

Remote sensing of sea surface temperature through infrared radiometry – a review

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MS received 14 March 1986; revised 20 August 1986

Abstract. A review on the estimation of sea surface temperature (SST) from space using infrared radiometry is presented. The principle of the remote sensing technique, the infrared radiometers and the theory of SST measurement have been explained briefly. The absorption of infrared radiation from earth by the atmospheric constituents has been a major problem in the retrieval of SST from space-borne sensors. The effect of clouds and other atmospheric constituents, especially moisture content/water vapour, has been discussed and the various techniques used for evaluating the atmospheric correction and their limitations are summarised. Development of separate algorithms for each oceanic area coupled with validation by realistic sea truth measurements has been suggested for improving the accuracy of SST measurements from space.

Keywords. Remote sensing; SST; infrared; clouds; water vapour content; field of view.

1. Introduction

Remote sensing is the means by which information about an object is obtained without any physical contact between the observer and the object. The interaction between electro-magnetic radiation and any object or medium is dependent on the physical and chemical structure of the object or medium. Each object, therefore, produces a different spectral response. The basic aim of remote sensing is to detect these responses and identify certain characteristics of the object based on the differences in the spectral response.

The techniques involved in remote sensing may be regarded as either passive or active: passive sensors detect energy emitted, scattered or reflected from an object or medium (e.g. thermal or photographic sensing) while active sensors have, in addition, their own illuminating source of electromagnetic radiation (e.g. radar).

All matter radiates energy in the electromagnetic spectrum with the peak intensity shifting progressively towards shorter wavelengths with increasing

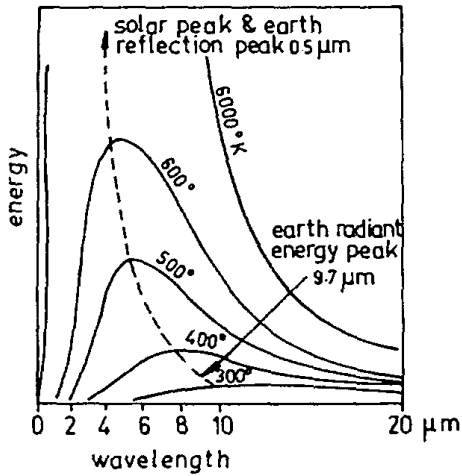


Figure 1. Spectral distribution curves of energy radiated from objects at different temperatures (from Colwell *et al* 1963, figure 2).

temperature (Wein's displacement law). Figure 1 shows the spectral distribution of energy emitted by a black-body at different temperatures. At the average surface temperature of the earth (300 K), the energy radiated is distributed as a broad curve with a maximum intensity at around $10 \mu\text{m}$. The radiation emitted by any body other than a blackbody depends on the emissivity of the medium as well. The emissivity of the medium is generally known, so that the temperature of the source can be determined from an accurate measurement of emitted radiation.

The atmospheric constituents selectively scatter and absorb electromagnetic radiation. The regions of high energy transmittance are the atmospheric transmission bands or windows. The transmission bands and the absorption bands specifying the gases responsible for the absorption are shown in figure 2. For effective remote sensing of the earth from space, it is necessary to use these transmission windows in order to minimise distortion of the terrestrial signal by the intervening atmosphere.

This paper deals with measurement of sea surface temperature (SST) through remote sensing. To measure SST, infrared (IR) and microwave regions of the electromagnetic (e.m.) spectrum have proved to be of great use. Earlier reviews by Gupta (1978a, b), Gupta & Rao (1983) and Brown & Cheney (1983) may also be

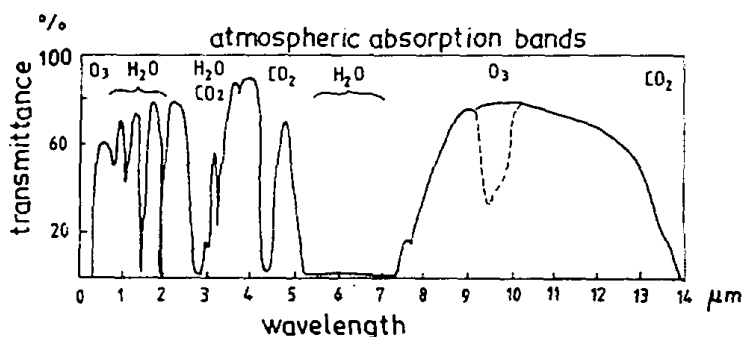


Figure 2. Transmission of energy through the atmosphere as a function of wavelength (from Santa Barbara Research Centre 1975).

referred to. In the present study, a comprehensive review on IR techniques for measuring SST is attempted.

The wavelength region in the electromagnetic spectrum ranging from 0.7 to 300 μm is the IR region. Of this 0.7 to 3 μm is the reflected-IR, 3 to 5 μm may be called mid-IR and 8 to 14 μm could be called the thermal-IR. The thermal-IR is important for the measurement of SST because the terrestrial radiation emitted from the ocean/land and received by a space-borne IR radiometer consists principally of thermal radiation (figures 1 & 2).

Remote measurement of SST is generally made in atmospheric windows, where the transmissivity is high. Due to the high absorption by ozone around 9 μm , this wavelength band is generally not used for remote sensing (figure 2). Because of their high transmissivity and low absorption by various atmospheric constituents, wavelengths around 3.5 and 10 to 12 μm are generally used. These wavelengths have got their own advantages and disadvantages and wavelengths used in combination are found to give better results. As mentioned earlier, the intensity of the radiated energy is maximum around 10 μm and since the reflected component of solar radiation is negligible, the measurement at 10–12 μm can be made both by day and by night.

Many authors emphasised the use of 3.5 μm wavelength for night measurement of SST. The data at this wavelength have been in use since 1968 (La-Violette & Chabbot 1968) to retrieve SST. Chahine (1981, p. 87) described it as a 'superwindow' which is far away from the major water vapour lines. Takashima & Takayama (1981) calculated the optical thickness of the atmosphere in order to estimate SST at 3.5 μm range considering the various atmospheric constituents. The radiation at 3.7 μm is less affected by the atmospheric constituents than in the 11 and 12 μm region (Weinreb & Hill 1980). Its absorption is mainly due to CO_2 . Even though it has a lower transmission in the polar atmosphere, its variability is less. Its higher transmissivity in the tropical atmospheres renders it useful for the measurement of SST over tropical oceans. Radiometric sensitivity at this wavelength is also found to be high (Llewellyn-Jones *et al* 1984). The non-linear response of the detectors used at 3.7 μm is minimum compared to that at 11 and 12 μm (Singh 1984). Measurements at this wavelength are more useful for cloud cover identification (Arking 1982, p. 42). However, the radiance measured at this wavelength band in the daylight hours may be based on nearly equal amounts of reflected and radiated IR energy. Since only radiated energy is of interest here, reflected IR is easily excluded by night operation. However, the 11–12 μm band is preferred for daytime operations.

Clouds play a major role in the radiative processes in the atmosphere both in the absorption and reflection of solar radiation and in the emission of thermal energy. Clouds absorb and scatter radiation of wavelengths less than 0.3 cm. This limits the use of thermal IR radiation in cloudy areas. Only microwave and longer wavelengths are capable of penetrating clouds without scattering, reflection or absorption.

2. IR radiometers

An IR radiometer is a device to measure radiant temperature quantitatively. To measure the temperature of a target the sensor views earth and then space and a

body of known temperature to provide two-point calibration data. The technique was first proposed by Stommel *et al* (1953). Basically, an IR radiometer consists of three main units: a thermistor bolometer detector, an inline reference cavity, and signal processing electronics. Such radiometers have been discussed in detail by Richardson & Wilkins (1958), Clark & Frank (1963), Colacino *et al* (1970), Sabins (1978, p. 128), Sinha *et al* (1980), and Robinson & Dewitt (1983, p. 293).

Since 1970, spaceborne radiometers have been in use to measure SST. A very high resolution radiometer (VHRR) was flown aboard the TIROS series of operational satellites. Indian Satellite INSAT-1B too carries a VHRR with one channel in the visible and one in the infrared. Recent satellites (TIROS-N launched in October 1979 and NOAA-6 launched in June 1979 and further follow-on satellites) carry improved versions of VHRR, namely, advanced very high resolution radiometer (AVHRR). Both VHRR and AVHRR are cross-track scanning systems, which measure upwelling radiation at IR wavelengths (NOAA 1981). VHRR had only two channels (one in the 11–13 μm wavelength range) compared to 4 channels (one each in the 3.5–3.9 μm and 10.5–11.5 μm wavelength ranges) of AVHRR. NOAA-7 launched in June 1981 carries a further improved version, AVHRR/2 (5 channels). The channels are chosen in such a way that the atmospheric effects are different for each channel. Additional information on AVHRR is available in Lauritson *et al* (1979) and Huh & Di Rosa (1981). An along track scanning radiometer (ATSR-M) has been proposed for ERS-1 (European Remote Sensing Satellite), which is scheduled to be launched in 1989. ATSR-M has the same IR channels as AVHRR/2. The measurement accuracy will be much more with ATSR-M because of the additional microwave channels at 23.8 and 36.5 GHz to determine the total water vapour content useful for atmospheric correction (ERS-1 programme proposal 1984). ATSR-M views the earth at 2 angles along the track. The advantage is that a given pixel is viewed twice along different atmospheric columns, and this helps improve atmospheric corrections.

3. Theory of measuring SST using infrared remote sensing

A detailed discussion on the theory of infrared SST measurement has been given by Warnecke *et al* (1971), Aoki (1979), Chahine (1980, p. 411), Agarwal (1983) and Chahine *et al* (1983, p. 165). However, to retain continuity for readers, a brief explanation is given here.

For a radiating surface, the emitted radiation detected at satellite altitude through a filter system can be described by the radiative transfer equation (Warnecke *et al* 1971).

$$N = \int_{\lambda_1}^{\lambda_2} \Phi(\lambda) \tau_s(\lambda) \varepsilon_s(\lambda) B(\lambda, T_s) d\lambda + \int_{\lambda_1}^{\lambda_2} \int_{h=0}^{h^*} \Phi(\lambda) \frac{\partial \tau(\lambda, h)}{\partial h} B[\lambda, T(h)] dh d\lambda, \quad (1)$$

where, N is the detected radiation; $\Phi(\lambda)$ is the filter function ($\Phi(\lambda) = 0$) outside the interval (λ_1, λ_2) ; $\tau_s(\lambda)$ is the transmissivity between the radiating surface and

satellite; $\tau(\lambda, h)$ is the transmissivity from a height h to the satellite at height h_s ; $\epsilon_s(\lambda)$ is the emissivity of the radiating surface; $B(\lambda, T)$ is the Planck function; and T_s is the temperature of the radiating surface.

The first part of the right hand side of the above equation is the contribution from land/ocean, thick clouds etc. and the second part is from atmospheric constituents like thin clouds, CO₂, ozone, dust etc. If the atmospheric window is absolutely clear where $\tau = \tau_s = 1$ then (1) becomes

$$N_c = \int_{\lambda_1}^{\lambda_2} \Phi(\lambda) \epsilon_s(\lambda) B(\lambda, T_s) d\lambda. \quad (2)$$

With the knowledge of $\Phi(\lambda)$ and emissivity $\epsilon_s(\lambda)$, (2) may be solved for $B(\lambda, T_s)$ from which T_s can be determined by inverting the Planck function (Lowan & Blanch 1940). The emissivity in the above equation at thermal-IR wavelength is nearly unity (Buettner & Kern 1965).

Practically no atmospheric window is ideal, that is, where the transmissivity is equal to unity. Therefore the absorption and emission of IR radiation by the predominant constituents like water vapour, CO₂, ozone, and aerosols, must be taken into account to retrieve SST from spaceborne sensors. The detected radiance (N) varies with the gas distribution $\partial\tau/\partial h$ and the gas temperature $T(h)$ along the optical path with wavelength λ . It also varies with the distribution of individual gases. Water vapour is the most complex of all and its distribution is highly variable in both space and time. To obtain SST using spaceborne sensors these factors are to be considered and the transmissivity has to be calculated accordingly. Various models are available to calculate transmissivity. Davis & Viezee (1964) developed a model which computes transmissivity of the atmosphere where water vapour and CO₂ are present. Considering the effects of water vapour, ozone, CO₂, methane and nitrous oxide, Anding & Kauth (1969) calculated the spectral radiance for 5 to 30 μm wavelength. Similar transmittance models have been developed by Wyatt *et al* (1964), Selby *et al* (1976), Stull *et al* (1964), and Idso (1981).

Further, if the measurement is made during daytime in the mid-IR, reflected sunlight is also added to the received radiance (Maul 1980) and hence the reflected solar radiation term, especially near 3.5 μm range must be added to (1).

4. Atmospheric correction

Since 1968, the remote measurement of SST (La-Violette & Charot 1968) has reached a stage where it can monitor thermal changes (Igeckis 1975; Tabata 1977; Simpson *et al* 1978; Roden 1980; Emery & Mysak 1980) to track eddies and currents (Strong & De Ryke 1973, p. 1069; Stumpf & Rao 1975; Harris *et al* 1978; Lee *et al* 1981; Christensen *et al* 1981) and upwelling (La-Violette 1974; Stumpf 1975; Dickson *et al* 1980; Strong *et al* 1972). But the correction due to clouds and other atmospheric constituents to retrieve the SST from spaceborne sensors still remains a major problem.

The measurement of global SST using satellites with greater accuracy is one of the goals of the World Climate Research Programme and a target of ± 0.2 K accuracy for the large scale processes (200–300 km) and ± 1 K for mesoscale processes

(5–10 km) has been quoted by Harries *et al* (1983). To achieve this goal using the infrared technique, the absorption, emission, reflection, and/or scattering due to atmospheric constituents and clouds must be considered. This section reviews the effect of cloud cover and atmospheric constituents.

4.1 Effect of cloud cover

Measurement of SST from space at IR wavelengths is hampered by the presence of clouds, since thick clouds are opaque to insolation and radiate at their own temperature like a black body, thereby effectively masking oceanic radiation at satellite altitude. It is therefore not possible to retrieve SST by IR remote sensing over cloud covered areas. In this section we discuss the methods for identification of cloud covered areas in a satellite image, and the possible errors in SST introduced when cloud identification is not achieved correctly.

The simplest method of cloud identification is the application of a threshold technique to each field of view (FOV). Cloud covered pixels are separated from cloud free fields on the basis of their radiances. Cloud covered fields generally have less radiance value compared to the radiance from the sea surface. Based on this, a "high daily average" (HDA) method was proposed by La-Violette & Chabot (1969) to eliminate the cloud covered FOV from the NIMBUS-II HRIR (High Resolution Infrared Radiometer) data. The drawback of this technique is that unless the FOV is filled with cloud this may introduce errors (Shenk & Salomonson 1972a). Another difficulty is that, if the water is as cold as the cloud tops (e.g. low clouds) or below the threshold value, the pixels will be classed along with the cloud pixels. To avoid this, simultaneous evaluation of clouds in the visible and IR wavelengths will be useful. Cloud tops have much higher brightness value in the visible region than the ocean surface (Bernstein 1982; Shenk & Salomonson 1972b; Fleming 1980, p. 503; Smith *et al* 1970). With simultaneous measurements utilising the visible and IR wavelengths, it is possible to extract the optical thickness of the clouds and by inverting the radiance measurements the cloud class can be identified. The one-dimensional histogram approach of Smith *et al* (1970) involves the search for peaks of high radiating temperatures. The high temperature side of these peaks is then used to estimate the clear sky radiances.

The effects of sensor spatial resolution on estimating the amount of cloud cover has been described by Shenk & Salomonson (1971, 1972b). They simulated the effect of various cloud distributions and sizes on the measurement of the cloud amount using the threshold technique. If the ratio of the aerial cloud size to the aerial resolution of the sensor (R) is known then the cloud cover can be determined with reasonable accuracy. For a cloud cover between 30 and 70% and if the ratio (R) is greater than 25 the estimated cloud cover becomes close to the true percentage. But when the ratio (R) becomes less, the estimated cloud cover is larger than the actual cloud cover. To improve the accuracy a nomogram has also been generated from simulated cloud patterns showing estimates of cloud cover.

Rao (1970) has shown that cloud amount can be determined using the visible channel and cloud height from the IR channel, when the FOV is filled with clouds, based on the cloud top temperatures. Maul & Sidran (1973) compared known values of SST with satellite derived temperatures in cloud contaminated areas. They found that when the cloud cover is 10% in a FOV, an error of 4 K is introduced into

the estimated SST when cloud altitude is 10 km, whereas the error is about 0.5 K when the altitude is 1 km.

Cogan & Willand (1976) separated the cloud contaminated pixels by comparison with visible data and tested the estimated temperatures over cloud free areas (T_B), to check whether they lie within the climatological limits of SST ($T_{\min} < T_B < T_{\max}$, where T_{\min} and T_{\max} are the climatological minimum and maximum temperatures, respectively). Pixels with T_B values outside the climatological limits are rejected.

To identify cloud covered pixels, Saunders (1982, p. 282) and Arking (1982, p. 42) emphasised the usefulness of 3.7 μm wavelength. During daylight hours both terrestrial emitted and solar scattered and reflected radiances contribute at this wavelength. Thus high cirrus clouds will have a low radiance due to its low emission and low reflectivity, whereas a low cloud has a high radiance value (higher emission and high reflectivity). The result of this is that the cloud free sea surface radiance has a value inbetween that of a low and high cloud due to the high emission and low reflectivity of the sea surface at 3.7 μm . For this method, both upper and lower thresholds have to be defined and it fails when the pixel is filled with partially low and high clouds.

Multichannel methods using visible and near IR region of the e.m. spectrum is possible only in the daytime, but they fail in regions of sunglint. Crosby & Glasser (1978), Crosby & De Priest (1980, p. 485) and Harris *et al* (1981, p. 199) described a method to remove the cloud contaminated pixels both by day and by night. The procedure involves estimating the SST from a large number of radiometer measurements using the theory of Truncated Normal Distribution (TND) (Johnson & Kotz 1970, p. 81). The principle is to fit a Gaussian curve to the warm edge of a histogram of the brightness temperatures and identify a cloud free value from the peak value of the fitted curve assuming that the SST is uniform and warmer than clouds. All cloud contaminated values cannot be removed using this technique and the result is found to be biased towards the warmest values when regions of SST variability are considered. To eliminate this residual cloud contamination from the TND technique, Fleming (1980, p. 503) modified the TND derived SST by an Error Detection and Correction (EDC) technique. The principle of the EDC technique is based on the fact that higher order finite differences approach zero for accurate, equally spaced data that fit a smooth polynomial, while on the other hand, any initial error in the data gets magnified at higher differences.

Another method of measuring clear and cloudy sky radiances is the spatial coherence method (Coakley & Bretherton 1982) which utilises the local spatial structure of the IR radiance in cloud free and completely covered FOV. Over cloud-free ocean the local standard deviation in brightness temperature will be small, whereas it is larger for cloud contaminated FOV. It relies on the assumption that over small horizontal distances, the radiance from the sea surface will be nearly constant, whereas from cloud tops it will vary appreciably. The advantage of this method over the threshold technique is that the FOV need not be filled with cloud cover. This method also fails for detecting multilayered and variable emissivity clouds as do most of the other techniques. Uniform low clouds may also have low local standard deviation, where the spatial coherence method fails. Llewellyn-Jones *et al* (1984) used this spatial coherence method to identify the cloud covered pixels followed by a visible threshold value (SCV method). A local standard

deviation threshold value is also determined by changing its value for some test cases.

4.2 Effect of atmospheric constituents

Sea surface temperature obtained from brightness temperature measured from space in the atmospheric window over a cloud-free ocean is, in general, lower than the actual SST. The absorption of IR radiation by various atmospheric constituents lowers the satellite derived SST ranging from a few tenths of a degree for arctic atmospheres to fairly high values ($\sim 10^\circ\text{K}$) for tropical atmospheres (Farmer & Houghton 1966; Shapiro & Gush 1966; Maul & Sidran 1973; White *et al* 1978). In fact, the main gases of the atmosphere (nitrogen and oxygen) contribute only slightly to the radiation absorption, whereas the variable atmospheric constituents such as water vapour, carbon dioxide, ozone, aerosols, nitrogen oxides and hydrocarbon combinations have absorption bands and lines in the IR spectrum. Figure 2 shows the absorption bands of atmospheric gases in the wavelength region 0 to 14 μm .

The absorption bands of nitrogen oxides (NO , N_2O , N_2O_4 , N_2O_5) and the hydrocarbon combinations (C_3H_8 , C_2H_6 , C_2H_4 , CH_4) are rather weak or very narrow in the IR spectrum. Since the energetical aspect of their absorption or emission is of no practical concern their effects can be neglected while measuring SST from space (Kondratyev 1969, p. 99). The effect of aerosols is very small and may cause only a slight error in the retrieved SST (Grassl 1974; Brower *et al* 1976; Stowe & Fleming 1980). However, aerosol absorption coefficient can become very large for hazy conditions with relative humidities approaching 100%, where they alter the retrieved SST to a considerable extent (Hanel 1976; McClatchey *et al* 1972). Water vapour, CO_2 and ozone contribute principally to the opacity in the atmospheric window. As shown in figure 2, the effect of ozone may be reduced by avoiding measurements at around 9.5 μm . The effect of CO_2 is generally much smaller than that of water vapour and since CO_2 is uniformly mixed and has a relatively constant distribution globally, it can be accounted for in retrieving the true surface temperature. The variability of water vapour both in time and space complicates any attempt to determine the sea surface temperature.

The total water vapour transmission in the window region can be expressed as the product of τ_c , τ_p and τ_l (Prabhakara *et al* 1974; Bignell *et al* 1963) where τ_c and τ_p refer to the water vapour continuum and τ_l refers to the water vapour lines. Continuum absorption is the absorption in the wings of strong water vapour lines, which peak far outside the window, and is associated with collision broadening of the distant lines, and τ_l is associated with absorption by lines which peak within the window and is strongly dependent on wavelengths (Sidran 1980). Computations of τ_c , τ_p and τ_l are given by Prabhakara *et al* (1974).

The water vapour absorption in the window region is dependent on the water vapour partial pressure, e . Bignell (1970) named it the e -type absorption. Water dimer molecules – two water molecules joined by a hydrogen bond – are responsible for this e -type absorption (Chahine *et al* 1983, p. 165). Many authors have reported this e -type absorption (Platt 1972; Grassl 1973; Coffey 1977). Houghton & Lee (1972), Prabhakara *et al* (1974) and Watkins *et al* (1979) emphasised the importance of e -type absorption in determining SST.

Differential absorption of water vapour at two wavelengths is found to be useful for estimating the magnitude of water vapour absorption and emission. If the absorption at the two wavelengths is different, then the temperature inferred from one wavelength will be different from the other, e.g., the absorption at 10.3–11.3 μm is greater than that near 11.5–12.5 μm . Therefore the temperature inferred from 10.3–11.3 μm will be less than that at 11.5–12.5 μm . Since the physical processes at these wavelengths are the same for absorption and emission, the radiance value at 10.3–11.3 μm must be a magnification of the effect near 11.5–12.5 μm . Thus a simultaneous measurement near these wavelengths will provide an estimate of water vapour content in the atmosphere.

The concept of differential absorption of water vapour at different wavelengths was first proposed by Anding & Kauth (1970). McMillin (1971) developed a physical basis for the empirical results obtained by Anding & Kauth (1970). Prabhakara *et al* (1974) used this technique successfully to retrieve SST from the Nimbus 3 & 4 IRIS (infrared Interferometer Spectrometer) data at wavelengths 10.5 and 13 μm . Sidran (1980) empirically modified the transmittance model of Prabhakara *et al* (1974).

The results of Morcrette & Irbe (1978), McClain (1981), and Deschamps & Phulpin (1980), show that this technique can also be applied to channels centred at 3.5 and 11 or 12 μm . Watkins *et al* (1979) and Barton (1981) used a similar approach to determine the water vapour absorption at 3.5–4.2 μm wavelength. At these wavelengths the *e*-type absorption is considerable and the temperature dependence of the Planck function gives better temperature discrimination than at longer wavelengths. The use of differential absorption at different viewing angles was suggested by Saunders (1967), McMillin (1975), and Chedin *et al* (1982).

Prabhakara *et al* (1974) found that the water vapour transmission function varies linearly with path length of the water vapour. Becker (1982, p. 151) presented a semi-empirical method to correct for atmospheric attenuation based on the principle of differential absorption considering the atmospheric constituents CO_2 , N_2O , CO , CH_4 , which are relatively constant in the atmosphere, and variable constituents like water vapour and ozone. McClain (1981, p. 73) used seasonally and geographically different sets of 59 cloud-free maritime radiosonde observations to calculate a transmittance model at 3.8, 11 & 12 μm brightness temperatures. The results were verified against buoy and XBT temperatures and an rms error of 1.1 K was obtained.

Grassl & Koepke (1981, p. 97) found that the correction is minimum at 11.1 μm below 30° nadir angle of observation. The correction is large with increase of angle of observation. This is due to the decreasing emissivity of the sea surface as well as increasing airmass. They have shown that accuracy of 0.5 K & 0.3 K is possible from one spectral channel (centred at 11.1 μm) and from two spectral channels (centred at 11.1 & 12.2 μm) respectively. While using a single channel, additional information on all absorbers are to be obtained from numerical weather forecasts. Measurements errors are neglected in their study.

Dalu *et al* (1981, p. 109) proposed a simulative study to measure SST with an accuracy of 0.3 K with the knowledge of total precipitable water using a channel at 18 μm wavelength. Instrument errors which get amplified when using multiple channels (Deschamps & Phulpin 1980) were not considered in their study; an error of 0.2 K is likely to cause an uncertainty of 1.2 K in the retrieved SST (Grassl &

Koepke 1981, p. 97). Prabhakara *et al* (1979) noted that the total water vapour can be determined from the water vapour line strength at $9 \mu\text{m}$.

Anding & Kauth (1970) demonstrated the selection of wavelengths where the difference in absorption is maximum and found that 10.96 & $9.19 \mu\text{m}$ give better results. Maul & Sidran (1972) obtained 10.96 and $8.60 \mu\text{m}$ with another model. Anding & Kauth (1972) modified their earlier model, taking into consideration *e*-type absorption, and obtained new wavelengths, 11.9 and $8.95 \mu\text{m}$. Deschamps & Phulpin (1980) with the use of $3.7 \mu\text{m}$ wavelength, found that the combination of 3.7 and $12 \mu\text{m}$ wavelengths gives better estimates of water vapour content. They also found that the temperature deficit, ΔT (difference in temperature between true SST and satellite derived SST), produced by the atmospheric emission is proportional to the absorption coefficient and also depends on air temperature. Their comparative study shows that absolute temperatures of better accuracy (0.14 to 0.39 K) can be expected from multiple window methods, if the noise equivalent difference temperature (NE ΔT) of the instrument is sufficiently low (0.1 to 0.2 K). Imbault *et al* (1981) studied the absorption at two wavelengths due to the line spectrum and continuum absorption of different absorbers in two pairs of channels; 8.7 and $11.6 \mu\text{m}$ on the one hand and 11 and $12 \mu\text{m}$ on the other. Their experimental study with an airborne radiometer showed errors varying from 0.3 to 0.8°C between the experimental and theoretical values.

McConaughy (1980) derived an empirical regression equation for the estimation of SST in the wavelength region 10.5 – $12.5 \mu\text{m}$ using 15 sea truth values and obtained a correlation coefficient of 0.93 . A regression equation was also proposed by Bernstein (1982) between sea truth values of SST and brightness temperatures at 3.7 and $11 \mu\text{m}$ for the cloud-free daytime observations. They obtained an rms deviation of 0.5 – 0.6°C .

To map the global SST, McClain *et al* (1983) described a multichannel SST (MCSST) method from the AVHRR on NOAA polar orbiting satellites. Various threshold and spatial homogeneity tests were applied to choose cloud-free samples. The brightness temperatures at the window channels are corrected for atmospheric attenuation with the brightness temperatures from either two or all three channels (3.7 , 11 and $12 \mu\text{m}$). The relationships between atmospheric correction and the combinations of brightness temperature difference for various channels were found to be linear with a small scatter and were used to develop MCSST equations both for day and night. Weekly MCSST charts are available on a routine basis from NOAA (e.g. figures 3 and 4). A comparative study by Strong & McClain (1984) showed that rms difference observed with MCSST/drifting buoys match-ups is 0.4°C less than with the MCSST/fixed buoy match-ups and 1.1°C less than with MCSST/ship match-ups.

A numerical simulation was made by Llewellyn-Jones *et al* (1984) to obtain SST from the AVHRR/2 for the north-eastern and tropical Atlantic area. Their study emphasised the dependence of airmass on determining SST from space. Errors upto 2 K may occur in the retrieved SST if airmass is neglected. For a NE ΔT , 0.12 of AVHRR/2, their simulation gives an rms error of 0.3 K and 0.2 K for the North Atlantic and 0.8 K and 0.3 K for the tropics using two (split window at 11 and $12 \mu\text{m}$) and three (triple window at 3.7 , 11 and $12 \mu\text{m}$) channels, respectively.

With a prior knowledge of SST, Aoki *et al* (1982) theoretically estimated the accuracy of retrieved SST from satellites on the basis of modern optimal estimation

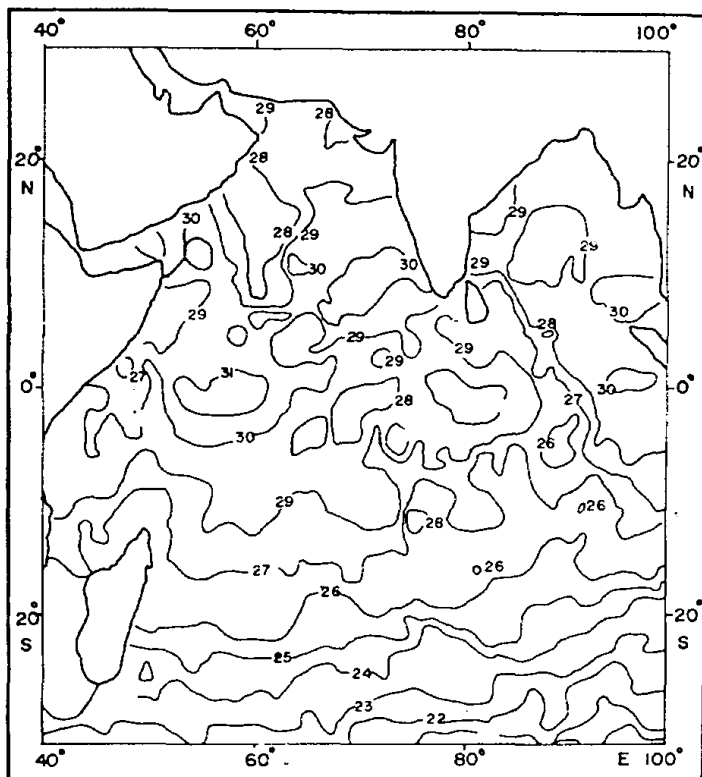


Figure 3. Satellite derived SST distribution over Indian Ocean 20 May 1986 (Source: NOAA-MCSST chart).

theory. They plotted the rms error versus the number of channels for two model atmospheres and found that the accuracy of measurement of SST increases with the increase of channels. It is also noted that in the tropics even the multichannel method does not increase the accuracy satisfactorily. The retrieved SST when compared with buoy and ship reports gave an rms deviation of about 0.8 K.

A comparison of SST obtained from satellites NOAA 3, 4, 5, 6 and TIROS-N showed that only NOAA 3 and 6 gave reasonable estimates of SST (Tabata 1981, p. 145). Pathak (1982) showed that about 60% of the ΔT values of TIROS-N lies within 1.5°C while the remaining is in the 2 to 4°C for the Indian Ocean. Legeckis & Bane (1983) reported a standard error of 0.5°C between the SST obtained from TIROS-N and AXBT, for a temperature range of 19–26°C for the Gulf Stream area. Legeckis *et al* (1980) compared the SST measured by polar orbiting VHRR (NOAA-5) and geostationary VISSR (GOES) and found that the VISSR (Visible Infrared Spin Scan Radiometer) derived SST is 1–2°C lower than that of VHRR. This error may be due to the calibration errors of VISSR (Bristor 1975). Legeckis *et al* (1980) also noted that the accuracy of VHRR derived SST is within 1°C for an SST range of 6–7°C and that the accuracy is reduced to 2–3°C in the 23–27°C range. In the tropics the difference is found to be greater still.

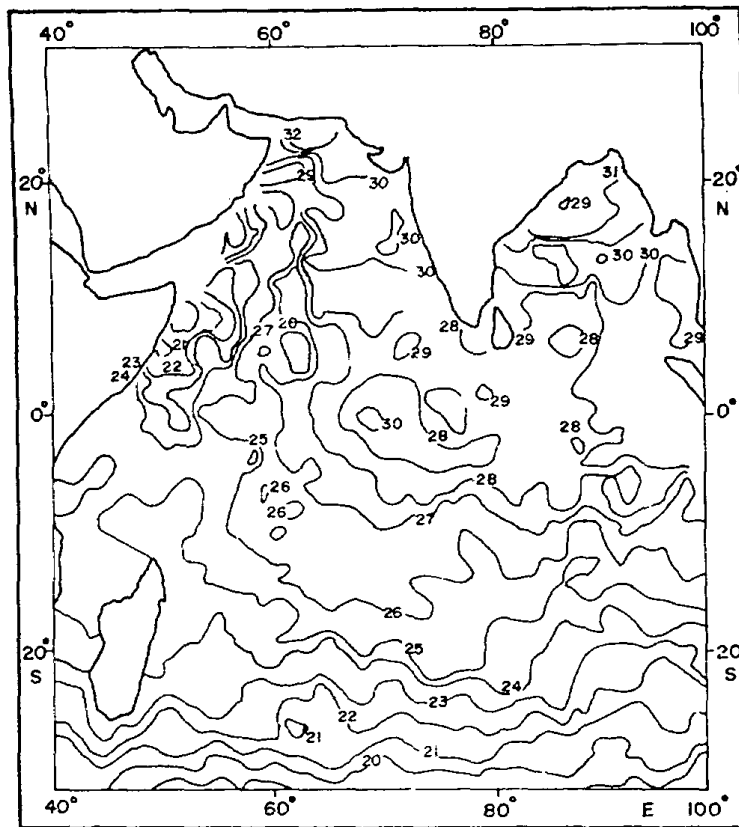


Figure 4. Satellite derived SST distribution over Indian Ocean 1 July 1986 (Source: NOAA-MCSST chart)

5. Discussion

About 98% of the incoming short wave radiation is absorbed in the top (≈ 0.1 mm) layer of the ocean (Ewing & McAlister 1960; McAlister & McLeish 1969). Therefore, the IR radiation emerges from the skin of the ocean surface. Thus the satellite measured SST is the skin temperature of the ocean instantaneously seen by the radiometer, whereas the conventional SST (bucket temperature) is the mean temperature of a well-mixed layer at least several decimeters deep. It has been shown that the skin temperature is colder than the lower turbulent layers by several decimals of a degree centigrade under laboratory conditions (Ramdas & Raman 1946; Spangenberg & Rowland 1960; Katsaros & Liu 1974) and in field observations (Woodcock 1941; Woodcock & Stommel 1947; Ball 1954; Ewing & McAlister 1960; Hasse 1963; Saunders 1967; Katsaros 1973, 1977; Simpson & Paulson 1980; Paulson & Simpson 1981). This temperature difference may be ascribed partly to the transfer of sensible heat to the air and partly to the effect of evaporation (Roll 1965, p. 227). The skin temperature is also controlled by local conditions such as wind speed, wave condition and cloud cover and hence is a variable, dependent on these conditions (Minnett *et al* 1984, p. 57). For climatic studies, algorithms involving SST could be based on either skin or bucket temperature. As the bucket temperature averages are more stable and reliable

their use has been preferred for studies under WCRP. This requires further work to relate the skin temperature to the bucket temperature.

Validity of SST retrieval algorithms are checked by comparison with sea truth measurements. Evidently this calls for accurate measurement of sea truth temperatures. Unfortunately, this requirement is not always satisfied. Various authors have pointed out discrepancies as high as 1°C between different *in situ* methods (Saur 1963; Collins *et al* 1975; Tabata 1978).

The temperature deficit (ΔT) described in the earlier section is also a result of the non-unity surface emissivity of the sea surface. The emissivity of sea surface at IR wavelengths varies from 0.900 to 0.993 depending on viewing angle and wavelength (Griggs 1968; McAlister 1964; Wolfe 1965; Katsaros 1980, p. 293). Cogan (1972) showed that the deviation of emissivity from a black body may cause a decrease of about 0.2 K in the retrieved SST. Singh (1984) estimated this emissivity correction as 0.45, 1.27 and 1.37 in the 3.7, 11 and 12 μm channels, respectively, for a surface temperature of about 280 K.

6. Some Indian studies

There is an increasing interest in India at present, both on the derivation of suitable algorithms for SST retrieval from satellite data under the humid tropical conditions that are relevant to the Indian Ocean, as well as in the use of satellite derived SST for oceanographic and meteorological applications. Scientists of the Space Applications Centre, Ahmedabad, have brought out a series of reports on algorithms and software for SST retrieval from infrared data (Garde 1978; Agarwal 1983; Murthy *et al* 1984). They recommend the use of some HIRS (High resolution Infrared Sounder) infrared channels in addition to the AVHRR channels. The HIRS forms part of the TIROS Operational Vertical Sounder (TOVS). Pathak (1982) has made a comparison of satellite derived and *in situ* SST measurements over the Indian Ocean. The VHRR data from INSAT-1B is being used by the India Meteorological Department (IMD) to derive SST.

The synoptic coverage, and quick availability of satellite derived SST makes it an attractive source of data for many applications. Garde *et al* (1980) studied the SST distribution around a tropical cyclone using the NOAA-5 VHRR data. Mishra (1981) has studied SST distribution patterns in the north Indian Ocean as indicator of changes in the oceanic circulation pattern during the SW monsoon period. Ramesh Babu *et al* (1985), Ramesh Kumar *et al* (1986) and Gopinathan & Sastry (1986) have related changes in SST pattern in the Indian Ocean, with monsoon rainfall over India. Venkateswara Rao *et al* (1986) have used satellite derived SST for yet another application – to remove bias due to SST changes in the evaporation rates estimated from satellite microwave data for the Arabian Sea.

7. Conclusions

Measurement of SST via remote sensing is important because the synoptic coverage of the ocean envisaged by this method will be very useful for several purposes, including studies on ocean-air interaction, heat budget, eddies, oceanic fronts,

climate research, fisheries industry and many military applications. Because of the importance of real-time SST data, considerable efforts have been and are being made to collect and analyse the satellite data.

Existing techniques for SST measurement from satellite seem to have a limit of about 0.6 K rms error. There are many difficulties that arise when comparing satellite observations with sea truth measurements. There is generally a difference in position and time of measurements. Sub-pixel cloud contamination of the IR observations under partly cloudy conditions also accounts for error. Further, studies have to be carried out to examine whether the errors are due to satellite instrumental noise, sea truth measurements, atmospheric or oceanic temperature variability. Realistic sea truth measurements are required for comparisons to determine the validity and optimum accuracy from this technique.

A sea truth campaign using many ships for different seasons would contribute to the development of a better model for each oceanic region. Separate algorithms for each oceanic area would contribute to increasing the accuracy of SST retrieval.

It is hoped that the forthcoming ATSR-M experiment on board ERS-1 would facilitate improved atmospheric corrections leading to a better accuracy in SST measurement from space.

The authors thank Drs V V R Varadachari, H N Siddiquie and J S Sastry for their encouragement and interest in this study. They also thank Shri M R Ramesh Kumar for critically going through the manuscript. The first author was supported by a CSIR fellowship.

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