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X-651-73-108
PREPRINT

NASA TM X- 66235

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(NASA-TM-X-66235) ANALYSIS OF SOME METHODS FOR OBTAINING SEA SURFACE TEMPERATURE FROM SATELLITE OBSERVATIONS (NASA) 23 p HC $3.25
N73-23467
Unclas
G3/13 02979

APRIL 1973

GSFC
GODDARD SPACE FLIGHT CENTER
GREENBELT, MARYLAND

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ABSTRACT

Satellite measurements of sea surface temperature must be corrected for atmospheric moisture, cloud contamination, reflected solar radiation and other sources of error. Procedures for reducing errors are discussed. It appears that routine accuracies of 1°C are possible, given low noise spectral measurements in the infrared.

I. Introduction

Although present satellite measurements are capable of providing accurate estimates of sea surface temperature in local areas (Smith, et al. 1970; Rao and Smith, 1972), the achievement of routine 1°C accuracies on a world-wide basis – a specific global atmospheric research program (GARP) requirement – has not yet been demonstrated. This paper investigates some of the problems which tend to interfere with accurate determination of sea surface temperature (SST). It is likely that with proper satellite instrument design and data handling procedures these error sources can be reduced to the desired levels.

Section II deals with techniques for measuring the atmospheric correction to be applied to "window" radiance measurements, and the necessary instrument specifications to hold the effect of random system errors to a low level.

Section III deals with the procedure suggested by Smith and Rao (1971) for using 3.7 and 11 micrometer spectral information to eliminate the effects of clouds. Recommendations for improvement are made, although the method still will not always produce unique answers. Section IV analyses the stability of statistical regression coefficients as an indicator of accuracy for prediction.

It must be noted that several types of error are not dealt with here. The first are system errors arising in calibration of the satellite instrument, telemetry, analog to digital conversion, satellite pointing errors, etc. For an analysis of a current system see the review by Leese et al. (1971). The second type of error arises from identifying radiance measurements with the nominal "bucket" temperature used by oceanographers. The difference between the "skin" radiance temperature and the internal temperature may go over 1°C (Paulson and Parker, 1972), but it seems unlikely that this will contribute significant error on a global scale.

Finally, reflected solar energy can cause a noticeable error (Whitehead, 1972; Greaves, 1972) in 11μm radiances under very calm surface conditions. For this reason we recommend the rejection of data which come from within 5° of the point of specular reflection of the sun.

II. A. The Use of Two Infrared Measurements for Eliminating the Effect of Atmospheric Water Vapor

Although measurements in the 11 micrometer atmosphere "window" yield reasonable estimates of SST, a correction for attenuation from atmospheric moisture must be applied if high accuracies are desired. The calculations described here show that the correction may be as large as 8°C for a very warm, moist atmosphere. Of course atmospheric water vapor is highly correlated with the temperature of the sea surface, so that the error in using 11μm radiances arises from the application of a standard correction as a function of SST. Thus

$$\begin{aligned}(\text{Temperature of sea surface}) &= (\text{Temperature from } 11\mu\text{m radiance}) \\ &\quad + \Delta T(11\mu\text{m temperature}) \qquad (1)\end{aligned}$$

where $\Delta T(T)$ is a prescribed function. In practice the quantity ΔT is not determined solely by the surface temperature, but also by the prevailing climate of the region, and by variations in the weather. Whether the error committed by taking ΔT in (1) to be a fixed function amounts to 1°C on a global basis is not known. Certainly the neglect of time and space variations will lead to systematic errors (e.g. dry high pressure areas will yield high estimates of SST), which will interfere with long range weather forecasting.

As described by Anding and Kauth, (1970) satellite measurements in two spectral windows, one "clean" and one "dirty," can provide additional information to make measurements accurate to 1°C on a case by case basis. The choice of these spectral channels is still open, as several factors must be taken into account.

In order to establish quantitative procedures for deriving sea temperature, we start with the equation of radiative transfer, in the usual notation,

$$I_\nu = B_\nu(T_s)\,\tau_\nu(p_s) - \int_0^{p_s} B[T(p)]\,\frac{\partial \tau_\nu(p)}{\partial p}\,dp \tag{2}$$

where I_ν is the observed radiance at frequency ν. We may rewrite 2 as

$$I_\nu = B_\nu(T_s)\,\tau_\nu(p_s) + [1 - \tau_\nu(p_s)]\,\langle B_\nu \rangle$$

where $\langle B_\nu \rangle$, the weighted atmospheric emission, is defined by

$$\langle B_\nu \rangle \equiv \int_0^{p_s} B[T(p)]\,\frac{\partial \tau_\nu(p)}{\partial p}\,dp \Bigg/ [1 - \tau_\nu(p_s)] \tag{3}$$

In the linear approximation of small absorption appropriate to atmospheric windows we have $1 - \tau_\nu (p_s) = \epsilon_\nu = k_\nu u$, where u is the path length and ϵ is much less than one. Also

$$\frac{d\tau_\nu}{dp} = - k_\nu \frac{du(p)}{dp} = - k_\nu q$$

where q is the mixing ratio. In this approximation

$$\langle B_\nu \rangle = \frac{1}{u} \int_0^{p_s} B[t(p)] \, q(p) \, dp$$

and we note that the atmospheric emission is independent of the absorption properties k_ν of the water vapor. Thus

$$I_\nu = B_\nu (T_s) - \epsilon_\nu [\langle B_\nu \rangle - B_\nu (T_s)]$$

In the spectral intervals we consider the moisture is at low levels, and we may expand the Planck function B about the surface temperature. For two spectral channels measuring radiance temperatures T_1 and T_2 we have

$$T_1 = T_{SST} - \epsilon_1 T_A$$

$$T_2 = T_{SST} - \epsilon_2 T_A$$

where T_A is the effective atmospheric temperature. We solve for T_{SST} and, following MacMillan (1971), define $\gamma = \epsilon_1 / (\epsilon_2 - \epsilon_1)$

$$T_{SST} = T_1 + \gamma (T_1 - T_2) \tag{4}$$

This theoretical relation is complicated in practice by several factors:

(1) spectral regions with stronger molecular absorptions tend to have strong line absorption, which does not obey the assumptions we have made.

(2) there is considerable evidence (Bignell, 1970) that the absorption coefficient k_ν for water vapor depends on q, so that the linear approximation is satisfied for only very small water vapor amounts.

It appears that both of these effects can be accounted for simply by generalizing (4) to

$$T_{SST} = T_1 + \gamma(T_1 - T_2) + C \tag{5}$$

where C is a constant to be determined from comparison of satellite data with measured SST's.

Table 1 presents the results of calculations of the atmospheric correction for clear sky atmospheres taken from the study of Wark et al. (1962). The calculations were carried out using the high spectral resolution radiative transfer program developed by Kunde and Maguire (1973). The high resolution results were averaged across the spectral band to produce the figures indicated.

The spectral interval 870 - 950 cm^{-1} (10.5 - 11.5 micrometers) has very little atmospheric absorption and has been used frequently on satellite instruments as an indicator of surface or cloud temperatures. The interval 775 - 831 cm^{-1} (12.0 - 12.9μm) has been recommended by Prabhakara et al. (1972) for estimating the atmospheric correction for SST. This recommendation is especially valuable as it is based on the examination of satellite data (Nimbus IV IRIS). The interval 1150-1250 cm^{-1} (8.0-8.7μm) was suggested by Kunde as a candidate because it is affected mainly by water vapor, but has stronger absorption than the 11 micrometer region of the spectrum.

By using a least squares fitting procedure we obtain the following equations for the surface temperature

$$T_{SST} = 3.67\,T_{11} - 2.84\,T_{12.5} + 45.6 \quad \text{rms error} = 0.7^{\circ}\text{C} \tag{6a}$$

$$T_{SST} = 5.35\,T_{8.4} - 3.90\,T_{11} - 105.6 \quad = 1.6^{\circ}\text{C} \tag{6b}$$

$$T_{SST} = 3.27\,T_{8.4} - 2.23\,T_{12.5} - 8.22 \quad = 0.3^{\circ}\text{C} \tag{6b}$$

It is clear that the third case is the best choice for the spectral channels according to the model. This is not surprising as the contrast in the water vapor absorption is strongest for this choice of spectral intervals.

Limitations on computer time prevented the running of more model calculations. Due to the uncertainty in knowledge of the absorption coefficients it would be desirable to test the predictions above with an experimental program.

With confidence in the qualitative behavior of the theoretical model (Kunde et al., 1973) we find that spectral information can be used to correct for attenuation from atmospheric moisture.

B. The Effect of System Errors on Two Channel Measurements.

In addition to the errors from imperfect information as to the atmospheric state we must consider errors in the observing instruments, data handling, data processing, etc. From equation 4 we may estimate the error δ (SST) arising from errors δT_1 and δT_2 in the respective spectral measurements.

$$\delta(\mathrm{SST}) = [\overline{\delta T_1^2}(1 + \gamma^2) + \delta T_2^2 \gamma^2 - 2\gamma(1 + \gamma)\ \overline{\delta T_1 \delta T_2}]^{1/2}$$

The quantity $\overline{\delta T_1 \delta T_2}$ may be assumed to be zero, except for bias errors; obviously we wish to have both biases of the same sign (both channels high, or both low). Ignoring $\overline{\delta T_1 \delta T_2}$, and assuming $\delta T_1^2 \sim \delta T_2^2$ we have

$$\delta(\mathrm{SST}) = [\overline{\delta T_1^2}(1 + 2\gamma + 2\gamma^2)]^{1/2} \tag{7}$$

Retracing the logic we see that we wish to have $T_1 - T_2$ as large as possible, in order to make γ as small as possible. However, when this quantity is large there must be strong absorption in one of the channels, so that the linear approximation fails. The tradeoff is between a strongly absorbing spectral region, in order to minimize the effect of system noise, and a weakly absorbing region such that the linear absorption approximation is justified.

From 6c and 7 we see that $\gamma \sim 2.25$ according to the model calculations. Thus error in the measuring factor is amplified by a factor $\sqrt{1 + 2\gamma + 2\gamma^2} \sim 3.8$ so that low noise measurements are essential.

Recently Anding and Kauth (1972) have carried out calculations similar to those described here. They recommend spectral channels at 8.9μm and 11.9 μm. Based on their published data we may obtain an estimate of $\gamma \sim 3.6$, which yields an error amplification factor of about 5.5.

The necessity for low noise measurements is clearly indicated by these results. We are optimistic that these requirements can be met with suitable instrument design.

III. Elimination of Cloud Effects by Combining Spatial and Spectral Information

Clouds represent a persistent problem in satellite meteorology. Total coverage by high clouds prevents the use of infrared data to determine SST in the affected region. A less obvious problem is the fact that partial cloudiness, or complete coverage by very low clouds may cause slightly lowered (and hence false) surface temperature results. The development of procedures and thresholds for rejecting these cloud contaminated results is a subject of continuing research.

Smith and Rao (1971) have described a method which uses pairs of spectral measurements (the 3.7 and 11 micrometer atmospheric windows) from spatially nearby points. The method rests on the assumption that clouds have constant spectral properties in a set of measurements in a given locale. The assumption is quite reasonable, and one can probably account for variations in cloud properties (height, thickness, etc.) by increasing the estimated magnitude of random error in the measuring system.

Granted this assumption, we may use the nonlinear temperature dependence of the Planck function to solve for the surface temperature. This results from the fact that clear and cloudy areas contribute different fractions to the total energy in different spectral intervals.

Smith and Rao suggest using numerous pairs of measurements in a given region to obtain a series of estimates of SST. The average result is then their best estimate of the true sea temperature. Because of non linear error propogation it is preferable to perform averaging on the raw data, and use the smoothed data in extracting the result.

Figure 1a shows the behavior of radiance temperatures in two spectral bands as a function of the percentage cloud cover, assuming that clouds radiate as black bodies. The variation in $T_{3.7} - T_{11}$ results from the fact that radiation varies as a higher power of temperature in the 3.7μm window.

Calculations (Curran 1972) show that clouds have a lower radiation temperature at 3.7μm than at 11. This effect may be modeled by assigning the cloud an emissivity less than unity at 3.7μ, with the result shown in figure 1b. We shall assign the cloud an emissivity of .75 at 3.7μm. In figure 2 radiance temperatures are plotted for two surface temperatures 280°K and 290°K for a number of cloud temperatures, with $T_c < T_{surface}$ for all cases. This figure suggests an algorithm for obtaining the surface temperature from satellite data. It will not be difficult to include information from a 9 or 12μm channel in order to correct for molecular attenuation. In this discussion we ignore molecular absorption.

A. System noise tends to thicken the lines on figure 2. As described in section IIB the error in $T_{3.7} - T_{11}$ is magnified because it is a differenced quantity. Furthermore we expect greater variability in cloud properties as cloud amount increases. For these reasons we identify warmest temperature, T_{ω}, from a set of measurements, and discard all data which are much colder, say $T_{\omega} - T > 10°$. If too few data pass this test no result can be obtained for surface temperature.

B. In order to average out random errors we fit a parabola through the remaining data. The root(s) where $T_{3.7} = T_{11} \geq T_{\omega}$ represent possible values for the surface temperature. There are now two possible difficulties.

1. The slope $d(T_{3.7} - T_{11})/dT_{3.7}$ of the parabola is near zero in the vicinity of the point $T_{3.7} = T_{11}$. In this case errors in the measured values result in greatly magnified errors in the estimate of SST. For cases where this slope is less than 0.3 we reject any extrapolation and fall back to the one channel histogram method of Smith (1970).
2. In figure 3 the case of 270° clouds may be identified with SST = 280° for small cloud amount and with SST = 290° for large cloud amount. This ambiguity may occur when the slope is positive and $T_{3.7} - T_{11}$ is negative. In this case again we revert to the one channel histogram for an answer.

For data processing on an operational basis this case may generally be resolved by comparing the predicted value with those obtained earlier in time. It is highly desirable to inject new results into a running time average in order to reduce the effects of random errors and to produce values of SST in regions where cloudiness causes frequent data gaps.

C. Because the extrapolation by a parabolic fit tends to magnify errors in the data we require that a reasonable fraction, say 10% of the measurements fall within 5° of the predicted temperature. It is apparent that this chain of logic does not always yield an answer. The method works better for high, cold clouds than for low, warm clouds, and it tends to fail as the total cloud cover increases. This result is very plausible on physical grounds.

By using the improved procedure described here results for SST can be obtained even in the presence of considerable cloud cover.

IV. Stability of Statistical Correlation Coefficients as an Indicator of Prediction Capability

Users of satellite data frequently have the task of deriving results from an incomplete set of measurements. For example a knowledge of low level water vapor is required for an accurate measurement of SST. This quantity is not determined by the measurements of current radiometers.

In this case one may tie together by empirical relationships the satellite measurements and independent ground truth measurements. These relationships may then be used to extend satellite results to regions lacking observations, or to produce results on a time scale for which ground truth data are not available.

It is desirable to know the probable error resulting from the extrapolation of such statistical relations into regions in time or space where independent information is not available.

For present purposes a simplified treatment is adequate. Let the desired quantity T be related to variables A and B by

$$T = \alpha A + \beta B \tag{8}$$

and assume that our instrumentation measures only A. We may take advantage of correlation between A and B by writing

$$T = \gamma A + C \tag{9}$$

where γ and C are obtained by fitting the measured values of A to independent measurements of T. We use $\langle\ \rangle$ to represent an average over measurements used in the statistical analysis. We obtain

$$\gamma = \alpha + \beta \left[\frac{\langle AB \rangle - \langle A \rangle \langle B \rangle}{\langle A^2 \rangle - \langle A \rangle^2} \right]$$

$$C = \beta \left[\langle B \rangle - A \frac{\langle AB \rangle - \langle A \rangle \langle B \rangle}{\langle A^2 \rangle - \langle A \rangle^2} \right]$$

which illustrates the dependence of γ and C on the correlation between A and B. From (9) and its average we have

$$T = \frac{[\langle A\rangle\langle T\rangle - \langle AT\rangle]}{\langle A^2\rangle - \langle A\rangle^2}(A - \langle A\rangle) + \langle T\rangle$$

so by identity

$$\gamma = \frac{\langle A\rangle\langle T\rangle - \langle AT\rangle}{A^2 - \langle A\rangle^2}$$

$$C = \langle T\rangle - \gamma\langle A\rangle$$

We may now estimate the error in T arising from a variation of the fitting coefficients

$$(\delta T)^2 = \left(\frac{dT}{d\gamma}\,\delta t\right)^2 = (A - \langle A\rangle)^2\,(\delta\gamma)^2$$

In a recent paper Shenk and Salomonson (1972) obtained equations relating SST to 11 μm radiances and other spectral measurements.

$$SST = .836\,T_{11} + \cdots \qquad 31 - 37°N$$

$$SST = 1.17\,T_{11} + \cdots \qquad 37 - 43°N$$

Comparing with equation (9) we have $\delta\gamma \sim (1.17 - .84) = .33$ and since $T_{11} \cong$ SST, we may estimate the variability of T_{11} from the variation of the derived values for SST. From their figure 10 we find the variation in each latitude band is about 4°C. Thus $\sqrt{(\delta T)^2} \sim (4°C) \times .33 \sim 1.3°C$.

This is a conservative (large) estimate for the possible error, as some of the variation of γ undoubtedly arises from climatic differences in the two latitude bands. If we consider the variation of γ about the average value, we find $\sqrt{8T^2} \sim 0.7°C$.

When other error sources are included such as instrument noise, errors in the ground truth data, etc., it is clear that a direct measurement of SST is preferable.

VI. Conclusion

It appears feasible to obtain sea surface temperature from satellite measurements, with an accuracy of 1°C. The effect of high cirrus (Braun, 1971) can be minimized by using 6.7μm water vapor measurements. Additional information from the microwave region of the spectrum will prove useful in overcoming the effects of clouds. Other cloud effects and system errors can largely be eliminated by the procedures described here. A suitable low noise multichannel instrument is planned for future satellite missions.

REFERENCES

Anding, D. and Kauth, R., Remote Sensing of the Environment, 1, p. 217-220, 1970.

Anding, D. and Kauth, R., Remote Sensing of the Environment, 2, p. 171-173, 1972.

Bignell, K. J., Quart. Jour. Roy. Met. Soc., 96, p. 390-402, 1970.

Braun, C., NOAA TM NESS 30, 1971, 28 pp.

Greaves, J. (private communication, 1972) has observed 11 μm solar reflections in Nimbus III data.

Kunde, V. G., and Maguire, W. C., submitted to Journal of Quant. Spect. and Radiative Transfer (1973).

Kunde, V. G., Conrath, B. J., Hanel, R. A., Maguire, W. C., Probhakara, C., and Salomonson, V. V., to be submitted to JGR.

MacMillan, L. M., Ph.D. Dissertation, Iowa State University, Ames, Iowa, 1971.

Paulson, C. A. and Parker, T. W., JGR, 77, p. 491, 1972.

Prabhakara, C., Conrath, B. S., and Kunde, V. G., NASA X-651-72-358, 1972.

Rao, P. K. and Smith, W. L., MWR 100, 1. p. 10-14, 1972.

Shenk, W. E., and Salmonson, V. V., J. of Phys. Oceanography, 2, 2, p. 157-167, 1972.

Smith, W. L., Rao, P. K., Koffler, R. and Curtis, W. R., MWR, 98, 8, p. 604-611, 1970.

Wark, D. Q., Yamamoto, G., and Lienesch, G., JAS 19, 5, 369-384, 1962.

Curran, R. J., American Met. Soc. Conference on Atmospheric Radiation, Fort Collins, Colo., August 7-9, 1972, p. 103-107.

Leese, J., Pichel, W., Goddard, B., and Brower, K., AGARD 17th Symposium, Colorado Springs, Colo., June 21-25, 1971, p. 25-31.

Smith, W. L. and Rao, P. K., 5th Symposium on Temperature and Its Measurement and Control, Washington, D. C., June 21-24, 1971.

Whitehead, Victor S., American Met. Soc. Conf. on Atmospheric Radiation, Fort Collins, Colo., August 7-9, 1972, p. 173-177.

Table 1

Sounding (1200 GMT)	Surface Temperatures	Radiance Temperatures		
		12.0 - 12.9 μm	10.5 - 11.5 μm	8.0 - 8.7 μm
Havana, Cuba 9-29-58 H_2O = 4.25 cm	299.00	288.21	292.52	290.04
Kenira, Morocco 9-29-58 H_2O = 2.89 cm	293.00	286.79	290.19	287.32
Oakland, California 9-29-58 H_2O = 2.60 cm	289.00	288.65	290.00	287.37
Maniwaki, Quebec 9-29-58 H_2O = 1.39 cm	274.00	272.35	273.56	271.65
Ft. Smith, Arkansas 9-29-58 H_2O = 0.63 cm	267.00	266.73	266.98	265.83
Chaguaramas Bay, Trinidad 4-1-58 H_2O = 3.90 cm	298.00	288.17	292.15	289.91
Washington, D. C., 4-1-58 H_2O = 0.97 cm	278.00	275.91	277.31	275.24
Kindley Berm, Bermuda 8-1-58 H_2O = 5.08 cm	300.00	287.58	291.93	289.88
Shreveport, Louisiana, 8-1-58 H_2O = 3.79 cm	298.00	291.40	294.87	292.16
Portland, Me. 8-1-58 H_2O = 1.78 cm	292.00	288.45	290.76	288.15
Eureka, N.W.T. 8-1-58 H_2O = 1.28 cm	278.00	275.74	277.20	275.26

FIGURE CAPTIONS

Figure 1a. This illustrates the variation of the difference between window channel measurements as a function of cloud amount, if clouds radiate as black bodies.

1b. The effect of decreasing the emissivity at 3.7μ m is shown in figure 1b.

Figure 2. Calculated values of $T_{3.7} - T_{11}$ as a function of cloud amount, for a cloud emissivity of .75 at 3.7μm. With a high percentage of 270° cloud cover it is not possible to distinguish between $T_s = 280°$ and $T_s = 290°$.

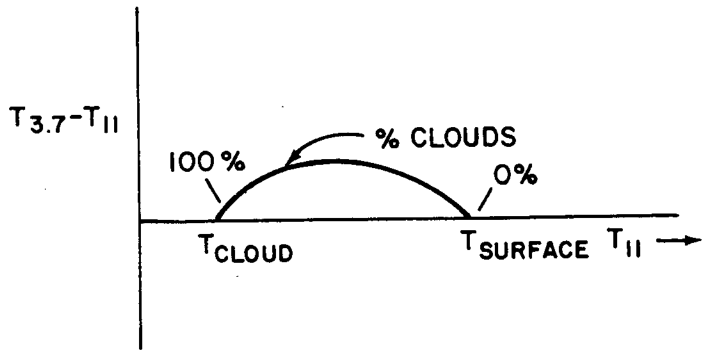


Figure 1a. This illustrates the verification of the difference between window channel measurements as a function of cloud amount, if clouds radiate as black bodies.

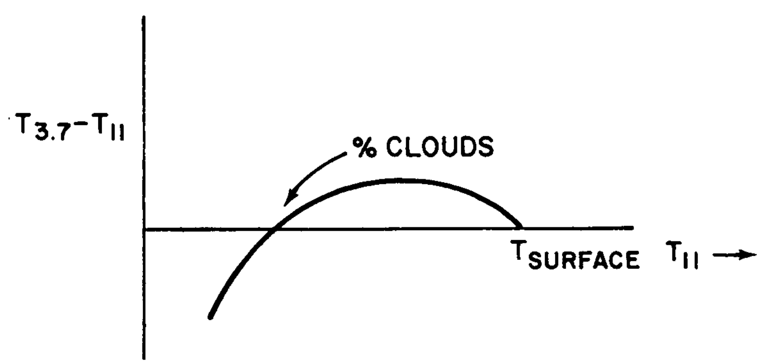
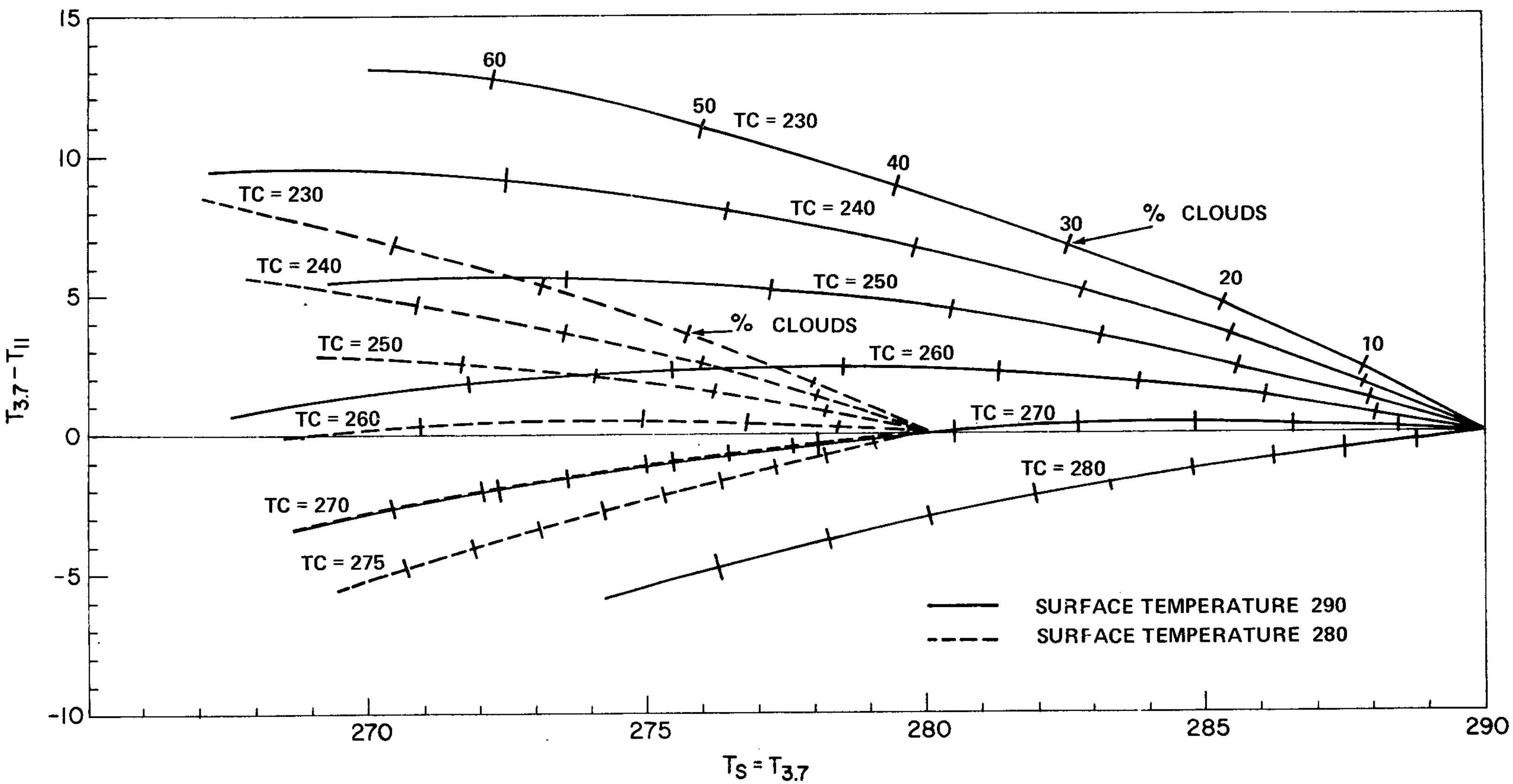


Figure 1b. The effect of decreasing the emissivity at 3.7 μm is shown in figure 1b.



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