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TURBIDITY OF AUSTRALIAN SKIES

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ABSTRACT

A mean Angström turbidity of 0.11 is derived for Australian skies. The method is based on an algorithm for computing clear sky global irradiance and uses global pyranometer data, which are more widely available than the spectral observations of direct beam irradiance conventionally employed. Comparisons with conventional turbidity measurements at Darwin and Aspendale provide support for the method.

INTRODUCTION

The turbidity of the atmosphere is a function of the amount of particulate matter in suspension, and is one of the factors influencing the amount of solar radiation reaching the earth's surface. A need for turbidity values exists for application in numerical models of the planetary boundary layer in which solar irradiance is computed at ground level, *eg*, Atwater and Brown (1974).

In a turbid atmosphere the predominant effect of suspended particles upon solar radiation is to scatter energy from the direct solar beam. Some of this scattered energy augments the diffuse radiation reaching the earth's surface. The overall effect is to reduce global irradiance, that is the sum of direct and diffuse irradiances, at ground level. Because of the dependence of scattering depletion upon wavelength it is possible and in fact usual to determine turbidity from measurements of direct beam irradiance in different spectral regions. There is a scarcity of such observations over Australia, and an attempt is made here to use unfiltered global irradiance measurements, which are available at a greater number of sites. The rationale employed in this study is to take an observation of clear sky solar global irradiance, and then with a knowledge of the solar distance and elevation, precipitable water and other factors relevant to the absorption and scattering of the solar beam, to seek a value of turbidity which provides a computed global irradiance matching that observed.

The particular global irradiance observations used were originally made to detect changes in instrument sensitivity in the Bureau's global radiation network. They consist of spot checks of clear sky global irradiance at local apparent noon, made about once monthly over several years. Data are taken from eight stations selected chiefly on the basis of availability and quality of data. Bias may be introduced into the computed turbidity values by restriction of the data sample to occasions of clear sky.

METHOD

The set of equations chosen to relate global irradiance to turbidity, based largely on Schüepp (1966), can not be inverted to provide an explicit expression for turbidity. Consequently an iterative method is used to arrive at a satisfactory turbidity value; a first guess value for turbidity is successively refined until the difference between computed and observed global irradiance is acceptably small.

Turbidity appears as an exponent in equations for scattering depletion. The two measures of turbidity referred to in this paper are the Schüepp decadic turbidity and the Angström turbidity. Choice between them is a matter of convenience. They are related as indicated by Eqn 3. Schüepp decadic turbidity B may be defined by

$$(t_s)_i = 10^{-Bm(p/1000)(2\lambda_i)^{-\alpha}}, \qquad \ldots 1$$

where $(t_s)_i$ is the fractional transmission of direct radiation in some wave band i, when depletion is due only to scattering by suspended dust, m is the cosec of the solar elevation, p the surface pressure in mb, λ_i the band wavelength in μm, and α is referred to as the wavelength exponent for scattering and is usually assigned a value of 1.3 or 1.5. Also the Angström turbidity factor β may be defined from

$$(t_s)_i = e^{-\beta m(p/1000)\lambda_i^{-\alpha}}, \qquad \ldots 2$$

where e is the base of natural logarithms. It follows from Eqn 1 and 2 that

$$2^{-\alpha} B \ln 10 = \beta. \qquad \ldots 3$$

With α = 1.5 as adopted in this paper to accord with Schüepp (1966), Eqn 3 implies that

$$B = 1.228\beta. \qquad \ldots 4$$

Schüepp turbidity B is used in the subsequent equations following Schüepp (1966).

Turbidity can be considered to play a dual role in relation to clear sky solar global irradiance, decreasing direct beam irradiance by scattering whilst increasing diffuse irradiance. The following equations describe the processes involved. Solar global irradiance S_G can be considered to be the sum of three components:

$$S_G = S + D + A, \qquad \ldots 5$$

where S is the direct beam irradiance, D is diffuse irradiance expected in the absence of ground reflection and A is diffuse irradiance arising from the return of some of the radiation being reflected skyward from the earth's surface.

Direct beam irradiance S is calculated for radiation in the wavelength range 0.25 to 3.05 μm. This contains about 98% of the energy in the solar beam. Nearly all of the remaining energy outside this range is absorbed before reaching the earth's surface; the longer wavelengths by carbon dioxide and water vapour, the shorter by ozone. The wavelength range 0.25 to 3.05 μm is divided into 28 intervals of 0.1 μm width, denoted by the subscript i in subsequent equations. Each interval is assigned an irradiance H_i according to Table 1. Table 1 is adapted from Robinson's (1966) Table 1.2, after correcting a printing error at 0.9 μm.

Table 1 Spectral irradiance of the solar beam in intervals of 0.1 μm centred on the tabulated wavelengths, after Robinson (1966).

Wavelength (μm)	Irradiance ($W\ m^{-2}$ per 0.1 μm)	Wavelength (μm)	Irradiance ($W\ m^{-2}$ per 0.1 μm)
0.3	61	1.7	18
0.4	154	1.8	15
0.5	198	1.9	13
0.6	181	2.0	11
0.7	144	2.1	9
0.8	113	2.2	8
0.9	89	2.3	7
1.0	73	2.4	6
1.1	61	2.5	5
1.2	50	2.6	4
1.3	41	2.7	4
1.4	33	2.8	3
1.5	27	2.9	3
1.6	22	3.0	3

Since scattering and absorption are wavelength dependent, the fractional transmission of energy in each band is computed separately. For each computation three kinds of depletion are considered; scattering by pure air, scattering by aerosol particles, and absorption by oxygen, carbon dioxide and water vapour. The extent of depletion by each of these processes is a function of the length of the path of the direct beam through the atmosphere *en route* to the earth's surface. The path length is expressed non-dimensionally in multiples of path length for a vertical beam. Thus

$$m = \text{cosec}\ (h) \qquad \ldots 6$$

where h is the solar elevation at the time of observation. The transmission factor t_i for wave band i is then given by

$$t_i = 10^{-[0.00386m\lambda_i^{-4.05} + Bm(p/1000)(2\lambda_i)^{-\alpha} + A_i(mW)^{C_i}]} \qquad \ldots 7$$

where A_i and C_i are coefficients derived from Schüepp's (1966) Table 4.1 of absorption data. Values of A_i and C_i are set out in Table 2 of the present report. W is precipitable water depth in mm. The three terms on the right hand side of Eqn 7 represent respectively Rayleigh scattering by pure air, scattering by aerosol particles, and absorption by oxygen, carbon dioxide and water vapour, according to Schüepp (1966). The expression $A_i(mW)^{C_i}$ of Eqn 7 is absent in Schüepp's notation. Instead it is represented by a single absorption coefficient a_w and tabulated as a function of mW in Schuepp's Table 4.1. The expression $A_i(mW)^{C_i}$ is introduced in the present study for ease of computing. Values of A_i and C_i have been generated to reproduce the tabulated coefficients exactly for water vapour paths of 10 mm and 40 mm. For other water vapour paths, transmission factors computed using A_i and C_i differ by less than half a percent from values calculated using Schüepp's a_w table.

Table 2 Coefficients for absorptive depletion of solar beam by oxygen, carbon dioxide and water vapour, for Eqn 7.

Wavelength (μm)	A_i	C_i
0.25 - 0.75	0.0	0.0
0.75 - 0.85	0.001289	0.9311
0.85 - 1.05	0.008507	0.6142
1.05 - 1.25	0.01538	0.5766
1.25 - 1.55	0.1385	0.3386
1.55 - 2.05	0.09215	0.2397
2.05 - 3.05	0.2056	0.3254

Two further factors influence the irradiance of the direct solar beam on a horizontal surface at the ground. One is the solar distance R, expressed non-dimensionall as a fraction of the mean solar distance. The other is the solar elevation. To give the direct beam irradiance S, the contributions reaching the earth's surface in each wave band, given by the products of irradiances H_i and transmissions t_i, are added and adjusted for solar distance R and solar elevation h. Thus

$$S = \sin(h)\ (1/R^2) \sum_{i=1}^{28} H_i\, t_i \qquad \ldots 8$$

The diffuse irradiance D expected in the absence of ground reflection is calculated by assuming that half the irradiance scattered from the direct beam is directed upwards and half downwards. If a purely absorptive transmission coefficient t'_i is defined so that

$$t'_i = 10^{-A_i (mW)^{C_i}} \qquad \ldots 9$$

then the value of D is given by

$$D = 0.5 \sin(h)(1/R^2) \sum_{i=1}^{28} H_i\, (t'_i - t_i) \qquad \ldots 10$$

To compute the diffuse irradiance A arising from radiation reflected at the ground, it is necessary to know the albedo a of the ground. The variation of albedo with solar elevation is described by Paltridge (1971) and Zillman (1972). This variation is satisfactorily represented by Eqn 11 for elevations in the range 10 to 90°.

$$a = a_n - 0.007 + 0.00628/(\sin(h) - 0.1365) \qquad \ldots 11$$

where a_n is the albedo for normal incidence. An equation for A is derived by making the following assumptions: the radiation reflected skyward by the ground has an irradiance given with adequate accuracy by (S + D)a. This radiation subsequently undergoes scattering in a similar way to direct beam radiation travelling earthward from the top of the atmosphere. That is to say the ratio of back-scattered to incident irradiance is the same in each case and is equal to $D/(\sin(h)(1/R^2)\sum_{i=1}^{28} H_i)$. It follows then that

$$A = a(S + D)\ D/(\sin(h)(1/R^2)\sum_{i=1}^{28} H_i) \qquad \ldots 12$$

A means of computing a clear sky global irradiance value for a given date, latitude, turbidity, precipitable water depth and albedo for normal incidence is provided by Eqn 5 to 12. Using a first guess Schüepp turbidity of 0.06 a value of global irradiance is computed and compared with that observed. After three or four iterations a solution of turbidity is obtained such that the computed global irradiance is within 0.01% of that observed.

DATA

A network of Eppley pyranometers is operated by the Bureau of Meteorology to record global radiation over Australia. Calibration checks are performed at each station at approximately monthly intervals by observing the global irradiance at local apparent noon on occasions of clear sky. These observations have been accumulated over several years. For the present study such observations are very suitable; the method of analysis involves computing global irradiance for a given set of relevant variables, and the particular conditions of local apparent noon and clear skies simplify the computations and make them considerably more reliable.

Since the computation of global irradiance by this method requires a knowledge of the precipitable water content of the atmosphere, data are taken only from pyranometer stations which are also radiosonde stations. There are twelve such stations. Of these Alice Springs is omitted because of suspected mis-levelling of the pyranometer, and three other stations are omitted because their radiosonde data are not immediately to hand. About twenty observations are taken from each of the remaining eight stations. These stations and their locations are listed in Table 3. For each observation of global irradiance a precipitable water value is computed from the radiosonde flight commenced about three hours earlier. The equation for calculating the precipitable water content (from the surface to 400 mb) is that presented by Pierrehumbert (1972), neglecting the possible presence of vapour below sensor threshold.

Albedos for normal incidence are assigned to the pyranometer sites with the aid of an albedo map due to Paltridge (1971). For coastal stations an albedo is chosen which is intermediate between the value on Paltridge's map and a value of 0.035 appropriate to a sea surface (Zillman, 1972). For this purpose land and sea surface albedos are weighted according to the relative proportions of land and sea areas within 65 km radius of the pyranometer.

ERRORS

Turbidity is typically responsible for a reduction of about ten per cent in global irradiance at the surface, so that a small fractional change in turbidity corresponds to an order of magnitude smaller change in global irradiance. Because of this the method used to compute turbidity is very sensitive to errors in observed global irradiance.

The pyranometers used are temperature compensated so that errors due to the effect of temperature variations on sensitivity are not expected to exceed 1% according to the instrument specification. Combined errors due to drift of calibration and non-linearity of response are not expected to exceed 2% according to specification. As an additional precaution against drift all data are selected from periods of at least one and usually two years duration during which the pyranometer at a given station was operated continuously without replacement, and visual scrutiny of the plotted observations suggests negligible drift. The instrument response is known to depart from proportionality to the sine of incident beam elevation and all observations of global irradiance are corrected accordingly. Corrections range from -6% at 10° solar elevation to zero at 90°. The total error in global irradiance after this correction is not expected to exceed 3%. Global irradiance errors are expected to be substantially site-specific and therefore to contribute bias to computed turbidities at each site.

To assess the errors arising in computed turbidity values as a result of errors in observed global irradiance, input data values are altered by the maximum error expected and the resulting changes in turbidity noted. This is done for several observations at Darwin and at Hobart. An increase of 3% in observed global irradiance gives a decrease of about 0.04 in computed Angström turbidity at both Darwin and Hobart

It is assumed that the major errors in precipitable water depths arise from failure to detect water vapour at concentrations less than sensor threshold. According to data presented by Pierrehumbert (1972) the largest error to be expected in computed precipitable water is about 9 mm. The annual mean error due to this cause is less than 1.5 mm at Darwin and less than 0.8 mm at Hobart. Increasing precipitable water at Darwin and Hobart by 9 mm is found to decrease computed Angström turbidities by 0.02 and 0.04 respectively at these stations. The small annual mean error in precipitable water values due to failure to detect water implies that the systematic over-estimation of annual mean Angström turbidities due to this cause is less than 0.004.

In selecting an albedo for normal incidence near coastlines, land and sea surfaces are weighted according to the relative proportions of land and sea within 65 km radius of the pyranometer. This distance is chosen since a horizontal plane 6 km above the ground (a median height for the atmosphere) and 65 km in radius, subtends a solid angle amounting to 90% of the field of view of the pyranometer. In view of the uncertainty in knowing what is the best radius to choose, and also given that for the case of a non-vertical sun the appropriate reflective area of the earth's surface is not centred on the pyranometer, the albedo values for coastal stations are in question. However, neither moving the centres the maximum distance appropriate to low solar elevations (14 km for Hobart in mid-winter) nor doubling the radii produces changes in weighted albedo in excess of 0.01. Daily and seasonal variations in albedo due to rain and plant behaviour are unknown but are not expected to contribute bias to annual mean turbidity values. It is found that an increase of 0.01 in albedo for normal incidence gives an increase of about 0.002 in computed Angström turbidity at Darwin and 0.004 at Hobart.

Errors in annual mean Angström turbidity due to errors in global irradiance are seen to be an order of magnitude larger than errors arising from precipitable water and albedo errors. Combined site-specific biases in annual mean Angström turbidity due to all three sources are not expected to exceed 0.04 at Darwin and 0.05 at Hobart.

An additional source of error is the global irradiance algorithm itself. Only one of the assumptions embodied in the algorithm is examined in relation to turbidity errors. This is the assumption of a constant value of 1.5 for the wavelength exponent for scattering, α in Eqn1. Individual determinations of α by Albrecht (1956) at Darwin gave a range of 0.07 to 3.1 in clear sky conditions, however it is plausible that annual mean values will be within 0.5 of 1.5. An error of 0.5 in α is found to imply an error of 0.02 in computed Angström turbidity.

RESULTS

To facilitate comparison with recently published values of Angström turbidity (Collins, 1972) the turbidity values obtained in the present study are expressed as Angström turbidities. The mean, range and standard deviation of the computed Angstöm turbidities for each station are shown in Table 3. The station means range from 0.06 to 0.17 with an average value of 0.11. The periods of observation vary from station to station but lie within the four years 1968 to 1971 inclusive. No significant year to year changes in turbidity are revealed, either for all stations combined or singly. There is an apparent winter maximum in turbidity at Forrest, Laverton and Wagga but this is suspected to be an artificial result of neglecting seasonal variation in albedo. No systematic variation of turbidity with latitude is evident.

Table 3 Angström turbidity statistics.

Station	Location		Mean Angström turbidity	Range	Standard deviation	Sample size
	Lat S	Long E				
Albany	34°57'	117°48'	0.11	0.06-0.16	0.03	22
Darwin	12°26'	130°52'	0.09	0.04-0.17	0.04	17
Forrest	30°50'	128°06'	0.14	0.06-0.21	0.03	23
Hobart	42°53'	147°20'	0.17	0.09-0.22	0.04	21
Laverton	37°52'	144°45'	0.06	0.00-0.12	0.03	21
Mount Gambier	37°45'	140°47'	0.14	0.06-0.21	0.03	17
Wagga	35°10'	147°28'	0.07	0.01-0.12	0.03	19
Williamtown	32°49'	151°50'	0.09	0.00-0.16	0.04	14

Previous determinations of turbidity at Australian sites have been made using direct irradiance measurements in two spectral regions. Albrecht (1956) made 450 such observations at Darwin. His method of analysis permits the wavelength exponent for scattering, α, to be determined. The variability of his α values in clear sky conditions confounds immediate comparison between his turbidities and the present results. Accordingly a selection of Albrecht's results has been made such that α is in the range 1.3 to 1.7. This gives 21 Schüepp decadic turbidity values which are converted to Angström turbidities by Eqn 4. Their mean and standard deviation are then 0.08 and 0.03, which compares with 0.09 and 0.04 obtained for Darwin in the present study.

Collins (1972), also using spectral measurements but assuming a constant value of 1.3 for α, has determined Angström turbidity values at Aspendale, Victoria. The mean for three years to 1971 is 0.05 and the range 0.00 to 0.20. These compare with the present results for Laverton, 33 km distant, where the mean and range are 0.06 and 0.00 to 0.12.

CONCLUSIONS

The method described is capable of providing turbidity values corresponding to given observed clear sky global irradiance, precipitable water, surface pressure, albedo for normal incidence, and solar elevation. It is sensitive to errors, particularly in global irradiance. However, if these are minimised by accurate observation or data averaging, then the method is correspondingly accurate. In the form in which it is described the method is not applicable when clouds are present.

When applied to about twenty observations at each of eight Australian sites the method gives an overall average Angström turbidity of 0.11 and a range in the station means of 0.06 to 0.17. Agreement between the present results and previous conventional spectral determinations of turbidity at Darwin and Aspendale, Victoria, is much better than the estimated maximum site-specific bias of 0.05, although this may be fortuitous.

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