Aircraft measurements of SST using IR-radiometer and temperature corrections due to atmospheric effects

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Abstract. Using an airborne scanning IR-radiometer, measurements of sea surface temperature (SST) were made from nine different levels in the Sandheads region of the Bay of Bengal on 5 October 1978. To retrieve SST from the observed radiances a temperature correction scheme, which uses the radiosonde data in the vicinity of flight area, has been generated. Atmospheric effects which have been considered include absorption due to water vapour and carbon dioxide, and the re-emission from different atmospheric layers. The radiances observed at different altitudes when corrected by our scheme yield a fairly consistent value of SST. The special ship measurements of SST, at the same location, are found to have very good agreement with the SST retrieved from the observed radiances using our scheme. The temperature corrections turn out to be 0.3 and 3.3°C at 600 and 3000 meters respectively for the type of atmosphere which has been used in our study.

Keywords. Sea surface temperature ; air-borne radiometer ; atmospheric effects ; precipitable water vapour.

1. Introduction

Sea surface temperature can be measured using airborne instruments such as radiometers, provided the net atmospheric attenuation of radiation from sea surface and the re-emission from atmospheric layers could be estimated in a realistic fashion. For a radiometer which works in the 10.4 to 12.2 μm band, water vapour is the strongest absorbing agent. In the actual atmosphere the vertical distribution of water vapour is known to have large spatial and temporal variations. Prabhakara *et al* (1974) followed an approach where knowledge of the vertical distribution of water vapour and temperature is not required for the estimation of SST. In this approach employing some simplifying assumptions and using the differential absorption properties of two IR channels, which have appreciable difference in the effective water vapour absorption coefficient, estimates of SST with an accuracy of 1°C had been made. But in cases where only one channel is available, as the present one, one has to use temperature and humidity distributions which are fairly close to ambient conditions.

A number of studies have been made to estimate the atmospheric effects up to aircraft altitudes (Griggs 1968; Weiss 1971 and Platt 1972) and satellite altitudes (Shenk and Solomonson 1972; Platt and Troup 1973; Maul and Sidran 1973). Griggs (1968) used an IR radiometer (8–14 μm) aboard an aircraft and made the SST measurements from a number of flight levels. In the same experiment Griggs also obtained the *in situ* measurements of SST by the bucket method. The radiometer-derived temperatures of Griggs had departures as large as 2° C during the down and upward legs of flight and were also not in good agreement with the bucket measurements. Weiss (1971) made airborne measurements of ocean and land surface temperatures using 10–12 and 8–14 μm radiometers. The temperature correction scheme adopted by Weiss was not very realistic as all the components of the atmospheric effects, such as absorption and re-emission, were characterised by an effective emissivity for the atmosphere in both the bands. However, the results of Weiss demonstrated very clearly the superiority of the 10–12 μm channel for the measurement of SST.

During the southwest monsoon of 1977 the authors, for the first time, made aircraft measurements of SST using an IR scanning radiometer in the Arabian Sea and the Bay of Bengal. This necessitated estimating the temperature corrections at different altitudes upto about 3000 m. Such an estimate needs the knowledge of meteorological parameters, upto 3000 m, in the vicinity of aircraft flight region. In order to ascertain that the temperature corrections thus estimated were realistic, we conducted an aerial experiment in the Sandheads region (latitude 21° N, longitude 88° 13' E) of the Bay of Bengal on 5 October 1978. In this experiment the aircraft equipped with an IR-scanning radiometer (10.4–12.2 μm) flew over a fixed area in the Sandheads region a number of times at different altitudes. During the timeframe of our experiment we had also made a special arrangement for making the SST measurements, in the same spatial area, from a stationary ship in the Sandheads region.

In this paper we present the results of our experiment on 5 October 1978 as well as the temperature correction scheme. We would be restricting ourselves to only those observed radiances which were obtained by nadir or near nadir viewings. The results of slant path viewing will be reported later. We find that our measurements from various altitudes, when corrected by the present scheme, yield a consistent value of SST which is also in good agreement with the *in situ* ship measurements made using the bucket method. The SST results of 1977 experiment will be published separately.

2. Instrument and calibration

Figure 1 shows the schematic of the scanning IR-radiometer system which was fitted on a DC-3 aircraft for the measurement of the spectral radiances. The detector is *HgTe* — *CdTe* photovoltaic type (S. N. PV 1162) and is housed in a double-walled metal dewar which is evacuated to 10^{-3} torr. The detector is cooled by liquid nitrogen to reduce the thermal noise. A filter having a response from 10.3 to 13.0 μm is placed in front of an anti-reflection coated Ge window which is positioned next to the detector as shown in figure 1. The overall response of the detector, window and filter section is shown in figure 2, the lower and

Figure 1. The schematic of air-borne IR-scanning radiometer.

Figure 2. Spectral response of entire radiometer system, i.e., detector, filter and window.

upper cut-off points being 10.4 and $12.2 \mu\text{m}$. The radiometer looks downward through a gold-plated mirror which is inclined at 45° to the optical axis of the system shown by a dashed line in figure 1. The mirror is connected to the optical axis of the system and rotates at a speed of 1 rps with the help of a DC motor.

The movement of the mirror provides scanning on the ground, produces synchronisation pulses and generates some other house-keeping data. The total field of view of the system is $\pm 45^\circ$ with respect to nadir and the instantaneous field of view is 10 milliradians. Figure 3 shows the area scanned by the detector and the smallest resolvable mosaic on the ground for a typical case when the aircraft is moving at a ground speed of 66 m/sec at an altitude of 3000 m. The black body (figure 1) is a beehive-type blackened copper plate and can be kept at 20° , 30° or 40° C with an accuracy of $\pm 0.1^\circ$ C with the help of an electronic temperature control system. Due to the rotation of the mirror the detector sees the scene and the black body alternately. When the detector looks at the black body the output of the detector after the preamplifier stage is stored in a capacitor. Next, when the detector looks at the scene, the difference of scene and black body signal after the preamplifier stage (stored in a capacitor) is amplified, digitised and recorded with the help of a logic and recording system (figure 1). Thus on a magnetic tape digital values are stored which we term as counts and denote by N .

The radiometer was calibrated just before the flight on 5 October 1978 at 0900 hr Indian Standard Time (IST). The calibration is done on the ground by keeping a water-bath underneath the detector system. The temperature of the water-bath is maintained constant for about 2 min and then it is changed to next value and so on. Corresponding to a fixed temperature of the water-bath, a number of N values are generated over a period of 2 min and from these one selects the statistically significant value of N . During our ground calibration the water-bath temperature was varied between 15° and 35° C. Figure 4 shows the plot of N values corresponding to different water-bath temperatures for our calibration of 5 October 1978. The solid circles represent the experimental points and the straight line is the least square fit and is given by

$$N = mT_{so} + C.$$

It can be demonstrated very easily that N should be linearly related to the scene temperature T_{so} in the wavenumber domain of our interest. From the calibration the m and C values are determined which turn out to be 3.18 and

Figure 3. Total and instantaneous field of view (IFOV) of the detector when the aircraft is flying at an altitude of 3000 m with ground speed of 66 m/sec.

Figure 4. Calibration curve for the radiometer for the ground calibration conducted on 5th October 1978 at 0900 hrs IST.

— 49.58 respectively. Using these m and C values any digital count N can be converted into the observed temperature at that altitude level.

The radiometer has some noise and due to that there is some error in the measurement of SST. The standard deviation of N at 303° K was 2.0 and hence the error due to noise was $\pm 0.6^\circ\text{C}$ at 303° K. The error in the measurement of the temperature of the black body which was maintained at 303° K was $\pm 0.1^\circ\text{C}$ and, therefore, the absolute error in the measurement of temperature was $\pm 0.7^\circ\text{C}$ at 303° K. The standard deviation of N at other temperatures was ≤ 2.0 and hence in all the temperature range the absolute error of measurement was $\pm 0.7^\circ\text{C}$ or less.

3. Aerial experiment

An aerial experiment was conducted on 5 October 1978 from 1300 to 1430 hrs IST in the Sandheads region to measure SST from a number of altitudes over the same location. First of all the aircraft fitted with the radiometer flew at 600 m level in the N-S direction for about 4 min and then it ascended up to 900 m and flew over the same spatial area in the S-N direction, again for about 4 min. In a similar fashion the aircraft ascended upto 3000 m altitude and then it flew at 3000 m in E-W direction. During the descent leg, the aircraft movement was in E-W and W-E direction and the flight levels were 3000, 2400, 1800, 1200 and 600 m respectively. We had also on board a TV camera and a video recorder with the help of which we ascertained that we were considering only those cases where there were no clouds in the field of view of the radiometer.

4. Temperature correction scheme

When remote sensing of sea surface is carried out from an aircraft the radiative transfer equation assumes the following form for absorbing and emitting but

nonscattering atmosphere.

$$\begin{aligned} \int_{\nu_1}^{\nu_2} J_\nu P_\nu(T_h) d\nu &= \int_{\nu_1}^{\nu_2} \epsilon(\nu) J_\nu P_\nu(T_o) \tau_{o,h}(\nu, T) d\nu \\ &+ \int_{\nu_1}^{\nu_2} \{1 - \epsilon(\nu)\} J_\nu P_\nu(T_o) \tau_{o,h}(\nu, T) d\nu \\ &+ \int_{\nu_1}^{\nu_2} \int_0^h J_\nu P_\nu(T_z) \frac{\partial \tau}{\partial z} dz d\nu, \end{aligned}$$

where J_ν is the spectral response of the radiometer system; $P_\nu(T)$ is the Planck's function which gives the spectral radiance of a blackbody at wavenumber ν and temperature T ; $P_\nu(T_o)$ is the radiance of the air as seen by a radiometer on the ground; ν_1, ν_2 are the lower and upper cut-off wavenumbers of the radiometer system; h is the altitude of the aircraft; z is the altitude of any intermediate isothermal layer; T_h is the experimentally observed temperature at an altitude h ; T_z is the temperature of the isothermal layer at altitude z ; T_o is the sea surface temperature; $\tau_{o,h}(\nu, T)$ is the atmospheric transmittance at wavenumber ν , of a slab, in altitude range $o-h$, and having a temperature T and $\epsilon(\nu)$ is the emissivity of the sea surface at wavenumber ν .

The left side of the above equation is the Planks spectral function corresponding to the observed temperature T_h at any given altitude h . The first term on the right side corresponds to the emission from the sea surface after its modification due to atmospheric transmittance $\tau_{o,h}(\nu, T)$. The second term represents the radiance of the air as seen by a radiometer on ground, reflected back $(1 - \epsilon(\nu))$ and modified by atmospheric transmittance $\tau_{o,h}(\nu, T)$ upto the aircraft. The third term represents the integrated re-emission from different atmospheric layers.

If one divides the atmosphere up to an altitude h , into a number of isothermal layers, the integral o to h , in the third term on right side, can be approximated by summation of contribution from all the layers. The third term would then be

$$\int_{\nu_1}^{\nu_2} [\sum_{i=1}^M J_\nu P_\nu(T_i) \Delta \tau_i(\nu, T)] d\nu$$

where M is the number of atmospheric slabs and, $\Delta \tau_i(\nu, T)$ is the difference in transmittance from the top of the i th layer to the observational level and, from the bottom of the same layer to the observational level.

In the present experiment, observations were made from altitudes varying between 600 m and 3000 m. We divided the atmosphere below 3000 m into 20 slabs of 150 m thickness.

One has to assume a certain radiative transfer model to compute the right side terms excepting $P_\nu(T_o)$. Once that is done, one can estimate the SST by an iterative procedure. Thus for a known observed temperature T_h , at an altitude h one can estimate the temperature correction. For evaluating the right side terms,

one has to take into account the different atmospheric effects. In the present analysis we included the following effects.

Based on the experimental measurements of Buettner and Kern (1965), the emissivity of the sea surface has been assumed to be unity in our band. In the absence of any experimental measurements, the emissivity of the sea surface is usually assumed as unity in the window region (Maul *et al* 1978; Prabhakara *et al* 1974). Another important consequence of this assumption is that the second term on the right side vanishes. The effect of absorption due to clouds has not been considered here which implies that our corrections are valid only for the measurements under cloudfree conditions. It may be added here that even under partially cloudy conditions it is possible to eliminate the cloud contaminated data for the retrieval of spectral radiances (Smith *et al* 1970). Attenuation and emission due to aerosols have been shown to be insignificant for this band in good visibility conditions (Deirmendjian 1960). The absorption due to aerosols is significant only for temperate and cold climates (Grassl 1973). In view of the above, aerosol effects have been neglected in the present study which means that the corrections are valid only for good visibility condition—a condition which was always satisfied during our experiment.

Water vapour is the principal absorbing agent in this spectral window. Water vapour continuum as well as lines contribute to the total absorption. Laboratory experiments of Bignell (1970) and Burch (1970) have shown that the most significant component of the water vapour absorption is the 'e' type absorption which depends upon water vapour partial pressure. Air-borne radiometric measurements by Platt (1972) and balloon borne measurements by Lee (1973) have confirmed the presence of 'e' type absorption in the atmosphere. Coffey (1977) has concluded that water vapour dimers are responsible for this absorption. The other components of water vapour absorption 'p' and 'l' type depend upon total pressure and line intensities respectively. CO₂ is another absorbing agent but its contribution to the absorption in this band is only very little. Both these effects have been included in the present study. The atmosphere absorbs the incoming terrestrial radiation and then the different atmospheric layers emit the radiation at their own temperatures. There is enhancement of the spectral radiance received at the detector because of re-emission from atmospheric layers. This effect has also been included here.

In the present experiment over Sandheads, no radiosonde data was available and, therefore, we decided to use the Calcutta radiosonde data on the day of our experiment. As Sandheads is only about 150 km south of Calcutta and, on the day of our experiment the atmosphere was very clear, it would be fairly reasonable to assume that atmospheric parameters over sandheads would be grossly similar to those over Calcutta. Table 1 gives the Calcutta radiosonde data which we have used.

As mentioned earlier we have to determine $\tau_{o,h}(\nu, T)$ and once we know that we can find out $\partial\tau/\partial z$ very easily. Since we are considering the absorption due to water vapour and CO₂ only, the total transmittance $\tau_{o,h}(\nu, T)$ would be given by

$$\tau_{o,h}(\nu, T) = \tau_{o,h}(\nu, T)_{H_2O} \times \tau_{o,h}(\nu, T)_{CO_2}.$$

Table 1. Calcutta radiosonde data.

Date : 5th October 1978

Time : 1730 hrs IST (12 GMT Flight)

Pressure (mb)	Height (gpm)	Dry bulb temperature (° C)	Dew point temperature (° C)
1003	0000	29	25
1000	0028	28	25
950	0450	26	22
900	0953	22	19
850	1447	19	16
834	1620	18	15
800	1967	17	14
750	2506	15	12
623	4100	06	-10
600	4369	04	-16

5. Water vapour and CO₂ transmittances

The water vapour transmittance $\tau_{0,h}(\nu, T)_{H_2O}$ was computed for every 10 cm⁻¹ interval in the wavenumber range 820–960 cm⁻¹ (10.4–12.2 μ m) using the following analytical expression given by Davis and Vieve (1964)

$$\tau_{0,h}(\nu, T)_{H_2O} = \exp [-(K_p \omega p)^{a_p}]$$

where K_p and a_p are the constants (refer to table 2), ω is the amount of precipitable water vapour in cm, and p is the ratio of *in situ* pressure to standard pressure.

This analytical expression was fitted by Davis and Vieve to the laboratory measurements of Roach and Goody (1958). Since we have considered 20 slabs upto an altitude of 3000 m, the experimental data given in table 1 have been interpolated to determine average temperature and humidity of all these slabs. The pressure at each slab was calculated by fitting an exponential. Another important parameter for these slabs is the amount of precipitable water vapour, ω , which has been calculated using the following formula:

$$\omega = \frac{1}{g} \int_{p_1}^{p_2} m(p) dp,$$

where m is the mixing ratio of water vapour, p is the pressure, and g is the acceleration due to gravity. $\tau_{0,h}(\nu, T)_{CO_2}$ has been computed by using the procedure adopted by Maul and Sidran (1973).

Once the transmittances are known one can determine the temperature correction for any given observed temperature T_h as mentioned earlier.

Table 2. Values of the constants K_p and a_p in the window region (Davis and Vieze 1964).

Wavenumber interval (cm^{-1})	H ₂ O window	
	K_p (Pr cm) ⁻¹	a_p
800-825+	0.170	0.775
825-850+	0.135	0.820
850-875+	0.115	0.860
875-900+	0.105	0.880
900-925+	0.095	0.885
925-950+	0.091	0.885
950-975+	0.091	0.885
975-1000	0.091	0.885
1000-1025	0.091	0.885
1025-1050	0.091	0.885
1050-1075	0.091	0.885
1075-1100	0.091	0.880
1100-1125	0.095	0.860
1125-1150	0.105	0.830
1150-1175	0.115	0.795
1175-1200	0.125	0.760

+ Denotes wavenumber intervals considered in the present work.

6. Results and discussion

A part of the radiation which leaves the sea surface is absorbed by the atmosphere and the remaining part, which does not interact with the atmosphere, reaches the radiometer. The values of the amount of precipitable water vapour, ω , and the net transmissivity of the atmosphere τ , upto different altitudes calculated for 5 October 1978 are given in table 3. The radiation which is absorbed by the atmosphere is re-emitted by different atmospheric layers. The energy received by an air-borne radiometer is made up of two parts—one emitted by the sea surface (ES) and the other which is emitted by the atmosphere (EA). Under the conditions of our experiment out of the energy received by a radiometer at 600 m, 84% is due to ES and 16% due to EA. At 3000 m, on the other hand, ES contributes only 62% and the remaining 38% is contributed by the EA.

Table 4 gives the observed temperatures (T_h), at different altitudes, which have been measured experimentally during the N-S and E-W motion of the aircraft, the temperature correction ΔT_h which has been evaluated using our scheme, and also the estimated value of the SST. The SST values, when calculated from observed temperatures at different altitudes range from 28.0° to 28.5° C during the N-S leg and from 28.2° to 29.0° C during the E-W leg.

The ship measurements of SST, arranged by us simultaneously with the aircraft measurements, were made using the simple bucket method with an accuracy of $\pm 0.5^\circ \text{C}$. The average sea surface temperature recorded by the ship was

Table 3. Precipitable water vapour and atmospheric transmittance at different levels for 5 October 1978 flight.

Altitude ' h ' (m)	Length of precipitable water vapour ' ω (cm)'	Net transmittance
600	1.33	0.82
900	1.87	0.77
1200	2.36	0.72
1500	2.79	0.69
1800	3.18	0.66
2100	3.55	0.63
2400	3.89	0.61
2700	4.20	0.59
3000	4.47	0.57

Table 4. Observed temperature (T_h), estimated temperature correction ΔT_h and estimated sea surface temperature (SST).

Altitude h (m)	N-S leg			E-W Leg		
	T_h °C	ΔT_h °C	SST °C	T_h °C	ΔT_h °C	SST °C
600	27.9	0.3	28.2	27.9	0.3	28.2
900	27.9	0.6	28.5	..	..	..
1200	27.2	0.8	28.0	27.5	0.9	28.4
1500	26.9	1.2	28.1	..	..	..
1800	26.6	1.6	28.2	26.6	1.6	28.2
2100	26.3	2.0	28.3	..	..	..
2400	26.0	2.4	28.4	26.3	2.6	28.9
2700	25.7	2.8	28.5	..	..	..
3000	25.3	3.1	28.4	25.7	3.3	29.0

29.2° C at the time of aerial flight. If one looks at table 4 which gives the estimated SSTs with an accuracy of $\pm 0.7^\circ$ and the ship measurement which has an accuracy of $\pm 0.5^\circ$ C, the agreement between the two appears to be very well within the experimental errors. In our opinion, this may be termed as very good agreement—a better agreement is not really called for due to following reasons. The emission of radiation takes place from a microlayer of the sea surface which is less than 20 μ m thick. As the microlayer is subjected to a number of air-sea interactions, such as flux of sensible heat and evaporation, etc., it is not uncommon that the temperature of the microlayer is -1.5° C to $+0.5^\circ$ C different from the temperature of water 10 cm below the surface. Also the

neglect of reflection by assuming unity emissivity would cause an error $\Delta T = (T_a - T_s)\rho\tau$ where T_a is the radiation temperature of the atmosphere, T_s is the bulk temperature of the sea, τ is the reflectivity and ρ is the atmospheric transmissivity. Changes in surface temperature ΔT range between -0.2°C to -0.7°C with an average of -0.5°C (Maul, private communication). Thus due to the neglect of reflection the radiometer-derived temperature would be lower than the actual microlayer temperature. Additionally, the temperature of the microlayer itself is lower than the bulk water below it due to air-sea interactions. Thus as observed by us the radiometer-derived temperatures should be lower than the bulk water temperature, which is measured by simple bucket method.

The present exercise shows that if the meteorological parameters in the region of interest are known to a reliable degree, the temperature corrections estimated by our scheme can be used to obtain the actual SST value. We propose to retrieve SST from our 1977 flight data, after knowing the true atmospheric conditions, preferably by making use of some observed data.

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