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A simulation technique for the determination of atmospheric water content with Bhaskara satellite microwave radiometer (SAMIR)

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Abstract. A simulation technique has been developed to estimate the integrated atmospheric water content over oceans using the 19·35 and 22·235 GHz brightness temperature data from satellite microwave radiometer (SAMIR) on board the Indian satellite Bhaskara. The results obtained have been compared with those from linear statistical regression and empirical methods as well as from the nearest radiosonde observations. Based on the simulation method, a map of total precipitable water for some of the Bhaskara passes in July 1980 over Bay of Bengal is given. The possible applications of such maps in the study of Indian summar monsoon and boundary layer characteristics have been pointed out.

Keywords. Simulation technique; Bhaskara SAMIR; atmospheric water content; brightness temperature.

1. Introduction

The total water vapour and liquid water contents over oceans have been derived by Staelin *et al* (1976) and Grody (1976) using data from Nimbus E microwave spectrometer. Basharinov *et al* (1969) and Gurvich and Demin (1970) have derived these quantities using Cosmos-243 microwave radiometer data.

In the present paper, the authors have developed a simulation technique to derive total water content over oceans using data from the microwave radiometer (SAMIR) on board the Indian satellite Bhaskara.

SAMIR is a two-frequency radiometer, one at 19·35 GHz and the other at 22·235 GHz. The ground resolution at nadir is 125 km and 200 km circle respectively for 19 and 22 GHz frequencies. The temperature resolution of either channel is about 1° K. The instrument collects data at + 5·6, + 2·8, − 2·8, − 5·6° from nadir along the subsatellite track at 19·35 GHz and at + 11·2, + 2·8, − 2·8, − 11·6° at 22·235 GHz. The data at 2·8° look angle have been used in the present analysis. The details are described by Pandey *et al* (1980).

The effect of ocean surface roughness was taken into account using an empirical expression suggested by Hollinger *et al* (1975), wherever information on ocean surface wind was known; otherwise the wind effect was not included.

Possible applications of total precipitable water and the maps generated therefrom for synoptic analysis, boundary layer studies and computation of fluxes of total precipitable water in the study of Indian summer monsoon have been pointed out.

2. Mathematical developments

For a nonscattering atmosphere in local thermodynamic equilibrium, the radiative transfer equation describing the intensity of radiation emerging from the atmosphere can be written in terms of brightness temperature (equivalent black body) as

$$T_B(\nu) = \int_0^H T(z) \frac{\partial \tau_\nu(Z, H)}{\partial Z} dZ + \tau_\nu(0, H)\,[\epsilon_s(\nu)\,T_s + (1 - \epsilon_s(\nu))]$$

$$\times \int_H^0 T(z)\,\frac{\partial \tau_\nu(0, z)}{\partial z}\,dz. \tag{1}$$

The three terms on the right side of equation (1) represents (a) the upward emission by the atmosphere, (*b*) emission by the surface attenuated by the intervening atmosphere and (c) downward emission of the atmosphere that is reflected by the surface and attenuated in its upwelling path by the intervening atmosphere.

The satellite altitude, emissivity and sea surface temperature are designated by H, $\epsilon_s(\nu)$ and T_s.

Equation (1) can be rewritten in an alternate form as follows :

$$T_B(\nu) = T_s\,[a(\nu) - \beta(\nu)\,\tau_\nu^2\,(0, H)\,(1 - \epsilon_s\,(\nu)], \tag{2}$$

where $$a(\nu) = 1 + \int_0^H (1 - \tau_\nu(Z, H))\,\frac{1}{T_s}\,\frac{\partial T(Z)}{\partial Z}\,dZ, \tag{3}$$

$$\beta(\nu) = 1 - \int_0^H (1 - \tau_\nu(Z, H))\,\frac{1}{T_s \tau_\nu(Z, H)}\,\frac{\partial T(Z)}{\partial Z}\,dZ, \tag{4}$$

using atmospheric data, the expressions for $a(\nu)$ and $\beta(\nu)\ \tau_\nu^2\,(0, H)$ could be written in terms of total transmittances.

$$a(\nu) = A + B\tau + C\tau^2, \tag{5}$$

$$\beta(\nu)\,\tau_\nu^2\,(0, H) = D + E\tau + F\tau^2, \tag{6}$$

$$\tau_\nu = \frac{-(B - Er) \pm \{(B - Er)^2 - 4\,(C - Fr)\,(A - Dr - [T_B(\nu)/T_s])\}^{\frac{1}{2}}}{2\,(C - Fr)} \tag{6a}$$

where $r = 1 - \epsilon_s(\nu)$.

The constants, A, B, C, D, E and F are determined using 300, atmospheric simulations with both clear and cloudy atmospheres. The solution of (2) after

Table 1. Values of the constants in the regression equations (5) and (6).

Frequency GHz	A	B	C	rms error	D	E	F	rms error
$\nu_1 = 19{\cdot}35$	0·5037	1·1106	−0·6289	0·0029	0·0272	−0·0021	0·9726	0·0005
$\nu_2 = 22{\cdot}235$	0·8611	0·2679	−0·1384	0·0012	−0·0023	0·0674	0·933	0·0009

substituting (5) and (6) is given in terms of brightness temperature by (6a). The clouds with liquid water contents in the multiple of 0·15 g m^{-3} with maximum 0·60 g m^{-3} were used in the simulation. The base and top of the cloud were also varied (1·5–5·5 km). The constants in the regression equations (5) and (6) and the r.m.s. error in the determination of $a(\nu)$ and $\beta(\nu)\,\tau_\nu^2(0, H)$ using exact expressions [equations (3)–(4)] are given in table 1.

In the case of cloudy atmosphere, the main absorbers of the electromagnetic radiation are water vapour, and oxygen molecule in gaseous phase and liquid water in the form of clouds. The total transmittance can therefore be written as

$$\tau_\nu = \tau_{cl}(\nu) \cdot (\tau_{H_2O}(\nu) \cdot \tau_{O_2}(\nu)). \tag{7}$$

The relation between bracketted parts of (7), for the two frequencies 19 and 22·235 GHz, obtained from atmospheric calculations, yielded the following regression equation:

$$\tau_{H_2O}(19)\,\tau_{O_2}(19) = a\,\tau_{H_2O}(22)\,\tau_{O_2}(22) + b, \tag{8}$$

where $a = 0{\cdot}3846$ and $b = 0{\cdot}6155$. The transmittance due to cloud is given by

$$\tau_{cl}(\nu) = \exp(-k\nu^2 Q), \tag{9}$$

where ν is in GHz and the parameter k is given by

$$k = 1{\cdot}11 \times 10^{[0{\cdot}0122\,(291 - T_{cl}) - 4]}$$

where T_{cl} = cloud temperature in degree K. From (9), the transmittances due to cloud at 19·35 and 22·235 GHz frequencies are related as

$$\tau_{cl}(\nu_2) = (a' + b'Q + c'Q^2)\,\tau_{cl}(\nu_1) \tag{10}$$

where the constants are given as $a' = 1{\cdot}0$, $b' = -0{\cdot}0318$, $c' = 0{\cdot}0005$.

Equations (7)–(10) can be solved in a straightforward manner to yield $\tau(22)$ and $\tau_{H_2O}(22)\,\tau_{O_2}(22)$.

From (7), (8) and (10), and using atmospheric simulations it is possible to derive the following expressions for total atmospheric water vapour and liquid water contents :

$$W = 19{\cdot}2676\,(1 - 1{\cdot}0365\,\tau_{H_2O}(22)\,\tau_{O_2}(22)) \tag{11}$$

and

$$Q = 7{\cdot}6279\left(1 - \frac{\tau(22)}{\tau_{H_2O}(22)\,\tau_{O_2}(22)}\right). \tag{12}$$

However, it should be noted that the above procedure requires the values of $T_{cl} \sim$ (260° K in the present calculations), sea surface temperature ($T_s \sim 300°$ K) and the emissivity calculated from Fresnel's formula. The increased emissivity due to wind-generated foams and surface roughness could be accounted using the empirical equation given by Hollinger *et al* (1975)

$$\triangle T_B = 0{\cdot}134 \, \triangle W f^{1/2} \tag{13}$$

where $\triangle W$ is in knots (> 7 knots) f in GHz, and $\triangle T_B$ in degree K.

3. Results and discussions

The results of the computation using simulation technique are presented in table 2 along with results obtained from other methods. It is seen that the agreement is satisfactory. The antenna pattern correction has not been taken into account while processing the SAMIR data.

The effect of different atmospheres used in the simulation to derive the coefficients has also been analysed and the results given in table 3 show that all the three regression equations give the values of W and Q and are within about ± 5%.

Based on the simulation method discussed earlier, it is possible to map the total water vapour and study its variations along with auxiliary data such as visible and infrared bands. Such types of global maps have been derived by Grody (1978) using the data from a scanning microwave spectrometer (SCAMS) that was flown on Nimbus 6.

Maps for total precipitable water and remotely-sensed sea surface temperature can be used to infer the gross characteristics of the boundary layer over the oceanic regions. This is possible because, on an average, the vertical distribution of water vapour in the atmosphere over the water bodies can be represented with a

Table 2. Comparison of total water vapour content (g cm^{-2}) and liquid water-content (kg m^{-2}) as determined from different methods.

Day 1979	Orbit No.	Simulation		Statistical		Empirical		Radiosonde	
		W	Q	W	Q	W	Q	W	Q
July 6	434	3·50	1·1	3·68	1·05	4·39	..	4·4	..
July 8	464	3·92	0·89	4·19	1·03	4·70	..	4·4	..
July 15	575	4·28	0·57	4·09	0·82	4·62	..	4·4	..
August 1	829	5·04	0·85	4·70	1·12	5·07	..	5·3	..
August 3	860	5·17	0·60	4·58	0·94	4·79	..	5·2	..
August 22	1149	3·65	0·56	3·72	0·73	4·34	..	4·0	..
August 23	1164	3·57	0·63	3·75	0·78	4·36	..	4·8	..
September 3	1327	5·92	0·36	4·78	0·84	5·06	..	..	..
October 10	1812	4·59	0·56	4·25	0·84	4·70	..	..	..
October 17	1998	4·55	0·72	4·37	0·97	4·81	..	..	..
October 21	2057	2·99	0·67	3·46	0·74	4·16	..	..	..

Table 3. Total water vapour ($g\,cm^{-2}$) and liquid water contents ($kg\,m^{-2}$) determined from simulation method using three different atmospheric conditions for simulation.

Day 1979	Orbit No.	Latitude	Longitude	$T_B(19)$ (° K)	$T_B(22)$ (° K)	1st regression		2nd regression		3rd regression	
						W	*Q*	*W*	*Q*	*W*	*Q*
July 8	464	07–42	72–33	167·8	209·4	3·90	0·88	3·99	0·86	3·89	0·88
July 15	575	10–55	90–11	159·8	206·2	4·28	0·57	4·32	0·56	4·27	0·57
August 1	829	12–42	73–02	172·0	218·6	5·04	0·84	5·10	0·81	5·07	0·83
August 3	860	16–08	70–36	165·3	215·3	5·17	0·60	5·21	0·58	5·21	0·59
August 22	1149	17–05	69–14	156·0	199·4	3·65	0·56	3·68	0·55	3·61	0·57
August 23	1164	07–23	74–19	157·8	200·2	3·57	0·63	3·62	0·62	3·54	0·64
September 1	1297	23–20	87–70	169·0	218·0	5·24	0·71	5·30	0·68	5·30	0·69
September 3	1327	21–32	83–35	162·0	218·0	5·92	0·36	5·94	0·35	5·99	0·34
October 5	1812	14–53	80–51	161·0	209·0	4·59	0·56	4·63	0·54	4·60	0·56
October 17	1998	07–20	80–31	166·0	212·0	4·55	0·72	4·61	0·70	4·56	0·72
October 21	2057	18–10	91-46	156·0	195·0	2·30	0·67	3·04	0·66	2·94	0·68
November 7	2311	08–01	84–35	193·0	219·0	2·04	2·03	2·06	2·03	1·93	2·09
November 14	2423	11-31	82–42	173·0	219·0	4·99	0·88	5·05	0·86	5·03	0·87
November 22	2542	09–54	88–53	177·0	223·0	5·27	0·98	5·34	0·95	5·35	0·96

simple model. Then some average value of precipitable water *W* can be associated to a given surface temperature. The departure of the measured *W* from the corresponding *W* relates to the atmospheric stratification in the boundary layer. For instance when $W > \bar{W}$, convective conditions (ITCZ) are present and when $W < \bar{W}$ stable condition prevails (Prabhakara *et al* 1978).

In addition, such types of maps could as well be used for calculating fluxes of total precipitable water affecting the monsoon. Ghosh *et al* (1978) have studied the Indian summer monsoon based on the conventional data alone but it is now conceivable to utilise satellite data for similar studies.

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