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ATMOSPHERIC TURBIDITY VARIATIONS IN WELLINGTON - A PILOT STUDY

E. Farkas

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Abstract

A pilot study of atmospheric turbidity measurements was carried out at Wellington during January 1975 to February 1978 using a three-channel Volz sun-photometer.

The various parameters of atmospheric turbidity derived from these observations are presented and discussed in this paper.

It was found that although the long term constancy of the filter characteristics in the photometer could not be monitored and the calibration constants were empirically established from only a limited number of observation series, the results were of the right order of magnitude and compared well with similar data at other places. Considerable doubt exists however as to whether the mean turbidity values obtained for the summer season represent real atmospheric conditions or have been influenced by sampling or instrumental errors.

The results of a short comparison between three Volz photometers are also presented and discussed.

Recommendations are made for further studies of atmospheric turbidities.

1. Introduction

(a) The instrument

The Volz sun-photometer is a relatively simple multispectral radiometer (Volz, 1974). Various combinations of absorption and interference filters in the visual and infrared wavelength range are available for single instruments. The photometer used in Wellington since 1975, with serial number 228, has three filter assemblies, designated blue, green and red, with nominal effective wavelengths of 0.44, 0.50 and 0.64 μm respectively. Each filter assembly also contains infrared blocking filters (Schott BG18, 1 mm) to obtain reasonably narrow bandwidths, quoted as 0.06 μm half widths (Volz, 1974). Each filter assembly is placed behind its own slit, in the front panel of a simple box construction. Each slit is alternatively aligned to the sun during observations, when the filtered sun-intensities are detected by silicon detectors suitably

positioned in the back of the box and their photovoltaic output read on a microammeter dial on top of the box.

Air mass values were calculated at Wellington from the times of observations. The effective field of view of the photometer slits is about 2^{o}.

The photometer filter calibrations and setting of the detectors are normally carried out by the instrument supplier, and calibration constants and the necessary corrections for filter temperatures or detector linearity are supplied with the instruments. Only the filter temperature corrections were necessary for the instruments used at Wellington and the original correction tables were used throughout the project.

(b) The basic equations

The basic equation of transmission of sunlight of wavelength λ through the atmosphere is presented in the form

$$J = J_o \cdot 10^{-M(\tau_R + \tau_0 + \tau_A)} = J_o \cdot 10^{-M\tau} \qquad (1)$$

where

J = observed instrumental response to radiation at λ;

J_o = instrumental response to extra-terrestrial radiation at λ at mean earth-sun distance;

τ_R = Rayleigh scattering coefficient at λ for unit airmass;

τ_0 = ozone absorption coefficient at λ for unit airmass;

τ_A = aerosol scattering coefficient at λ for unit airmass;

M = relative optical airmass;

τ = total extinction, including absorption and scattering at λ for unit airmass;

Absorption by other atmospheric constituents such as water vapour and carbon dioxide can be neglected for the wavelength ranges used in our Volz photometer.

In the application of this type of equation, corrections have to be made for actual earth-sun distance at the time of observation, for the different path-lengths of radiation through air, ozone and aerosol and for the changes of Rayleigh scattering with atmospheric pressure.

The Rayleigh scattering coefficients for the various wavelengths involved were taken from Volz (1959), together with the ozone absorption coefficients which were slightly revised to be applicable to an average total ozone amount of 0.300 cm rather than to the original 0.240 cm taken by Volz.

The values of the extra-terrestrial constants or instrumental constants J_o are normally supplied for each photometer. The difficulties of determining such instrumental constants are well known in radiometry and commonly, empirical methods using atmospheric rather than laboratory standard measurements are employed. From equation (1)

$$\log J = \log J_o - \tau . M$$

Accordingly, a set of atmospheric observations (J, M) made during days of horizontally homogeneous, constant optical density conditions can be used to determine graphically the value of J_o from the intercept and the value of τ from the slope of the straight lines determined by the above relationship. This is the well known "Langley plot" method which is theoretically simple to execute, but in practice its success depends on whether a sufficiently large number of days can be found for the calibration series, during which the atmospheric conditions specified above are closely fulfilled.

In recent literature some doubts were expressed concerning the value of turbidity data from the Volz instruments (Laulainen and Taylor, 1974). The large uncertainties in the turbidity data found by these investigators were however consequently traced back to the usage of the Langley plot method for obtaining J_o values under improper conditions (Russell and Shaw, 1975; Volz, 1975).

At the commencement of observations at Wellington, the instrumental constants for photometer number 228 were checked by a number of Langley plots from a carefully selected throughout-the-day series of measurements made during February to April 1975. The same procedure was carried out during January-February 1978. Some examples of Langley plots are given in Fig.1. The instrumental constants resulting from these plots showed little change from 1975 to 1978, but were less than the constants originally given by 28%, 35%, and 2% for the blue, green and red channels respectively.

These empirically established constants were used in the analysis of all data from observations with the number 228 photometer at Wellington.

(c) Calculations of the parameters of turbidity

It is useful to clarify the relation of the terminology given above to some other terms commonly used in the literature, and to describe the derivations of other turbidity parameters from the basic scattering or aerosol extinction

coefficients.

The often used "turbidity coefficient", B, is equivalent to the decadic aerosol scattering coefficient $\tilde{\tau}_A$ at wavelength 0.5 μm, i.e. as measured on the green channel of the Volz photometer.

This is also often called Schuepp's turbidity coefficient. Schuepp first used this wavelength and the base 10 for turbidity calculations (Schuepp, 1949) instead of the previously more commonly used indices of Linke or Angstrom.

Linke's turbidity factor T_L (Linke and Boda, 1922) relates the total atmospheric extinction to the Rayleigh extinction at all wavelengths and cannot be easily converted to other indices of turbidity as water vapor absorption has to be taken into account.

$$T_L = \frac{\log (I_o/I)}{M \tilde{\tau}_R}$$

where

I_o = solar constant;

I = measured total radiation;

$\tilde{\tau}_R$ = total Rayleigh scattering.

The turbidity factor used by Volz is expressed as

$$T = 1 + \frac{B}{\tau_R} \cdot \frac{p_o}{p} \simeq 1 + 16\, B\, \frac{p_o}{p}$$

where p_o = 1013 mb, p = surface pressure
This factor can be obtained from the green channel photometer observations and can be used in comparisons with turbidity factors derived from suitable pyrheliometric or other radiometric observations.

The Angstrom turbidity coefficient β is defined by the empirical formula

$$p_\lambda = \exp(-\beta / \lambda^\alpha)$$

where λ is wavelength, α is the "wavelength exponent" and p_λ is the aerosol scattering coefficient at wavelength λ (Angstrom, 1961). This relationship was empirically derived from many years of data given by the Smithsonian Institution and the Uppsala Solar Observatory transmission coefficient observations, which referred to the 1 μm wavelength (Angstrom, 1961).

The relationship between the Angstrom and the Schuepp turbidity coefficients can be found by taking into account

that both expressions must give the same amount of depletion of solar radiation, and the difference of wavelengths to which they refer, thus

$$e^{-\frac{\beta}{\lambda^{\alpha}}} = 10^{-\frac{B}{(2\lambda)^{\alpha}}} \quad \text{or} \quad \beta = 2.3 * 2^{-\alpha} * B \qquad (2)$$

It is seen that $\beta = B$ if $\alpha = 1.21$.

The value of the wavelength exponent α is related to the aerosol particle size distribution, with an average empirical value of about 1.3 (Angstrom, 1929; Mani et al., 1969). It can have large variations in individual conditions when particle size distributions depart markedly from the average, or when the observations are in error due to other causes such as cloud interference. Generally, values of α near 1 are thought to indicate a "classical" size distribution with small particle predominance in the submicron range, although the validity of talking about such typical size distributions has been disputed recently (Georgievskiy and Rozenberg 1963). Small or even negative values of α have been observed and thought to be due to the increased large particle aerosol fraction. Both the wavelength exponents, and the turbidity coefficients can have variations due to differences in the aerosol loading and distribution of air masses of different origin and age. Variations can also be caused by unusual natural events such as volcanic eruptions, fires and dust storms, and man-made pollution also will have a great influence.

The values of the wavelength exponent α at Wellington were calculated from the ratios of the scattering coefficients on two different channels, using the Junge type relationship of

$$\tau_{\lambda_1} / \tau_{\lambda_2} = (\lambda_1 / \lambda_2)^{-\alpha} \qquad (3)$$

For instrument No. 228, the exponent α was derived from the scattering coefficient ratios of the blue and red channels.

The Junge type exponential aerosol particle size distribution can be expressed as

$$dN(r)/d \log r = C\ r^{-V}$$

where N is the number of aerosol particles per unit volume with radii between r and r + dr, C is a constant proportional to the aerosol concentration, and V is the Junge exponent. This latter usually lies between 2 and 4(Junge,1963).

It can be shown that the relation with the Angstrom wavelength exponent is

$$\alpha = V-2$$

By making some simplifying assumptions with a value of $\alpha = 1$ and for aerosol radius range of 0.1 - 1.0 μm, the total aerosol number or mass loading of the atmosphere can also be calculated using this relationship (Volz, 1959).

However, when the Junge type exponential particle size distributions do not apply, these calculations cannot be used, but the wavelength dependence of the scattering coefficients can be derived from simultaneous observations on several wavelength channels of the photometer.

Another parameter, the height of the homogeneous aerosol atmosphere or aerosol scale height, can be calculated from the aerosol scattering in the vertical as measured by the Volz photometer, and the horizontal extinction, which is related to the meteorological range or visibility (Volz, 1959).

It can be shown, that the following approximate relation holds for a wavelength exponent $\alpha \simeq 1.5$, for a vertical aerosol distribution which is constant with height:

$$H \simeq 0.51 \times B \times S \text{ (Km)} \qquad (4)$$

where H is aerosol scale height, B is the Schuepp turbidity coefficient and S is horizontal visibility in Km.

The parameters of turbidity described above were obtained from the observations made in Wellington. Only the most significant of these parameters will be discussed in the following sections.

2. The variations of turbidity parameters in Wellington

Distributions during the year of average monthly mean turbidity parameters for the whole observation period and their standard deviations are given in Table 1. The number of observations each month of each year varied greatly and were generally small, ranging from one to ten, after doubtful data and observations obviously influenced by the presence of cloud were eliminated.

The wavelength exponents α given in the Table were derived according to relationship (3) from the ratios of the scattering coefficients observed on the blue and red channels. Using these derived α values, and the scattering coefficients obtained on the green channel (B), the Angstrom turbidity coefficients β were calculated from relation (2). Aerosol scale heights were calculated according to relation (4). Monthly mean values of these various

parameters are also seen in Figure 2.

All turbidity parameters show marked annual variations. There were phase shifts of the annual maxima-minima of the scattering coefficients between wavelengths, only the blue channel results showing the summer turbidity maximum usually found at other places (Angstrom, 1961; Mani et al., 1969; Flowers et al., 1969). The longer green and red wavelength turbidities show annual maxima occurring progressively earlier during the winter months, and very small turbidity values during the summer.

The large divergence in the summer half of the year between the blue and red wavelength scattering coefficients resulted in unrealistically large values of the wavelength exponents from November to December, and these are not given in Table 1. For the same reason the summer values of the Angstrom turbidity coefficient β were very small, or zero during this time. If the "classical" value of $\alpha = 1.3$ is used for all derivations of β from the green channel scattering coefficients, the resulting summer turbidity values become more acceptable as can be seen from Fig. 2(c). It is not unreasonable to assume that either the smallness of the sample, or an unknown instrumental defect affecting the red wavelength channel observations is responsible for the unsatisfactory absolute values of these results in the summer. The Jan-Sept. mean exponent values give $\alpha = 2$ which is readily comparable with the exponent values in the range of 0.9 to 1.5 obtained at Aspendale (Atmospheric Turbidity and Precipitation Chemistry Data for the World, 1973 and 1974).

It is not possible to say with certainty whether the differences in the phases of the annual maxima and minima at the different wavelengths represent real physical effects peculiar to the geophysical location of Wellington, or again are the result of sampling or calibration errors.

The values of the measured green scattering coefficients and the Angstrom turbidity coefficients were close to the range of minimum values observed in the United States, and never reached the comparatively high average annual values near 0.100 found there for the rural, reasonably unpolluted areas (Flowers et al., 1969). Similarly, the same order of magnitude for minimum turbidity was found at stations in India, while maximum values varied greatly during seasonal changes of air mass characteristics and at different geographical locations (Mani et al., 1969).

The monthly mean scattering coefficients at 0.5 μm at Aspendale, Australia, and the wavelength exponents calculated from two wavelength measurements at 0.38 and 0.5 μm are shown in Fig. 2(a). Values are higher at Aspendale than in Wellington except in June-July, when

they were about the same. Their annual variations were out of phase, as Aspendale turbidities were highest in the summer.

If these differences are real, they must be related to differences in air mass characteristics or differences in the "background" distribution of aerosols between Wellington and other places, such as Aspendale.

An attempt to classify turbidities according to different wind-directions at 900 mb, possibly characteristic of different air masses, resulted in the figures below.

Direction range	$\tilde{\tau}_{0.5}$	No.obs.	Stand.Dev.
E-S-W	0.021	99	0.018
W-N-E	0.015	35	0.020

The poor sampling qualities of the turbidity data and the likelihood that air mass categories cannot be separated by wind-direction categories alone, make these results inconclusive.

3. Comparisons between Volz photometers

During January-February 1978, nearly simultaneous observations with three Volz photometers were made at Wellington. Two of the three photometers (Nos.221 and 229) had the same construction, containing filters with the nominal wavelengths of 0.5 μm (green channel) and two infrared channels at 0.88 μm and at 0.94 μm. These two infrared channels can be used to obtain the amount of precipitable water from the ratio of their observations (Volz, 1974).

Although these latter have been calculated from the observations giving reasonable orders of magnitude, the instruments' calibration constants could not be checked, as required, by upper air humidity soundings at Wellington, and will not be further discussed here.

The third instrument taking part in the comparison was the photometer No.228 previously discussed in this paper. Thus there was only one common channel on the three instruments, the 0.5 μm green wavelength channel. During the comparison period, all photometers were calibrated by the Langley plot method to check and revise their original instrumental constants as described in Section 1(b). The original constants on instrument Nos. 221 and 229 had to be increased by 3% and 1% according to these atmospheric tests. These were rather small corrections compared to the large corrections that were found to be necessary at the blue and green wavelengths for instrument No. 228 (See Section 1(6)).

There were 14 days during January-February 1978 during which 31 separate comparisons were made. There was good correspondence between instruments 228 and 221 on the green channel, the average difference between calculated turbidities having been 0.001 or about 3% of the mean turbidity values for this period. Between instrument Nos 228 and 229 the correspondence was not so good, the mean turbidity difference was 0.012 or about 33% of the mean, with No 229 giving consistently higher turbidity values. As all the observed turbidity values were of the right order of magnitude, it is impossible to draw conclusions as to which instrument more nearly represented the real atmospheric values. If the instruments have not suffered filter deterioration or other mechanical defects during the three years' of use and storage at Kelburn, the possibility still exists that the instrumental constants have not been adequately determined (See Section 1 (6)), and still need revisions.

As would be expected, the wavelength exponent values derived from the 0.5 μm and 0.88 μm channel readings for instruments Nos 221 and 229 did not give similar values. Exponent values from observations throughout days when conditions were judged to be suitable for J_o determinations were however internally consistent for each instrument. This indicates that instrumental differences rather than calibration errors in the J_o values were responsible for the discrepancies.

4. Summary and recommendations

The results presented on the turbidity measurements with a three channel Volz photometer at Kelburn were reasonably consistent with other turbidity studies overseas, in spite of the non-routine nature and the limitations of the data. Measurements on one channel at 0.44 μm nominal wavelength have given annual variations with summer maxima as found elsewhere over the world, and the magnitudes of the monthly mean turbidities were close to those found at Aspendale, Australia. Readings on the longer wavelength channels however, resulted in winter maximum turbidities, and summer values were low, particularly at the red, 0.64 μm wavelength. In the summer, unrealistically high wavelength dependence of turbidity in this range resulted, and this was suspected to have originated from instrumental or sampling errors, rather than from real physical causes.

From a short comparison of three photometers, only two gave sufficiently close results, although each of the three was internally consistent throughout the period. Again, the causes of the discrepancies with the third photometer cannot be ascertained without rechecking all three instruments by laboratory and further atmospheric tests.

Since no other turbidity measurements are presently made and will probably not be made in New Zealand in the foreseeable future by other more sophisticated radiation instruments, the establishment of a network of well calibrated multi-channel Volz photometers at a number of selected meteorological stations should be considered, to provide a long series of aerosol scattering data needed for studies of atmospheric turbidity.

	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
τ Blue	0.085	0.057	0.074	0.050	0.057	0.035	0.044	0.065	0.055	0.060	0.066	0.076
S.D.	0.046	0.033	0.062	0.029	0.030	0.025	0.019	0.010	0.021	0.014	0.024	0.036
τ =B Green	0.021	0.012	0.018	0.011	0.024	0.019	0.028	0.039	0.030	0.026	0.026	0.018
S.D.	0.027	0.030	0.031	0.010	0.011	0.017	0.015	0.008	0.018	0.010	0.011	0.008
τ Red 64	0.003	0.005	0.039	0.027	0.048	0.024	0.026	0.026	0.027	0.005	0.002	-0.009
S.D.	0.022	0.020	0.039	0.020	0.020	0.010	0.016	0.013	0.020	0.007	0.006	0.011
No. years	3	3	3	3	3	2	3	2	3	2	2	2
No.Obs	11	22	19	22	12	9	18	7	18	7	8	4
α	2.9	3.4	1.8	2.2	0.2	1.0	1.7	2.2	2.7	5.8	-	-
β	0.000	0.002	0.020	0.011	0.042	0.020	0.033	0.028	0.018	0.002	-	-
H(km)	0.42	0.52	0.73	0.34	0.47	0.47	0.54	0.92	0.61	0.63	0.78	0.35

Table 1. Average monthly mean turbidity coefficients (τ_λ) and their standard deviations (S.D.) from observations at Wellington during 1975-1978, with Volz photometer No.228. Derived parameters are: Wavelength exponent α , Angstrom turbidity coefficient β , and aerosol scale height H.

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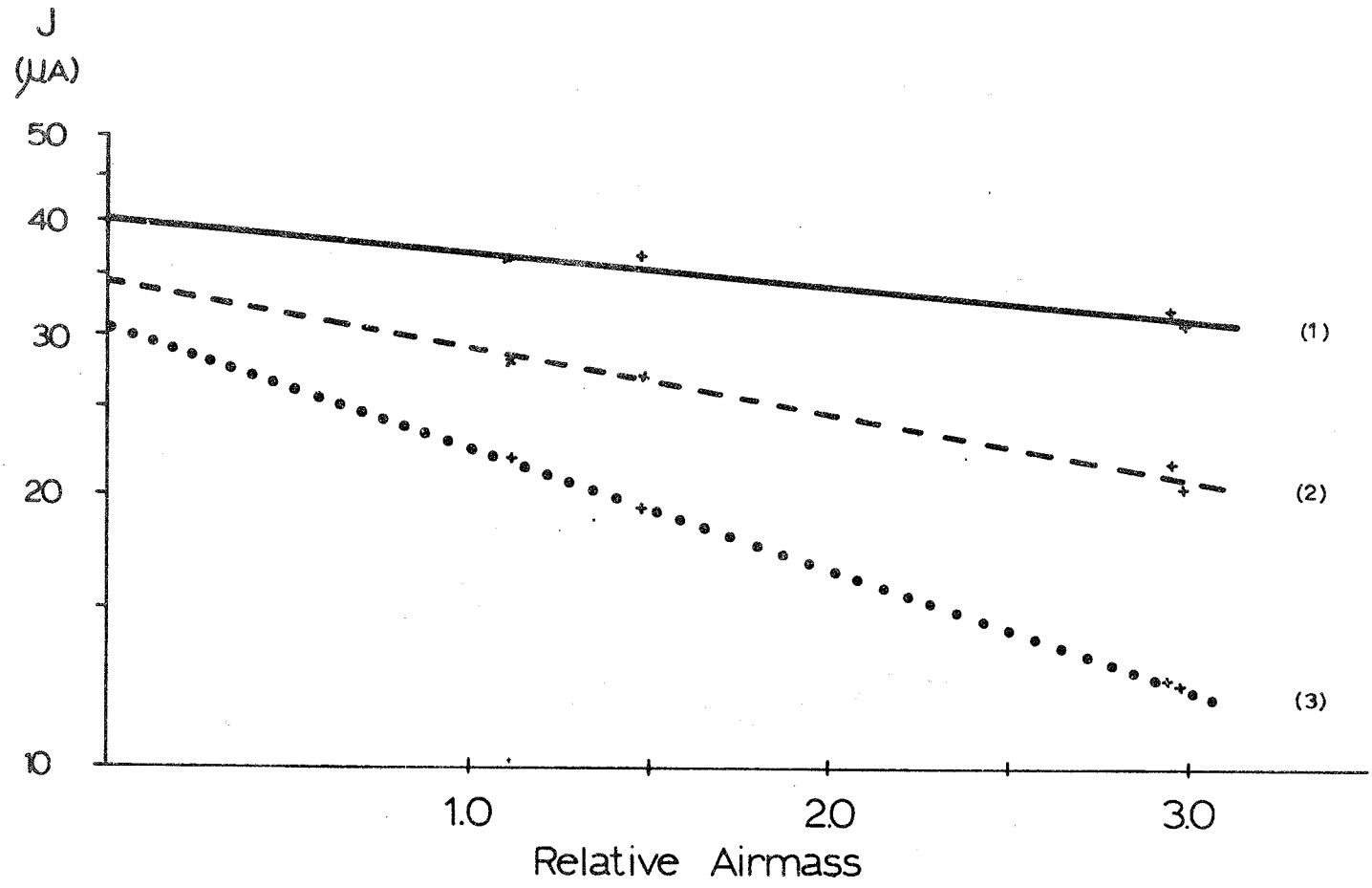

Fig. 1. Langley plot of Volz photometer (No.228) readings on the red(1), green (2) and blue(3) channels during a clear day at Wellington.

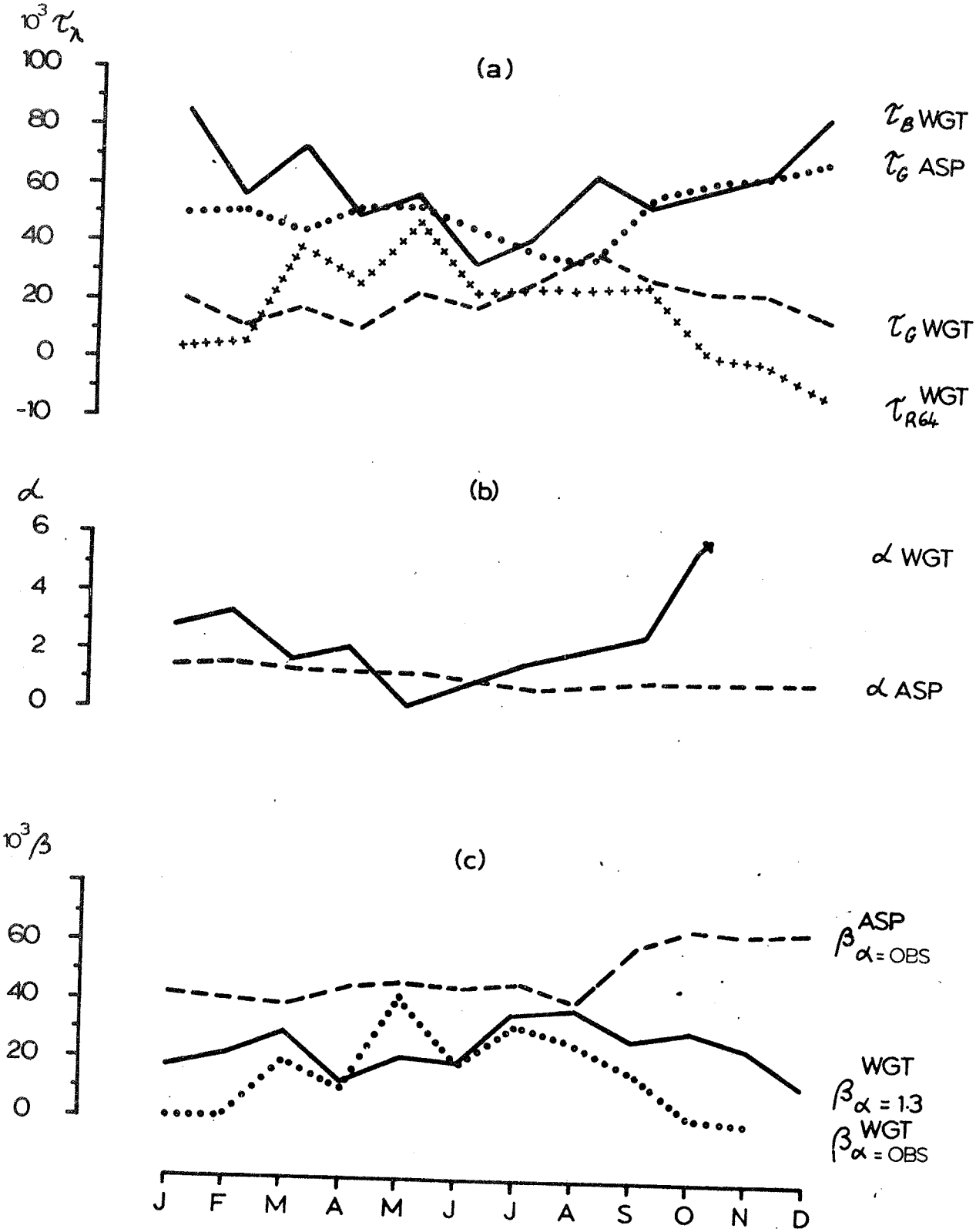

Fig. 2. Annual variations of:
(a) turbidity coefficients at Wellington for the blue(B), green (G) and red (R64) wavelengths during 1975-1978 and at Aspendale for the green wavelength for 1972-1973;
(b) corresponding wavelength exponents;
(c) Angstrom turbidity coefficients derived from τ_G and observed or constant values of the wavelength exponent α .