NOTES AND CORRESPONDENCE

A Note on the Sea Surface Temperature Sensitivity of a Numerical Model of Tropical Storm Genesis

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ABSTRACT

In a three-dimensional numerical model of a tropical disturbance, a spectrum of development stages, from a weakening wave to a mature tropical storm, was obtained with a 5 K range (298 to 303 K) sea surface temperature (SST). However, the apparently large SST sensitivity of the model was found to be modulated by other factors including the large-scale environmental temperature and humidity. Through the use of this model, problems concerning a critical value of SST necessary for storm development were discussed.

1. Background

The impact of sea surface temperature (SST) variation on tropical storm development has been extensively investigated since the pioneering work of Palmén (1948). Climatologically, tropical cyclone formation is associated with SST's of 26–27°C or higher (e.g., Wendland, 1977). Some observations have indicated this strong relationship exists on the shorter time scales of individual storm cases (e.g., Miller, 1958). But this has been contradicted by Ramage (1974). In the controlled environment of numerical simulation, the high sensitivity of development of an axisymmetric vortex to SST is well documented by the investigations of Ooyama (1969), Sundqvist (1970), Rosenthal (1971) and Anthes and Chang (1978). The dependence of mature storm intensity on the interaction with the ocean has been investigated by Chang and Anthes (1979), who concluded that, except for relatively slow moving storms, cooling of SST due to upwelling and mixing processes has little impact on storm intensification. This study will investigate the impact of SST variation on tropical storm formation from a rather realistic shallow easterly wave. This is in contrast to previously performed model investigations on the intensification of hurricanes from axisymmetric, vertically well developed vortices. The mode and the degree of influence of SST on transformation of a wave and that on intensification of a circular vortex may not be identical, partly because of the differences in the distribution patterns of the surface wind, evaporation and the latent heat release in these two cases. Furthermore, the dependency of the results on atmospheric temperature and moisture conditions and the existence of a critical value of SST for storm development will also be discussed.

2. Main effects of SST sensitivity

For this investigation, the numerical model described in Kurihara and Tuleya (1981, hereafter referred to as KT) has been utilized. The model domain is a 25 × 25° cyclic channel and contains a uniform mesh grid with 0.625° resolution. The initial steering flow was that of an idealized GATE III period with an approximate easterly flow of 5 m s^{-1} at 18°N. The basic state temperature and moisture were also a GATE III period average at 80°W. The SST of this experiment was prescribed as 302 K. This model successfully simulated the genesis of a tropical storm, from a shallow easterly wave, with a developing warm core and an associated upper level anticyclone. The low level vorticity (~950 mb) increased from 4.3 $\times 10^{-5}$ to 23.7 $\times 10^{-5}$ s^{-1} during the 96 h integration as the surface winds increased to storm strength (>17.5 m s^{-1}).

In order to isolate the effects of SST, additional integrations were run to 96 h for SST's ranging from 298 K (24.8°C) to 303 K (29.8°C). In each experiment, the initial conditions, including the lowest model level (~992 mb) temperature of 299.6 K, were identical, except for SST, to the above described experiment, i.e., Exp. 1 in KT. Table 1 summarizes the results at 96 h, which is representative of the full development stage of each experiment. Some fluctuations appear in the table (e.g., maximum vorticity at 302 K SST) because the time of maximum development does vary somewhat between experiments. Notice that a full range of development states is represented: from that of a decaying wave (298 K SST); to a weak depression (300 K SST); to a mature tropical storm (303 K SST). The amount of development changes monotonically, but not linearly with SST. The surface pressure, as well as other indicators of

TABLE 1. Disturbance intensities at 96 h as a function of SST. Initial intensity and 96 h disturbance movement are also shown. Extreme values are taken from grid point values, whereas disturbance movement is evaluated from surface pressure charts. The warm temperature anomaly is defined as the difference between the temperature above the disturbance center and a reference temperature at the same latitude as the disturbance.

Sea surface temperature (K):		298	299	300	301	302	303
	At 0 h	*At 96 h*					
Surface pressure min. (mb)	1008.4	Wave pattern	1008.0	1006.8	1003.6	1002.6	1000.4
Maximum surface wind ($m\ s^{-1}$)	7.9	8.1	10.0	11.8	15.8	17.8	21.2
Maximum vorticity ($10^{-6}\ s^{-1}$) at ~950 mb	43	28	82	140	292	237	323
Warm temperature anomaly (K) at ~335 mb	−0.1	0.3	1.6	2.0	3.3	3.3	4.7
Maximum evaporation ($10^{-5}\ cm\ s^{-1}$)	—	0.13	0.24	0.43	0.84	1.24	1.78
Maximum precipitation ($10^{-5}\ cm\ s^{-1}$)	—	5.2	16.7	24.6	36.2	32.7	79.2
Westward movement in 96 h (deg)	—	22.6	21.2	20.9	23.1	24.0	25.4
Northward movement in 96 h (deg)	—	Wave pattern	1.8	1.9	3.7	3.9	5.1

intensity, such as 950 mb maximum vorticity, display the largest difference between SST's of 300 K (26.8°C) and 301 K (27.8°C). Despite nearly identical surface pressures and winds at 48 h, a significant difference between the experiments is already evident. Careful analyses of the synoptic, vorticity budget and heat budget fields reveal some interesting details. The maximum divergences associated with the outflow from the disturbances at 48 h are 3.3×10^{-5} and $4.9 \times 10^{-5}\ s^{-1}$ for the 300 and 301 K SST cases, respectively. In the 300 K case, the vertical motion and the surface pressure fields are more vertically aligned than in the 301 K SST case. In this respect, the 301 K SST development is very much like that of KT (Figs. 9, 10, 13 and 16) in which the surface low position is off the area of major convective activity and maximum upward motion. At ~335 mb in an 8.1 × 8.1° area centered above the surface pressure center, the maximum instantaneous net heating rates are 18×10^{-5} and $30 \times 10^{-5}\ K\ s^{-1}$ for the 300 and 301 K SST cases respectively. Furthermore, the areal coverage of net positive heating tendency in the above-mentioned area for the 301 K case is two times that of the 300 K SST case. In the 301 K case a low level vorticity budget analysis indicates a net tendency similar to that of KT (Fig. 13) in which a concentration of cyclonic vorticity occurs near the pressure center and a general decrease in the surrounding region. In the 300 K SST case, this tendency is considerably reduced. The disturbance structures in the 301 K SST case, as well as in the developing cases at 302 K and 303 K SST, are inherently not axisymmetric. In short, a crucial factor for a tropical disturbance to grow into a tropical storm seems to be the degree of formation of a warm anomaly at upper levels within a few hundred kilometers of the surface pressure low, though not necessarily right above the surface low. Whether this condition evolves or not in an individual case, depends on the initial structure of the disturbance, the environmental conditions, such as the conditional instability and the basic wind field, and the surface conditions including SST. The influence of some of these factors will be discussed in the next section.

In the experiments in this section, the behavior of the model disturbances is apparently related to the evaporation, advection and release of latent energy. As shown in Table 1, there is a systematic decrease of maximum evaporation in the experiments as the SST is reduced. A systematic decrease is also observed in the domain average evaporation. The evaporation rate is dependent on the surface wind speed, the mixing ratio of the surface air and the saturation mixing ratio at the sea surface. The decrease of evaporation is, of course, initially caused by the reduction of available moisture from the ocean surface which is a direct function of SST. This explanation holds as long as the low level atmospheric moisture content remains relatively fixed. As the developing systems evolve, the low level wind speed increases which, in turn, causes the evaporation rate to increase in the experiments. Fig. 1 displays profiles of the equivalent potential temperature θ_e and the saturated equivalent potential temperature θ_e^* at the disturbance center for the initial state (the same for all experiments) and for the two extreme cases of 298 and 303 K SST at 36 h. The conditional instability $-\partial\theta_e^*/\partial z$ of the low levels is significantly reduced in the 298 K case and increased in the 303 K case by 36 h. Analysis reveals that the low level change in equivalent potential temperature is due both to the modification of the air temperature toward that of the SST and to the change in absolute humidity. This feature is similar to the portrayal by Gray (1975). The difference in the low

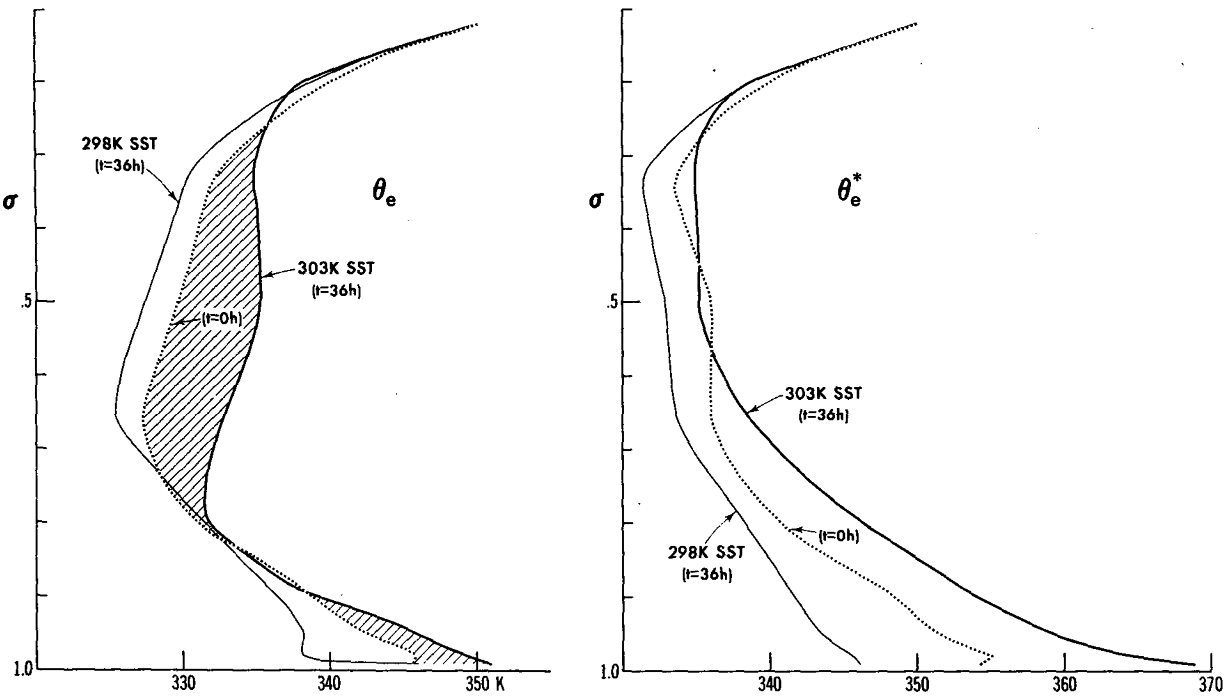

FIG. 1. The equivalent potential temperature θ_e profile and the saturated equivalent potential temperature θ_e^* profile at the disturbance center initially (dotted line) and at 36 h for experiments with a SST of 298 K (thin line) and 303 K (thick line). Shading indicates increase of θ_e from the initial value. Profiles are obtained from grid point values immediately surrounding the surface pressure center.

level profiles of θ_e and θ_e^* between low and high SST experiments is a general feature of the surrounding environment as well as being evident in the disturbance center. The effects of convection in the high SST case at the disturbance center are manifested in the θ_e profile between 250 mb ($\sigma \approx 0.25$) and 850 mb ($\sigma \approx 0.85$). Temperature and/or moisture increases are responsible for the mid-level increase in equivalent potential temperature as the upper level warm, moist core and the low level vorticity intensify. The mid-level cooling in the 298 K SST case is caused primarily through net radiative cooling.

Another interesting feature revealed by Table 1 is the sensitivity of storm track to SST. In the high SST experiment of 303 K, the intensifying storm moved approximately 4.5° longitude further west than that of the 300 K SST case in which only a weak depression evolved. The intense disturbance also moved 3.2° latitude further north in the 96 h period. Note that the weakening disturbance with a low SST of 298 K moved westward faster than that of the 300 K SST case. The phase velocities of the disturbances in this series of experiments are a complicated function of many factors. According to Adem (1956), the beta effect causes a barotropic vortex to drift to the north and west relative to the mean wind, at a rate proportional to the intensity and size of the vortex, respectively. This appears to agree with the results obtained, in that the disturbances all propagate to the northwest relative to the mean flow with the most intense propagating the furthest north. Moreover, it is observed that in the initial two days of the storm producing disturbances, i.e., SST $\geq$ 302 K, upper level convective and net heating maxima appear ~2° to the northwest of the disturbance center (e.g., KT, Figs. 9, 16 and 17). It is speculated that this heating may also affect the storm phase velocity. Another effect which causes deflection of a rotating vortex to the right of the steering current is that of the frictional drag (Kuo, 1969; Jones, 1977). In a study somewhat related to the present one, Chang and Madala (1980) have shown model sensitivity of mature traveling hurricanes to spacial SST anomalies. They found that a tropical cyclone intensifies (decays) when encountering a broad area of warmer (colder) SST and propagates faster (slower) ahead and to the right than a control experiment with uniform SST.

3. Influences on SST sensitivity—other factors controlling genesis

Obviously, only in a model simulation can the effects of SST be isolated from the many other factors which play a part in hurricane development. One of these other factors is the basic flow in which the incipient disturbance is embedded. Fig. 2 displays the behavior of an initially identical disturbance for two different zonal flows at both 298 and 302 K SST. Type NS and type ES zonal flows correspond to Exp. 1 and 6 described by Tuleya and Kurihara (1981),

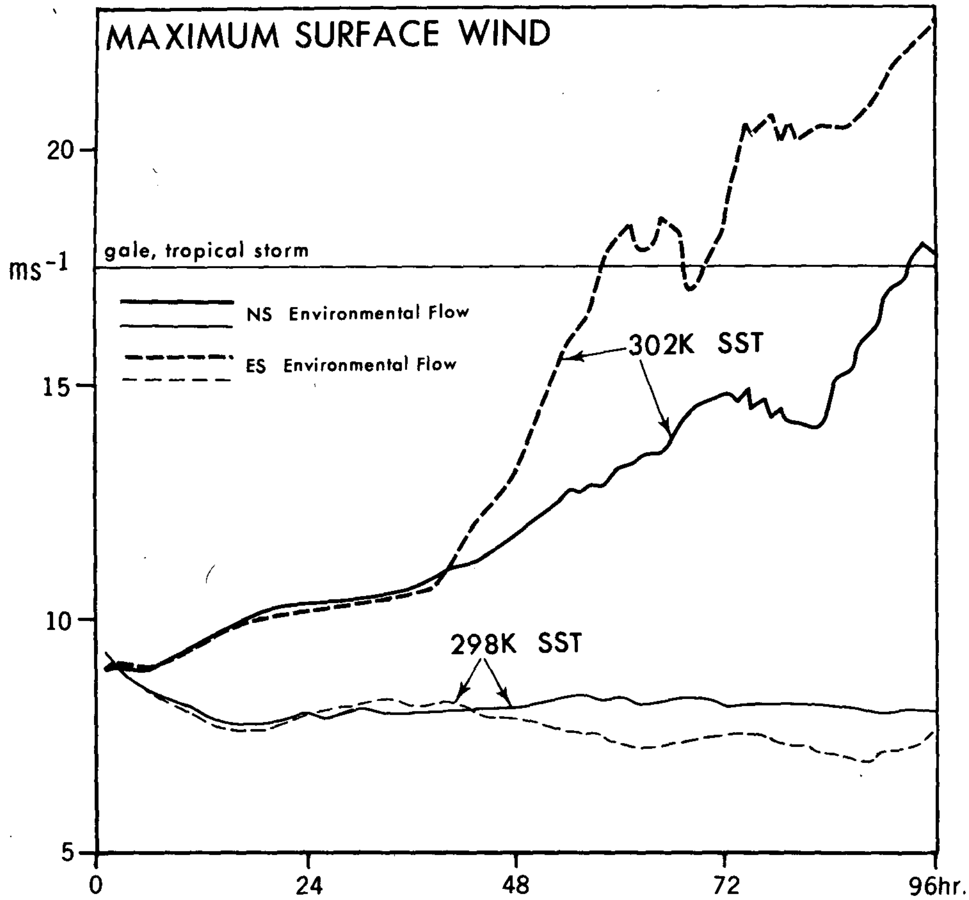

FIG. 2. Time history of maximum surface wind speed for experiments with two different SST (298 and 302 K) and with two different basic flows. NS (solid lines) and ES (dashed lines) indicate experiments with basic flows identical to Exps. 1 and 6 of Tuleya and Kurihara (1981) respectively.

in which the initial vertical shears were zero and moderate easterly, respectively. Notice that at 302 K the characteristic flow of the environment has a large influence on the rate and amount of storm development. However, at 298 K SST, a value below that climatologically found in genesis regions, little development is found for either basic flow. It appears that a SST value exists below which no development occurs, regardless of the environmental flow.

There are many other complicating factors which tend to mask the sensitivity of genesis to SST variations, including fluctuations in the intensity and structure of the initial disturbance and variations in the basic state temperature and humidity. In the experiments described above, the SST was rather arbitrarily changed. In reality, the SST changes in turn modify the lower tropospheric temperature and moisture. This was seen in the evolution of the experiments previously analyzed. Two cases will now be studied in which the initial atmospheric temperature and humidity will be specified differently from that of the previous control cases.

One experiment was performed to investigate the impact of reducing the initial atmospheric temperature by 5 K everywhere with a 298 K SST. The initial relative humidity was identical to that of the control experiments. In this experiment, the initial evaporation rate is approximately equal to that of the control experiment with a higher SST of 302 K. Therefore, one may expect that higher evaporation rates than the control case of 298 K SST would perhaps lead to a different disturbance evolution despite the same low SST. The results of this experiment at 24 and 96 h are shown in Table 2, in which they can be compared with the control cases of 298 and 302 K SST. Notice that at 24 h the maximum surface wind in the -5 K case is approximately the same as the 302 K control case. However, for the same evaporation rates, the maximum precipitation rates are reduced by more than a factor of two. This is because the boundary-layer moisture is reduced when the initial relative humidity is the same as the control experiments. In fact, the maximum integrated moisture convergence below ~600 mb at 24 h is reduced from

TABLE 2. Disturbance intensities at 24 and 96 h for control experiments with 298 and 302 K SST and for two supplementary experiments. In the −5 K experiment, the initial atmospheric temperature was 5 K cooler than the control experiments. A tropical sounding based on the cloud cluster and clear area profiles of Williams and Gray (1973) was used in the supplementary experiment WG. Parameters are evaluated in the same manner as described in the Table 1 legend.

		Control experiments		−5 K experiment	WG experiment
Sea surface temperature (K):		298	302	298	302
24 h	Maximum surface speed (m s^{-1})	8.0	10.3	10.3	9.9
	Maximum evaporation (10^{-5} cm s^{-1})	0.07	0.61	0.61	0.72
	Maximum precipitation (10^{-5} cm s^{-1})	1.3	7.0	3.1	2.5
	Maximum integrated moisture convergence (600–1000 mb) (10^{-6}g kg^{-1} s^{-1})	21	121	59	74
96 h	Surface pressure minimum (mb)	Wave	1002.6	1006.9	1007.5
	Maximum surface speed (m s^{-1})	8.1	17.8	10.2	9.8
	Warm temperature anomaly (K) ~335 mb	0.3	3.3	−0.4	0.7

121×10^{-6} in the 302 K control case to 59×10^{-6} g kg s^{-1}, despite the similar initial mass convergence. The reduced moisture convergence dramatically suppresses the ability of the disturbance in the −5 K experiment to develop, despite the relatively high evaporation rates. After 24 h, the −5 K case disturbance failed to develop any further and a cold anomaly appeared by 96 h at ~335 mb. Nevertheless, the −5 K case disturbance, contrary to the 298 K control case, did develop somewhat and maintain itself as a weak tropical depression. In an experiment in which the lower atmosphere was more moist than in the −5 K case, an even stronger disturbance evolved despite low values of SST. It can be concluded from the −5 K experiment that a large evaporation rate, by itself, will not necessarily lead to development without a sufficiently moist ambient atmosphere. Palmén and Riehl (1957) have demonstrated the importance of moist inflow in addition to local evaporation in the maintenance of mature hurricanes.

In a second experiment involving different atmospheric conditions, an averaged profile for cloud cluster and clear areas based on Williams and Gray (1973) for the tropical Pacific region was utilized. The temperature and relative humidity values are given and compared to those used in the control experiments in Table 3. The SST was fixed at 302 K. The moisture content of the Williams and Gray (WG) profile is considerably lower with the biggest discrepancy (20%) at 500 mb. The mid-tropospheric lapse rate is also slightly more stable in the WG profile than in the control experiments. The integration results of the WG experiment are shown in the last column of Table 2 and can be compared directly with the 302 K SST control case which has identical initial conditions other than the atmospheric temperature and humidity profiles. Despite the high SST of 302 K, only a weak tropical depression developed in the WG experiment. At 24 h the evaporation rate was considerably higher than the 302 K control case because of the low humidity in the surface layer. Similarly to the −5 K case, the moisture convergence was restricted because of the lack of moisture at low and middle levels. As mentioned by Palmén (1948), high SST alone is not a sufficient condition for development of a tropical storm.

TABLE 3. Atmospheric temperature and moisture profiles utilized in the control experiments, i.e., GATE III average at 80°W, and an average cloud cluster–clear area profile WG, based on that of Williams and Gray (1973) for the tropical Pacific. Differences between the two profiles are also shown.

		Temperature (K)			Relative humidity (%)		
Level	Pressure (mb)	Control experiment	WG experiment	Difference	Control experiment	WG experiment	Difference
1	31	218.6	218.6	0.0	10	1	−9
2	120	201.0	200.6	−0.4	20	20	0
3	215	224.7	223.8	−0.9	30	24	−6
4	335	246.9	248.2	1.3	38	28	−10
5	500	267.3	268.4	1.1	51	31	−20
6	665	279.7	280.7	1.0	61	46	−15
7	800	288.6	288.6	0.0	71	58	−13
8	895	294.5	293.7	−0.8	78	67	−11
9	950	297.5	296.9	−0.6	82	74	−8
10	977	299.1	299.1	0.0	83	79	−4
11	992	299.6	300.2	0.6	84	76	−8

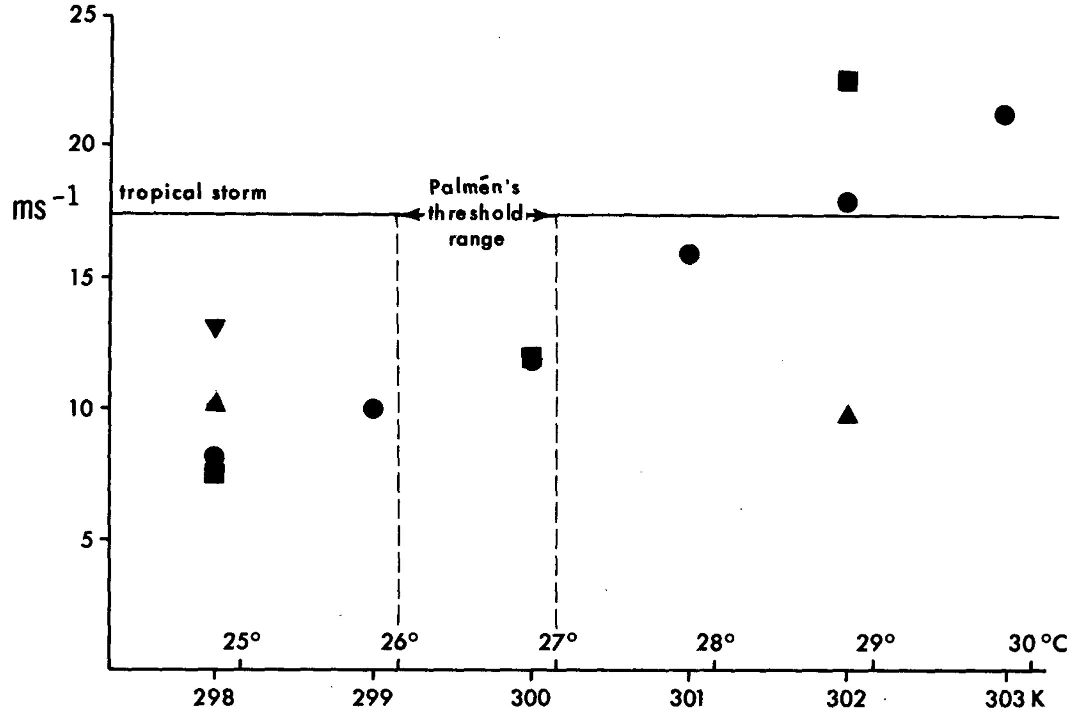

FIG. 3. Maximum surface wind speed at 96 h as a function of SST. Circles define control experiments explained in Section 2. Squares define experiments with type ES zonal flow. Triangles define supplementary experiments, −5 K and WG. The inverted triangle defines the supplementary experiment identical to Exp. −5 K except that lower atmospheric humidity was specified as 100%.

4. Remarks

In the experiments performed, one can see the direct, isolated effect of SST on the genesis of a tropical storm. Utilizing the basic flow of Phase III of GATE at 80°W, one obtains a spectrum of tropical disturbance development stages from a weakening wave to a mature tropical storm with a range of SST from 298 K (24.8°C) to 303 K (29.8°C), respectively. The largest change in intensity was at a SST of ~300 K (26.8°C) although the intensity increased monotonically from 298 to 303 K SST.

The apparently large SST sensitivity of the model is found to be modulated by other factors. One obvious masking effect is that of the basic environmental flow which may be a more practical factor in determining whether a disturbance intensifies. The large scale environmental temperature and humidity also significantly alters the sensitivity of storm genesis to SST. The modulation of the SST sensitivity in this study can be seen in Fig. 3. There is a wide range of development stages (i.e., very weak depression to a tropical storm) for experiments with identical SST. For even low values of SST, the amount of development depends, to some extent, on the moisture content, the static stability and the wind shear of the environment. These factors, in turn, determine evaporation, moisture convergence and ultimate strength of the disturbance. One may draw an upper limit envelope above all the model results with realistic initial states, and define a critical value of SST for development where this envelope intersects tropical storm strength. Such a cut off value may not be that abrupt. However, it is encouraging that this value for the present model corresponds rather well with Palmén's threshold value of 26–27°C. Of course, the placement of this envelope is subjective and may be dependent on many model parameters and the experimental design.

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