Photosynthetically available radiation in the central and eastern Arabian Sea

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In this article, we present the analysis of the photosynthetically available radiation (PAR, 400–700 nm) in the eastern and central Arabian Sea (21° 30' N, 64°E to 12°N, 74°E) during February–March, 1995 using data collected with quantum PAR sensor and maritime solar radiation models for clear sky. The peak values observed were in the range of about 1700 to 1830 μmole $s^{-1}m^{-2}$ or equivalent to about 365 to 435 Wm^{-2}. The radiative transfer models of Gregg and Carder, and Frouin *et al.* were used to investigate the relationship between the quantum and energy of solar radiation available in the PAR region. The average ratio of this quanta per energy was found to be about 4.2 μmole $W^{-1}s^{-1}$. Aerosol optical depth obtained using the sunphotometer at 498 nm for the region was found to vary from 0.07 to 0.19.

LIFE in the ocean is dependent on the amount of sunlight available. The amount of solar irradiance reaching the ocean surface is important in all disciplines of oceanography. From a physical oceanography point of view, the total incident solar irradiance constitutes a major boundary forcing for ocean circulation and determines meridional heat transport. From the biological point of view, solar irradiance in the photosynthetically active interval PAR, regulates marine primary productivity and therefore the evolution of aquatic ecosystems[1,2].

The Arabian Sea is a highly productive oceanic region. One of the goals of JGOFS is to understand the process in controlling the carbon flux in the Arabian Sea. Its unique feature is the regular oscillation of high rates of primary production under relatively constant levels of solar radiation[3]. Though the sunlight is available on the surface at a wide spectral range, the useful light available in the upper layers of ocean is limited in PAR region in the range of 350–700 nm. PAR is defined in terms of energy (Wm^{-2}) as

$$E_{PAR}(z) = \int_{350}^{700} Ed(\lambda,z)d\lambda, \qquad (1)$$

where $Ed(\lambda,z)$ is the downward spectral irradiance (Wm^{-2}) at wavelength λ (nanometers) at depth z (meters). The total quanta available in the photosynthetic range is given by

$$Q_{PAR}(z) = 1/hc\int_{300}^{700} \lambda\, Ed(\lambda,z)d\lambda, \qquad (2)$$

where h is the Plank's constant and c is the velocity of light in vacuum. Q_{PAR} has units of quanta $s^{-1}m^{-2}$.

Usually PAR is taken as a constant fraction of total solar irradiance. PAR includes more than 45% of total radiation reaching the earth's surface for low zenith angle or high elevation above 30°.

Materials and methods

During February–March, 1995 PAR values were measured in the eastern and central Arabian Sea (21°30' N, 64°E to 74°E) under the JGOFS (India) programme aboard the research vessel *ORV Sagar Kanya* (Table 1). Q_{PAR} values were measured using quantum PAR sensor LI-190S (LI-COR Inc., USA) (Figure 1). The sensor was placed at a clear site on a raised platform on the ship to obtain sunlight without any obstruction. The PAR range of this sensor is 400–700 nm. The unit is given in μmole s^{-1} m^{-2}. (1 μmole s^{-1} m^{-2} = 6.022. 10^{17} quanta s^{-1} m^{-2}). E_{PAR} is evaluated using two solar radia-

Table 1. Stations covered in Arabian Sea for the measurements of PAR data

Station	Data	Latitude N	Longitude E
1	12.2.95	21° 22'	64° 10'
2	13.2.95	21° 25'	64° 16'
3	21.2.95	15° 00'	65° 50'
4	22.2.95	14° 58'	63° 49'
5	23.2.95	14° 55'	63° 48'
6	25.2.95	11° 00'	64° 00'
7	26.2.95	10° 55'	65° 30'
8	1.3.95	11° 40'	74° 20'

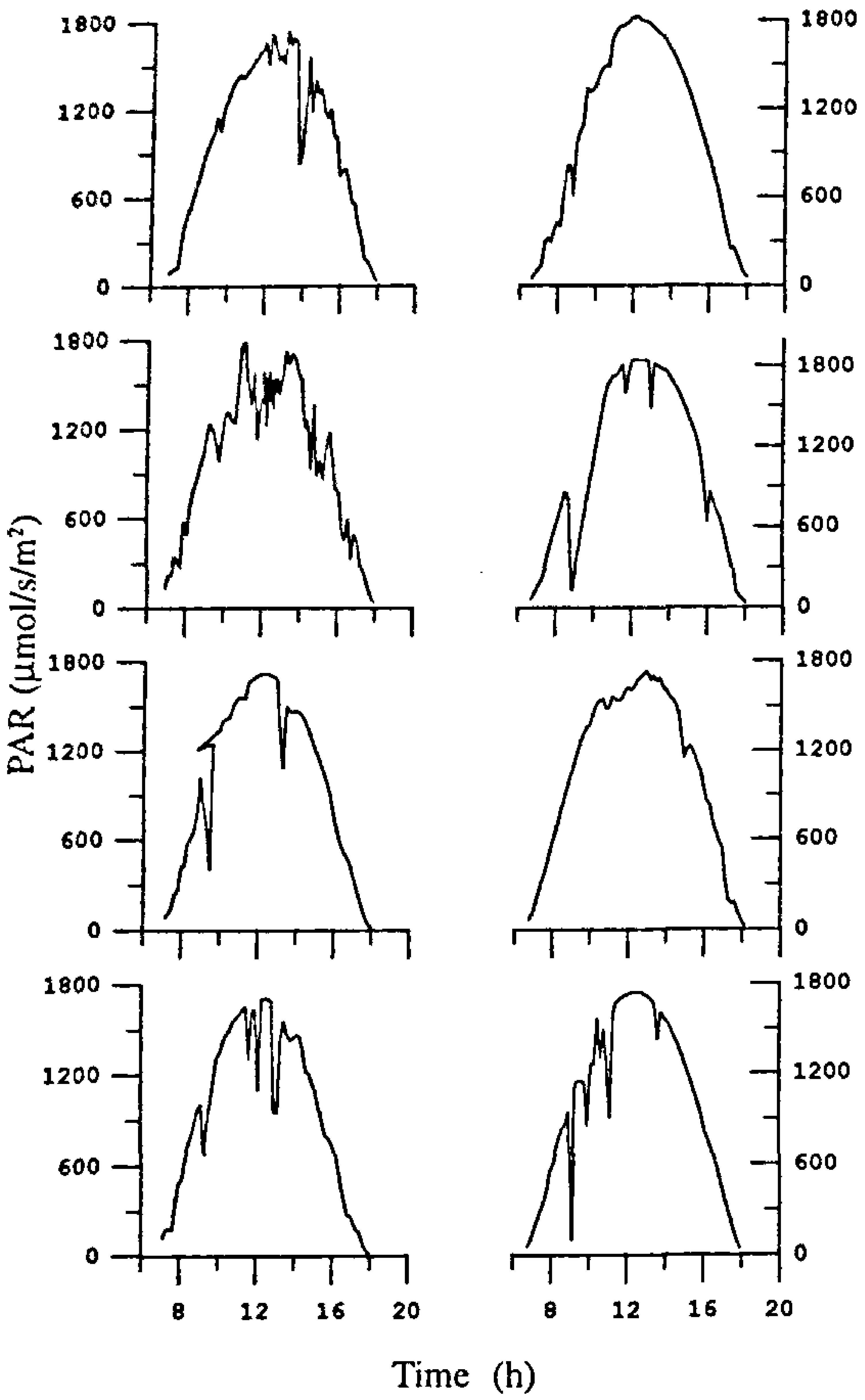

Figure 1. PAR data using sensor LI-190 at different stations.

tion models of Gregg and Carder[4], and Frouin *et al.*[5]. The meteorological parameters were measured with an automatic logging meteorological station[6,7]. The atmospheric pressure recorded every minute was averaged over a ten minutes interval, while wind speed, wet and dry bulb temperatures were recorded every 3 hours. A single wavelength hand held sunphotometer was used to obtain the aerosol optical depth (wavelength 498 nm, FWHM = 13 nm). The unit was calibrated using the Langley method at a clear sight at a high altitude station, Mount Gurushikhar.

Since all measurements were made with reference to Indian Standard Time (Longitude 82° 30' E), corrections were applied to obtain the local time at the meridian of observation. In the model, airmass is assumed to be unity, which is typical for maritime atmospheres. Ozone and water vapour contents were not measured. Average values for water vapour were obtained from the atlas of Ramesh Kumar *et al.*[8], average ozone content was taken as 320 Dobson Units, which gives ozone scale height to be 0.32 cm. Total precipitable water vapour was taken from the atlas[9] as 5.5 g cm^{-2}.

Solar radiation models

Spectral model of Gregg and Carder[1]

The high resolution solar spectral irradiance model of Gregg and Carder[4] was used to evaluate E_{PAR}. This model was adopted as it was suited for maritime atmosphere and it provided spectral information at a high resolution of 1 nm. The model computed contributions from direct and diffuse sunlight available at any time in the range of 350–700 nm. We have used the range of 400–700 nm for compatibility with the quantum sensor output.

Attenuation of solar irradiance in the visible and near-UV wavelengths is attributed to atmospheric processes such as scattering by gas mixture (Rayleigh scattering), absorption by gas mixture, absorption by ozone, and scattering and absorption by water vapour. Irradiance that is not scattered out of the direct beam but toward the surface is the diffuse irradiance. The sum of the direct and diffuse components defines the global irradiance available for use.

$$Edd(\lambda) = F0(\lambda)\cos(\theta)\, Tr(\lambda)\, Ta(\lambda)\, Toz(\lambda)\, To(\lambda)\, Tw(\lambda), \quad (3)$$

$$Eds(\lambda) = Ir(\lambda) + Ia(\lambda), \quad (4)$$

where *dd* and *ds* refer to direct and diffuse components, $F0(l)$ is the mean extra terrestrial irradiance corrected for earth–sun distances and orbital eccentricity, θ is the solar zenith angle, *Tr, Ta, Toz, To* and *Tw* represent transmittance after absorption or scattering by air molecules, aerosol, ozone, oxygen and water vapour respectively. Similarly *Ir* and *Ia* represent the diffuse components due to Rayleigh scattering and aerosol.

For maritime environment, the above equations on incorporating air–sea interactions change to

$$Edd(\lambda) = Edd(\lambda).\,(1-\rho_d), \quad (5)$$

$$Eds(\lambda) = Eds(\lambda).\,(1-\rho_s), \quad (6)$$

where ρ_d is the direct sea surface reflectance and ρ_s is the diffuse reflectance.

Aerosol optical depth

Marine aerosols, being larger than continental aerosols, reduce solar radiation by absorption and large scattering. Total optical depth is a sum of optical depths due to Rayleigh scattering, (τ_R) aerosol extinction (τ_A) and gaseous absorption (τ_G). Total columnar optical depth

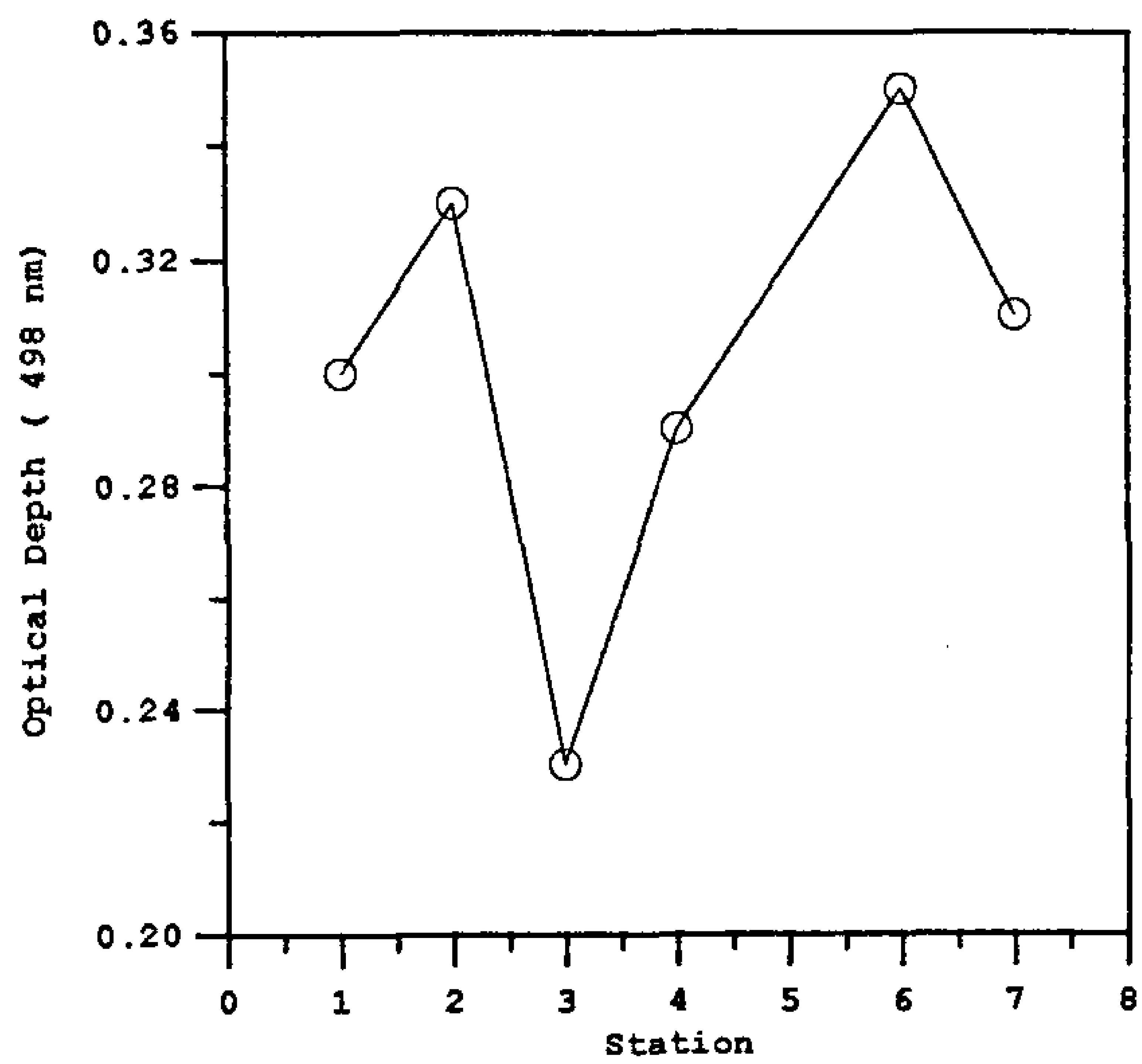

Figure 2. Total columnar optical depth at various stations

measured at 498 nm using the sunphotometer (Figure 2) is given as

$$\tau(\lambda) = \tau_R(\lambda) + \tau_A(\lambda) + \tau_G(\lambda). \quad (6)$$

Rayleigh optical depth is expressed as [10, 11]

$$\tau_R(\lambda) =$$

$$(p/p_0)\ 0.00865.\lambda^{-(3.916+0.074\lambda + 0.050/\lambda)}. \quad (7)$$

Here λ is given in micrometer, p is the barometric pressure (mb) and p_0 is the standard atmospheric pressure ($p_0 = 1013.25$ mb).

Optical depth component due to ozone is calculated at $\lambda = 498$ nm and is given by $\tau_{OZ}(498) = 0.01$ and contributions due to other gases are assumed to be the same as at 484 nm, whose value is given as, $\tau_G(484) = 0.005$. Using the equation (7) for τ_R at 498 nm and optical depth for gases as given above, we calculate the aerosol optical depth $\tau_A(498)$ at each station. Since no data for optical depth was available at stations 8, for this station we have taken the average value of the total columnar optical depth of 0.3. With the help of aerosol optical depth at 498 nm, we determine the aerosol depths at other wavelengths using the Angstrom formula

$$\tau_A(\lambda) = \beta\,\lambda^{-\alpha} \qquad (\lambda \text{ in } \mu\text{m}). \quad (8)$$

Here β is the turbidity coefficient representing the aerosol concentration and α is the Angstrom component. Angstrom component α is given in terms of Junge exponent γ as[13]

$$\alpha = -(\gamma + 3). \quad (9)$$

Aerosol size distributions can be expressed by two expressions, Junge distribution[14] and by an expression dependent on relative humidity and wind speed[15]. The latter expression is evaluated for three radii of particles, 0.1, 1.0 and 10 nm. Logarithmic regression of these with the three radii is computed and the regression coefficients give the Junge exponent. Using equation (9) we then obtain the Angstrom component. Having obtained α, β can be obtained from equation (8) with the observed value of $\tau_A(498)$. $\tau_A(\lambda)$ is then calculated for all other wavelengths.

Model of Frouin *et al.*

This is a simple, yet accurate analytical formula to determine the PAR at the ocean surface under clear skies[5]. The formula is given as follows

$$E_{PAR} = F_{PAR}\ T_1 T_2 T_3 \quad (10)$$

F_{PAR} is the monochromatic extraterrestrial irradiance integrated over 400–700 nm, taking into consideration the earth–sun separation.

$$T_1 = \frac{\cos(\theta)\exp[-a+b/V)/\cos(\theta)}{1-\rho\,(a' + b'/V}, \quad (11)$$

$$T_2 = \exp[-av\,(U_v/\cos(\theta))^{bv}], \quad (12)$$

$$T_3 = \exp[-a_0(U_0/\cos(\theta))^{bo}], \quad (13)$$

ρ is average surface reflectance, V visibility, U the vertically integrated absorbed amount, θ the zenith angle, and subscripts o and v denote ozone and water vapour respectively.

The regression coefficients for maritime atmosphere over spectral region 400–700 nm are given as

$a = 0.068$ $b = 0.379$ $a' = 0.117$, $b' = 0.493$,

$av = 0.002$, $bv = 0.87$ $a_o = 0.052$ $b_o = 0.99$.

Visibility V (km) is calculated using the aerosol optical depth data. Aerosol extinction coefficient at 550 nm is expressed in terms of visibility[16]

$$C_a(550) = 3.91/V. \quad (14)$$

This extinction coefficient is related to aerosol optical thickness by

$$\tau_A(550) = C_a(550)\,H_a, \quad (15)$$

where H_a is the aerosol scale height, which is assumed to be 1 km (ref. 17). We can determine aerosol optical depth at 550 nm using the method described earlier in Gregg and Carder[4] model.

Results and discussion

The two models were evaluated for eight stations in the Arabian Sea. They were found to agree closely with the real time data collected using a quantum PAR sensor. With the exception of ozone and water vapour contents, all other meteorological parameters required for the models were measured at the stations. The measured peak values of PAR were found to vary from about 1700 to 1830 mmole $s^{-1} m^{-2}$ or equivalent to about 365 to 435 W m^{-2} evaluated using the solar radiation models. Figure 3 shows the measured and model data given for a station. In the absence of observed meteorological data for the duration of calculation, the last observed values were considered, assuming that variations over the period were not significantly large. The models seem to agree better at stations when there were less disturbances in PAR values, such as at station 11°N, 64°E (25 February) and 11° 40'N, 74° 20'E (3 March). A discontinuity was usually observed with the quantum sensor data around 1000 hours, which is attributed to the shadow of a mast on the deck while the others at 1300 to 1500 hours probably caused by clouds. Since the models were tuned for clear sky, the observed values were filtered to remove all such spurious data.

We have used the models to analyse the behaviour of $Q_{PAR} : E_{PAR}$. The ratio of observed PAR quantum values to the solar irradiance values evaluated using the models of Gregg and Carder[4] and Frouin *et al.*[5] were 4.08 and 4.04 respectively (Figures 4 and 5). After filtering the data for disturbances, the ratios were boosted to values 4.24 and 4.20 for the Gregg and Carder and Frouin *et al* models respectively. The difference in the ratios for both the models was found to be marginal (< 0.02%), when the solar irradiance was estimated for zenith angles below 68°, 80° and 90°. Morel and Smith[18] obtained a value of 2.77 × 10^{18} quanta $W^{-1} s^{-1}$ or 4.6 μmole $W^{-2} s^{-1}$ for 400–700 nm. (1 μmole $s^{-1} m^{-2}$ = 6.022. 10^{17} quanta $s^{-1} m^{-2}$). Their observations were for solar zenith angles < 68° and the ratio is for monochromatic radiant energy at 550 nm, the central wavelength of the spectral region 400–700 nm. Baker and Frouin[19] obtained a ratio of 2.68 × 10^{18} quanta $W^{-1}s^{-1}$ or 4.38 μmole $W^{-1}s^{-1}$d λ is taken as the central wavelength, in the range 350–700 nm.

Investigation of aerosol content is also part of the study under JGOFS (India) and it is also an important parameter required for remote sensing, validation of algorithms and models. Atmospheric turbidity obtained using sunphotometer at 498 nm was found to vary from

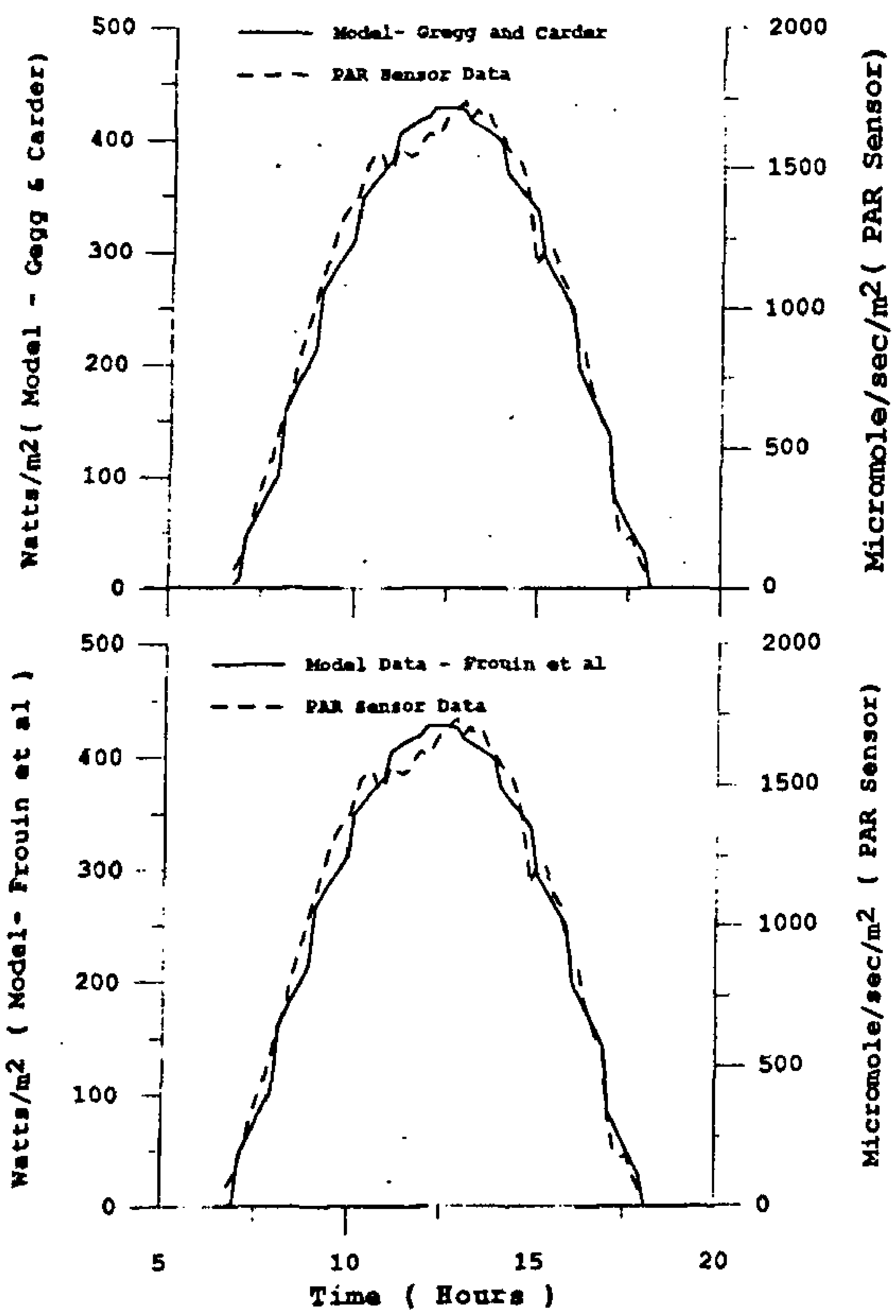

Figure 3. PAR sensor and model data for station 6 (25 March).

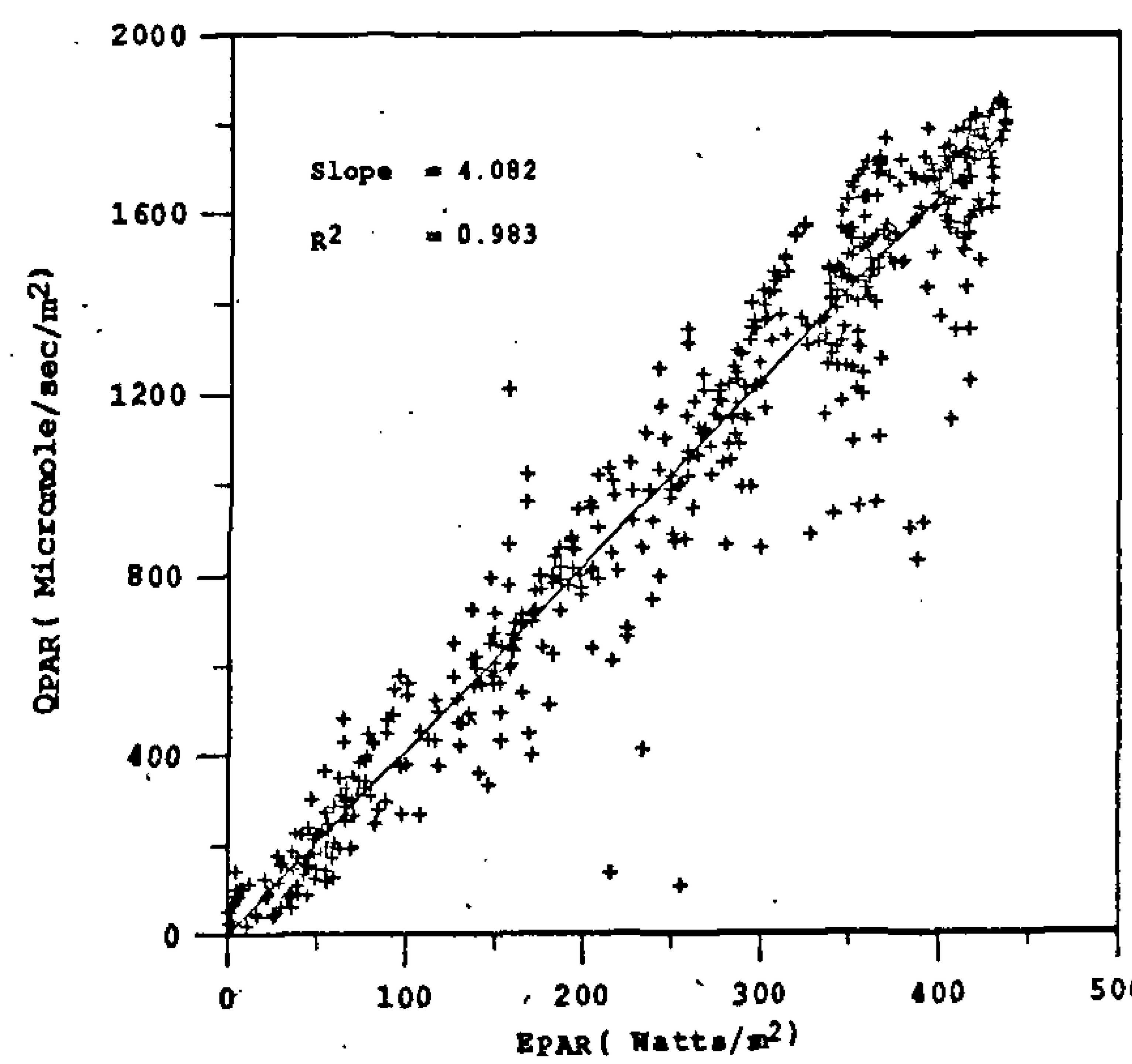

Figure 4. PAR sensor data vs solar irradiance model of Gregg and Carder.

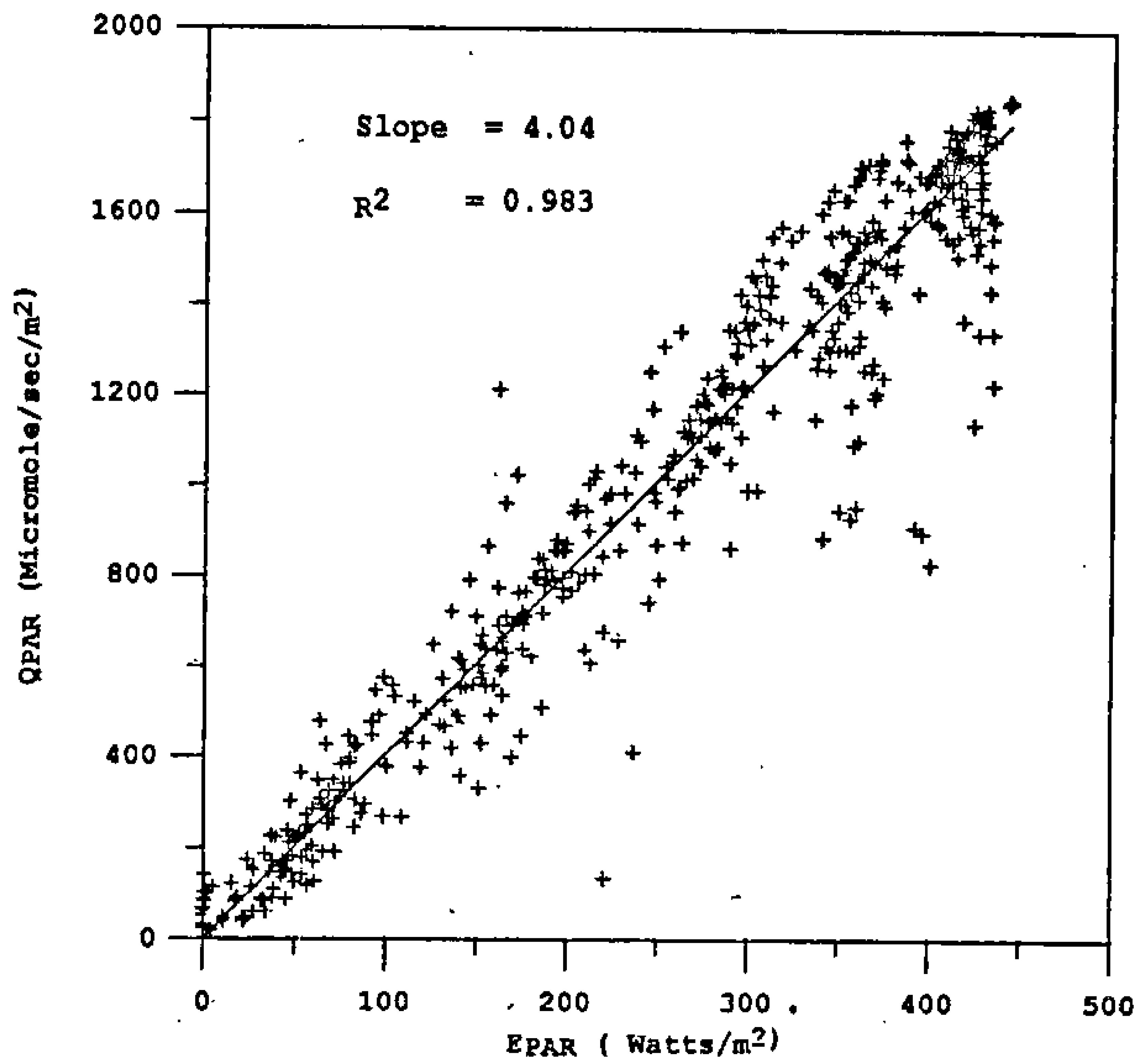

Figure 5. PAR sensor data vs solar irradiance model of Frouin *et al.*

0.23 to 0.35. The aerosol optical depth derived from this was found to vary from 0.07 to 0.19.

The analytical model of Frouin *et al.*[5] was found to be simple compared to the rigorous high resolution spectral model of Carder and Gregg[4]. The model of Frouin *et al.* could be used to obtain the much required parameter of aerosol optical depth. Solar irradiance values in PAR range can be measured using the PAR sensor and the same can be equated to the model of Frouin *et al.*[2] to obtain the aerosol optical depth.

1. Platt, T., *Deep-Sea Res.*, 1986, **33**, 149.
2. Satheyendranath S. and Platt, T., *J. Geophys. Res.*, 1988, **93**, 9270.
3. *Arabian Sea Process Study*, US JGOFS, vol. 13, 1991.
4. Gregg, W. W. and Carder, K. L., *Limnol. Oceanogr.*, 1990, **35**, 1657.
5. Frouin, R., Ligner, D. W., Gautier, C., Baker K. S. and Smith, R. C., *J. Geophy. Res.*, 1989, **94**, 9731.
6. Prabhu Dessai, R.G. and Desa, B. A. E., Proceedings MDS-86, Int. Data Systems Symposium, New Orleans, USA, 1986, pp. 290.
7. Prabhu Dessai, R. G., Desa, E. and Vithayathil, G., *Oceanography of the Indian Ocean*, Oxford & IBH, 1991, pp. 719.
8. McClain, C., J., Comiso, R., Fraser, J., Firestone, B., Schieber, E., Yeh, K., Arrigo and Sullivan, C., *SeaWiFS Technical Report Series*, **13**, 1994.
9. Ramesh Kumar, P.V., Sathe, P. M., Muralidharan, L. V. and Rao, G., *An Atlas of Precipitable Water Over the Tropical Indian Ocean for 1979*, Technical Report, 1993.
10. Frohlich, C. and Shaw, G. E. *Appl. Opt.*, 1980, **19**, 1773.
11. Young, A. T., *J. Appl. Meteorol.*, 1980, **20**, 328.
12. Korotaev, G. K., Sakerin, S. M., Ignatov, A. M., Stowe, L. L. and McClain, E. P., *J. Atmos. Ocean Tech*, 1993, **10**, 725.
13. Van de Hulst, H.C., *Light Scattering by Small Particles*, Dover, Dodrecht, 1981.
14. Junge, C. E., *Air Chemistry and Radioactivity*, Academic, New York, 1963.
15. Gathman, S. G., *Opt. Eng.*, 1983, **22**, 57.
16. Fitzgerald, J. W., *Appl. Opt.*, 1983, **28**, 3534.
17. Gordon, H.R. and Castano, D. J., *Appl. Opt.*, 1987, **26**, 2111.
18. Morel, A. and Smith, R. C., *Limnol. Oceanogr.*, 1974, **19**, 591.
19. Baker, K. S. and Frouin, R., *Limnol. Oceanogr.*, 1987, **32**, 1370.

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