is an important problem of the satellite oceanology.

To evaluate the probability of tropical cyclone generation in a given water area and to forecast the precipitation amount in the zone of cyclone propagation, we have analyzed typical conditions for cyclone generation. This analysis has yielded some combined criterion Ω characterizing the probability of cyclone generation and its power.

Concise characterization of the process of tropical cyclone generation

A tropical cyclone is the process of fast rise of a warm and humid air, accumulated in the boundary zone near water surface, to the tropopause (12–14 km), during which a huge amount of moisture is lifted to high altitudes, moves, and then falls as powerful precipitation. In this case, the upward airflow is swirled around the axis (cyclone). The observed tangential wind velocity is very high at relatively low radial and vertical wind components. For the tropical cyclone to occur, a stock of potential energy is needed, which then is transformed into the kinetic energy of extreme wind velocities. The potential energy is gradually accumulated in the atmosphere above the ocean, before the beginning of cyclone generation, and its source is, certainly, the solar energy.

The solar energy flux heats the upper layer of the ocean water. The water surface heats air, an intense evaporation begins (so-called hidden heat flux) and, as a result, the boundary layer of several kilometers thick is filled with warm and humid air (according to measurements, the relative humidity at the height of 10 m above the surface achieves 85% and higher). If the surface air layer begins to rise, then the moisture plays the role of fuel, because a considerable amount of heat is released as it condenses during the ascent. Air expanding due to heating, as a piston of a heat engine, raises the "heavy" atmosphere, and an unstable equilibrium is established. Thus, the first condition for TC generation is a presence of prerequisites, allowing the warm and humid air to ascend to the tropopause, that is, prerequisites for appearance of the through-troposphere convection.

Another condition is a presence of "suction" of the warm and humid surface air from surrounding areas to the place of appearance of the through-troposphere convection. The suction appears in a region, where the Coriolis force results in swirling of airflows directed to the center and, consequently, in the pressure reduction. This, in turn, causes air inflow from surrounding regions.

Our criterion for estimation of the TC generation probability was formulated from these two conditions. Consider factors leading to their realization.

Effect of air temperature and humidity in the above-surface ocean layer on the possibility of through-troposphere convection

Let us analyze the conditions, under which the heated surface air can rise to the tropopause.

Consider the physical processes occurring upon motion of some bulk of air from a lower layer to some

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upper one. Let a "trial" air volume ascend from the ocean surface fast enough, so that the heat exchange with the surroundings can be neglected. The temperature of the volume obeys the adiabatic law:

$$T/T_S = (P/P_S)^{\gamma-1}, \quad (1)$$

where T_S is the temperature of the ocean surface; P_S is the atmospheric pressure above the surface; $\gamma = C_P/C_V$, C_P is the molar specific heat of air at a constant pressure, C_V is the molar specific heat at a constant volume, $\gamma \cong 1.284$. This is so-called dry adiabat.

The process of ascent of the trial air volume obeys this law up to the height h_0 (pressure p_0, temperature T), at which the partial pressure of water vapor in this volume becomes equal to the saturation vapor pressure $p_{w.v} = p_s(T_0)$. Find this height. Let k_S be the relative humidity of air near the ocean surface. Then the water vapor partial pressure near the ocean surface is

$$p_{Sw.v} = k_S p_s(T_S). \quad (2)$$

Here $p_s(T)$ is the saturation vapor pressure at the temperature T:

$$p_s(T) = 1.9 \cdot 10^{-9} \cdot 10^{-2320/T}. \quad (3)$$

The ratio of the water vapor partial pressure to the total pressure keeps unchanged with height until the condensation or evaporation occurs; consequently, at h_0

$$k_S p_s(T_S)/P_S = p_s(T_0)/P_0. \quad (4)$$

Since the values of P_S, P_0 and T_S, T_0 satisfy Eq. (1), then

$$k_S p_s(T_S)/p_s(T_0) = (T_S/T_0)^{1/(\gamma-1)}. \quad (5)$$

Upon substitution of Eq. (3) into Eq. (5) and calculation of logarithm of the both sides, we obtain the equation for determination of T_0:

$$2320 \ln 10(1/T_0 - 1/T_S) + $$
$$+ 1/(\gamma - 1) \ln(T_0/T_S) = -\ln k_S. \quad (6)$$

An approximate solution of this equation can be easily obtained, taking into account the fact that $(T_S - T_0) \ll T_S$:

$$T_S - T_0 \approx -\ln k_S\, T_S^2/[2320 \ln 10 - T_S/(\gamma - 1)]. \quad (7)$$

The equation for the pressure at h_0 is derived with allowance for Eq. (4):

$$P_0 \approx P_S \exp[-2320 \ln 10\, (T_S - T_0)/T_S^2]/k_S. \quad (8)$$

Consider the process of further ascent of the trial air volume from the height h_0. The air temperature decreases with height, and, consequently, the saturation vapor pressure decreases as well. Determine the amount of moisture condensed upon ascent of the trial air volume from h_0 to some h_1. The pressure and temperature of the trial air volume at this height are P_1 and T_1. The fraction of water vapor in one mole of air M is proportional to the ratio of the water vapor partial pressure to the total pressure, therefore the amount of the condensed vapor can be determined from the difference of these ratios at different heights:

$$\Delta m = M(p_s(T_0)/P_0 - p_s(T_1)/P_1). \quad (9)$$

Upon condensation, some amount of heat is released in one mole of air

$$\Delta W = \Delta m \xi, \quad (10)$$

where ξ is the molar heat of evaporation of water ($\xi \cong 10746$ cal/mol). This heat is consumed for heating of the trial air volume.

Figure 1 depicts the PT-diagram of the process. From the ocean surface to some height h_0 with pressure P_0, the temperature variation follows so-called dry adiabat, obeying the law expressed by Eq. (1). If water vapor did not condense at high altitudes, at h_1 the trial air volume would have the temperature:

$$T_1^* = T_0(P_1/P_0)^{\gamma-1}. \quad (11)$$

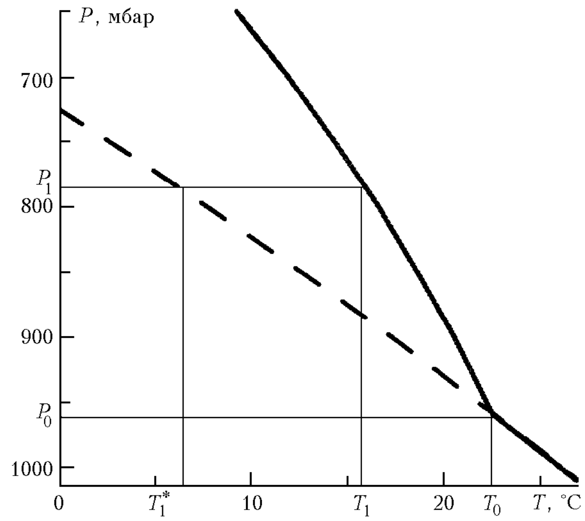

Fig. 1. Temperature conditions of trial air volume rising from the ocean surface; air temperature near surface is 27°C, relative humidity is 85%; dry adiabat (dashed line), adiabatic expansion of humid air (solid curve); P_0 is the height (in units of pressure), at which water vapor begins to condense.

However, because of heating due to condensation, its temperature turns out to be higher, and

$$C_P(T_1 - T_1^*) = \Delta W. \quad (12)$$

Upon substitution of Eqs. (9), (10), and (11) into Eq. (12), we obtain the equation for determination of the temperature of the trial air volume at h_1 (pressure P_1):

$$C_P(T_1 - T_0(P_1/P_0)^{\gamma-1}) =$$
$$= \xi(p_s(T_0)/P_0 - p_s(T_1)/P_1). \quad (13)$$

The calculated results at the ocean surface temperature of 27°C and relative humidity of 85% are shown in Fig. 1.

Let us determine whether the through-troposphere convection is possible under the values of air temperature and humidity typical near the ocean surface. To do this, compare the obtained curve with the typical vertical profile of the air temperature in the stable tropical atmosphere. We use the experimental data on the tropical atmosphere obtained at the *Akademik Kurchatov* and *Professor Shtokman* research vessels of the Institute of Oceanology RAS during the many-month PIGAP (Program for Investigation of Global Atmospheric Processes) mission in the equatorial zone of the Atlantic (3°N – 3°S).[2] The study of air temperature vertical profiles obtained from the data of more than 200 radiosondes has shown that for the ocean surface temperatures, which are characteristic of tropical latitudes (23–29°C), the *PT*-curve of the atmosphere is almost independent of the ocean surface temperature starting from the height of 2–3 km and higher. Figure 2 depicts such a typical *PT*-curve (*1*).

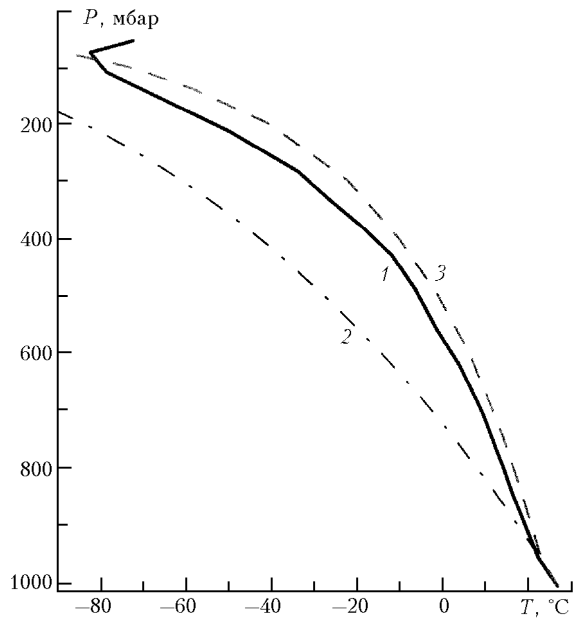

Fig. 2. *PT*-curves for the atmosphere above the ocean surface: measured in the equatorial zone (*1*), calculated for the ascending trial volume of dry air (*2*), calculated for the ascending trial volume of air with the relative humidity of 85% at the level of 10 m above the surface (*3*). Air temperature near the ocean surface is 27°C.

Let us estimate the conditions, under which the typical tropical atmosphere becomes instable. This means that the trial air volume, starting to rise, reaches the tropopause. Let the trial air volume, having the temperature of 27°C near the ocean surface, lift conditionally to different heights. If the initial relative humidity of this volume were zero, the process would follow curve *2* in Fig. 2 (the dry adiabat). It can be seen that in this case up to the height h_0 (P_0) the temperature of the trial air volume is equal to the temperature of the surrounding air, that is, free mixing of air is taking place. Above h_0, at any height, to which the trial air volume can rise, its temperature would be lower than the temperature of the surrounding air, and the density, consequently, higher than that of the surrounding air; therefore, the trial volume would come back to the initial layer, that is, into the layer of free mixing ($0 - h_0$). This effect can be called an adiabatic lock, since it prevents the vertical convection.

A quite different pattern is observed, if we take into account the air humidity. With regard for the heat released upon condensation of water vapor during the ascent, according to Eqs. (9)–(13), for the relative air humidity of 85% at the level of 10 m above the surface and the air temperature of 27°C near the ocean surface, as was observed in the April expedition, we have calculated the parameters of the trial volume (P, T), that is, the so-called wet adiabat was obtained (curve *3*, Fig. 2). It can be seen that above the height h_0 (0.5 km or 960 mbar) the temperature of the trial air volume is higher than that of the surrounding air, therefore, the trial air volume ascends freely up to the tropopause. Consequently, at the ocean surface temperature of 27°C and the relative air humidity of 85% near the surface, the through-troposphere convection is observed in the typical tropical atmosphere, which is a prerequisite for TC.

As is known, TCs are almost always generated at the ocean surface temperature higher than 26°C. Consider what happens at temperatures below this threshold. Let the surface temperature T_S be 24.8°C and the relative air humidity near the surface k_S be 85%. The calculations under these conditions have shown that the temperature of the trial air volume exceeds the temperature of the surrounding air only up to a height of 2.5 km. Consequently, the warm and humid lower-layer air cannot rise above this height. The available potential energy in this case is, apparently, insufficient for formation of the though-troposphere convection and TC.

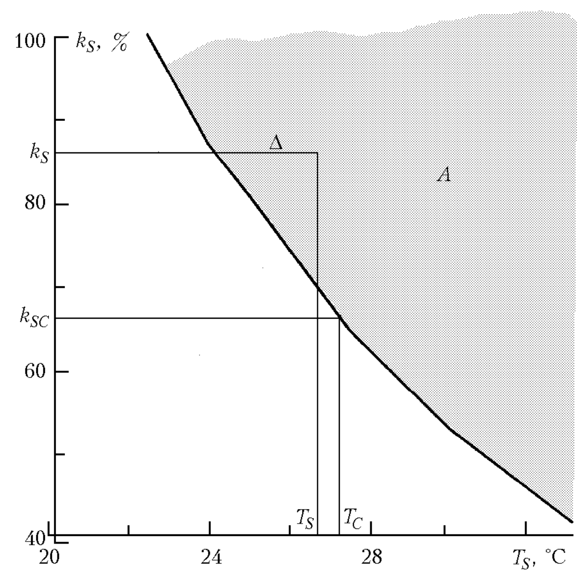

Fig. 3. Relative air humidity at the level of 10 m above the surface and the ocean surface temperature, determining the possibility of tropical cyclone generation: TC is probable in the shaded range A.

Thus, the existence of some temperature threshold for TC formation finds an explanation. However, we

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know that water vapor plays a very important role in the appearance of the atmospheric instability. Calculations have been performed for different values of the relative humidity of air above the ocean surface at various water temperatures. We have found the range of temperature and humidity values, starting from which the through-troposphere convection can occur (Fig. 3). In Fig. 3, this range is located to the right from the curve (A). The range of the T_S and k_S parameters, at which the through-troposphere convection and, consequently, TC generation are unlikely, lies to the left from the curve.

Formation of the temperature–humidity criterion for estimation of the possibility of TC generation

Using the dependences obtained, we formulate the criterion Ω so that to take into account not only the temperature, but also the humidity of the surface atmosphere.

Assume that the greater is the distance from the point (T_S, k_S), characterizing the parameters of some state of the atmosphere above the ocean, to the threshold curve (see Fig. 3), the higher is the probability of the through-troposphere convection and, correspondingly, TC generation. Linearize the threshold curve at some its central point ($T_C = 27.5$°C, $k_{SC} = 67\%$) inside the temperature range, in which storms are observed most often. It is remarkable that this range is very small: from 25.5 to 29.5°C (Ref. 4). Denote it as $\Delta T_{max} = 4$°C; the related increment of k_S along the curve (see Fig. 3) is $\Delta k_{Smax} = 24\%$. Introduce some parameter Δ, characterizing the distance from the point (T_S, k_S) to the threshold curve in degrees centigrade. Dividing the parameter Δ by ΔT_{max}, we obtain a more convenient dimensionless parameter

$$\delta = (T_S - T_C)/\Delta T_{max} + (k_S - k_{SC})/\Delta k_{Smax}. \quad (14)$$

The conditions for generation of the through-troposphere convection are observed most often in the near-equatorial zone. However, many-year observations show that no TCs are generated at least at latitudes from 5°N to 5°S. In this latitudinal belt, the process of the through-troposphere convection proceeds in the following way. Giant columns of cumulus clouds grow, separated by several tens of kilometers from each other. At a fine weather, a seafarer onboard a ship can see 10–12 such cloud columns simultaneously. These columns are similar in shape and can achieve very high heights. Obviously, they are results of the through-troposphere convection, but TC is not generated in this case.

The vertical convection in an air column is accompanied by "dragging" of humid air from the surrounding lower troposphere. The same process proceeds near other columns as well. As a result, there arises a gentle differently directed wind, generating ripples on the water surface. The ripples are observed by the satellite-borne radio telescopes as atmospheric cells.

In the near-equatorial zone, though the air coming to the region of the through-troposphere convection has a low angular momentum, it is insufficient to generate a cyclone. With the distance from the equator, the angular momentum significantly increases due to the effect of the Coriolis force, together with the area, from which the humid air comes to the center of every column. Larger columns take up smaller ones, and the TC generation occurs. The second main condition for TC generation is a nonzero Coriolis acceleration. Therefore, some value proportional to $\sin\varphi$, where φ is the latitude of the location, is used by us as the second necessary parameter.

Now we can formulate the general criterion for the TC generation probability in a given water area:

$$\Omega = \delta \sin\varphi. \quad (15)$$

Thus, we have formulated some criterion depending on the ocean surface temperature, near-surface relative humidity of air, and the latitude. Let us show that it is proportional to the probability of TC generation.

Use of TC generation statistics for evaluation of the temperature–humidity criterion

We use the available data on TC generation for 16 years.[3] Let us compare the distribution of Ω over the latitudinal zones in oceans with the probability of TC generation in the corresponding water areas. The criterion Ω was calculated from the averaged climatic parameters (temperature and humidity of the surface atmospheric layer). The frequency of storms in the considered water areas served an estimate of the true probability of TC generation.

We have chosen two regions of the global ocean: the Indian Ocean to the south from the equator and the Atlantic Ocean to the northwest from the equator. Figure 4*a* depicts distributions of the number of TCs for 16 years in the 5°-wide latitudinal belts: 0–5, 5–10, 10–15°, and so on in the Indian Ocean. Figure 4*b* shows the distributions of the TC number for 16 years in the Northern Atlantic to the west from the 35°W meridian in the 10°-wide latitudinal belts: 0–10, 10–20, 20–30°, an so on. These distributions are normalized to the maximal TC number in a given latitudinal region. It is clear that the height of every column characterizes the probability of TC generation in the considered region.

As expected, the probability of TC generation near the equator is close to zero; it increases to the tropics and then decreases with the further increase of latitude. It can be also seen that the regions analyzed give different patterns of the probability distribution: in the Indian Ocean the peak falls on the latitudes from 10 to 15°S, while in the considered area of the Atlantic Ocean it is shifted from the equator by more than 10° and falls on the latitudes of 20–30°N.

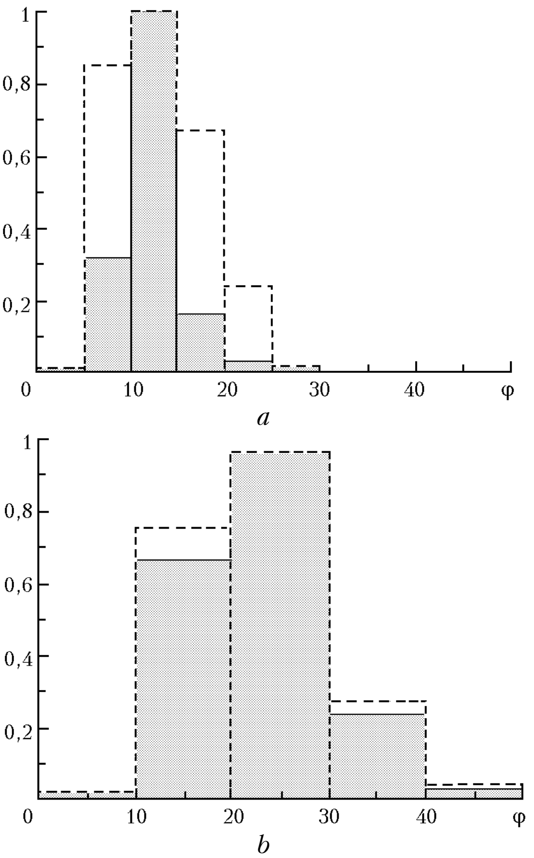

Fig. 4. Distribution of the frequency of TC occurrence (shaded area) and dependence of Ω averaged for a year (dashed curves) on the latitude φ to the south from the equator in the Indian Ocean (*a*) and to the north from the equator in the Atlantic Ocean (*b*).

Then we have estimated Ω averaged for a year in every latitudinal belt. The results, shown by the corresponding plots, were normalized to the maximal value of Ω for each ocean. The mean temperature near the surface was taken from the *Atlas of Oceans*.[4] The relative humidity was obtained by averaging the experimental data on humidity at a height of 10 m above the water surface,[2] and it turned out to be 80%.

It turned out that Ω adequately reflects the latitudinal regularities of TC occurrence: in the Indian Ocean the peak of the storm activity, according to Ω, lies in the belt from 10 to 15°S and, consequently, coincides with the real latitude of generation of the maximal number of TCs. In the Northern Atlantic, due to the Gulf Stream, the warm waters are observed in the higher latitudes, and, therefore, Ω achieves its maximal values in the 20–30°N belt, which is characterized by the maximal frequency of TC generation for 16 years. Thus, in the both cases, the agreement can be considered as satisfactory, especially, between the latitude dependent positions of maxima.

Compare now the energy released in the TC and the energy accumulated in the zone of TC development. Within our scheme of TC development, the "working medium" in this giant heat engine is the water vapor located in the lower troposphere. As estimates show, only the air located in the layer no higher than 500–700 m above the ocean surface can be involved into the through-troposphere convection. Assume that, upon the ascent to a height of 10–12 km, all water vapor from this layer condenses, giving its heat to the development of TC. Then the stored thermal energy can be estimated as:

$$E \cong LkuS \approx 10^{20}\ \text{J}, \tag{16}$$

where L is the specific heat of evaporation ($L = 2.4$ kJ/g); u is the mean amount of water vapor in the 1-cm^2 atmospheric column, which is characteristic of the TC zone ($u = 5$ g/cm^2); S is the area of the ocean surface, over which TC develops (mean cyclone path length of 1200 km, radius of 300 km); k is the coefficient accounting for the fraction of water vapor involved in the process of cyclogenesis ($k \approx 0.8$).

Assuming the efficiency of the TC "heat engine," i.e., the efficiency of conversion of the stored energy into the kinetic energy of the air mass motion, equal to several percent, we obtain the estimate of the kinetic energy released upon passage of a cyclone being of the order of 10^{18} J. The estimate obtained well agrees with the available estimates of the total and kinetic energy of TC calculated by other methods.[5]

The agreement between the estimate of the TC generation probability by the proposed criterion and the long-period frequency of TC occurrence, as well as between the estimates of the total and released energy with our estimates indicates a possibility of using the proposed temperature–humidity criterion Ω, as well as the need in the further investigation of physical principles that have allowed us to formulate it.

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