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MODIFICATION OF AN EMPIRICAL SHORT-WAVE RADIATION FORMULA USING SATELLITE OBSERVATIONS

by

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ABSTRACT

Recent estimates of the brightness and albedo of the earth-atmosphere system, averaged for individual months from meteorological-satellite data, are used to test and modify a formula for the short-wave radiation budget. The formula is used in a numerical model designed for long-range forecasting experiments.

Whereas previously the model formula assumed that the average transmissivity of the clouds is a slowly-varying function of latitude alone, the tests discussed here show that there is a linear inverse relation between cloudiness and transmissivity. This is due mainly to a strong tendency for the average thickness of the different cloud types, as well as for the relative frequency of the more vertically-developed types, to increase as the monthly-mean cloudiness increases.

This modification has an important effect on the heat budget of the earth atmosphere system, and is easy to introduce into the model.

Introduction

Indirect measurements of the clouds, the geopotential and kinematic structure, and the heat budget of the earth-atmosphere system from earth-orbiting satellites are becoming increasingly available. Within the foreseeable future it will be possible to obtain these factors over the globe in great detail from instruments such as the satellite infrared spectrometer (procedure described by Smith et al, 1970). These advances in measurement are matched by an enormous improvement in the global coverage and speed of collection.

It is necessary to accompany this improvement in observations of the environmental parameters with a corresponding improvement in equations for simulating them; because such equations are essential for generating the parameters in numerical models designed to forecast the general circulation and its weather.

It is the purpose of this report to summarize some work, bearing on this problem, which is part of a continuing effort to utilize satellite products in deriving fields of environmental parameters for long-range forecasting research. This particular report deals with an empirical formula for the short-wave budget of the atmosphere which has been in use for several years in a model designed for monthly and seasonal forecasting experiments (Adem, 1964a). This formula was tested using certain estimates of the albedo of the earth-atmosphere system (called hereafter the "top albedo") based on satellite measurements, and a simple but important modification was found. Although the principle of this modification has been known for many years, it has apparently not yet been used in numerical models.

A previous comparison of the formula with satellite observations of top albedo was carried out by Adem (1967).

Formula for top albedo

The method of handling the short-wave radiation budget in the thermodynamic model may be incorporated in the following formula for top albedo (see Adem, 1967, for a somewhat modified version):

$$\alpha_T = 1 - (a_2 + \varepsilon b_3) - [1 - \varepsilon(1-k)](1-\alpha)\frac{(Q+q)_0}{I} \qquad (1)$$

Here, α_T, ε and α are respectively the fractional top albedo, cloudiness and surface (land or ocean) albedo; $(Q+q)_0$ is the solar and sky radiation reaching the ground with a clear sky; "I" is the incoming solar radiation on a horizontal surface at the top of the atmosphere; a_2 and b_3 are respectively the fractions of incoming radiation absorbed in the air and within an overcast cloud deck; and k is the fractional reduction in radiation reaching the ground in the presence of clouds. The fraction k will be called for brevity the "cloud transmissivity".

Formula (1) was assembled (Adem, 1964a) from component parts which synthesize many statistical studies of short-wave radiation, based largely on surface observations, such as those summarized by London (1957) and Budyko (1956). For example, the last ~~item~~ term is the well-known Brunt-Ångström (or Savino-Ångström) formula for short-wave radiation absorbed at the earth's surface.

It should be noted that the formula has not been compiled by the direct method of adding up the radiation reflected or scattered to space from the atmosphere, clouds and ground; but rather indirectly, by subtracting from one the fraction of radiation absorbed by these factors. Therefore, the cloud reflectivity is not explicitly a part of the formula, but may be obtained approximately by setting the cloudiness equal to one.

As used in the model until recently, the coefficients a_2, b_3 and k were considered to be weak functions of latitude only, or of latitude and season. The short-wave radiation reaching the ground with clear sky $(Q+q)_0$ was determined from surface measurements, and the incoming radiation at the top of the atmosphere is a function of the solar constant and the mean daily solar zenith angle for different latitudes and seasons. Convenient tables of these quantities have been given by Adem (1964a, b).

Use of satellite brightness levels to estimate top albedo

The first estimates of top albedo used in testing (1) were based on macro-scale measurements of brightness derived by Taylor and Winston (1968) from digitized satellite video pictures (Bristor *et al*, 1966). Taylor and Winston recognized that in spite of ingenious efforts to develop a uniform scale of illumination, the digitized values still contain considerable variation. Therefore they reprocessed the data in an attempt to correct this problem. They combined the brightness values into larger 5-degree latitude-longitude "squares", and further adjusted the absolute values (on a scale from 0 to 10) so that certain selected control surfaces (cloud-free desert, snow-covered and oceanic areas) always had the same numerical reflectance. Their technique depends on considerable "hand" processing of the data, so that only a single 13-month period (February 1967 to February 1968) of fully-adjusted monthly and seasonal mean data are available for research. The resulting macro-scale digitized brightness values at each 5-degree "square" over the globe will be referred to here as "brightness levels". It is clear that this stabilization of the absolute reflectance over uniform surfaces means that the brightness levels are related to top albedo. Indeed, Winston later (1969) established a linear relation between these two quantities.

Brightness levels compared to computed top albedo

In order to compare the top albedo computed from (1) with the Taylor-Winston brightness levels, it was necessary to obtain mean values of cloudiness and surface albedo for the same months and 5-degree squares used by these authors. This was done for the contiguous United States (U.S.) only, where published values of the needed parameters are readily available for a large number of weather stations (ESSA, 1967-68). Twenty-five 5-degree squares were selected which lie mainly over land areas of the U. S. (Fig. 1).

Mean-monthly cloudiness was obtained for each "square" from the published station values, while surface albedo was obtained from the formula:

$$\alpha = \alpha_s f_s + \alpha_o (1 - f_s) \tag{2}$$

where α, α_s and α_o are respectively the fractional total surface albedo, albedo with and albedo without snow on the ground; and f_s is the frequency of one inch or more of snow on the ground as observed once each day during a given month. The "snow frequency" (f_s) was also obtained from the published records. The values of albedo with and without snow were extracted from charts of climatological surface albedo (Posey and Clapp, 1964) for the months January and July, respectively. It is assumed that the albedo with or without snow on the ground is independent of season.

With 25 5-degree squares and 13 months, there are 325 separate computations of top albedo to compare with the corresponding brightness levels. A plot of top albedo (expressed in percent) versus brightness level (times 10) is shown in Fig. 2. Because of the large number of cases, only 1/3 of the data points (every 3d point) can be shown in the figure. However, an "envelope" enclosing all but about 10 of all 325 points is shown by the closed dashed curve. Also, the points have been separated into 3 categories of surface albedo, as shown in the figure.

It will be noted that there is a fairly good relationship between the two factors. The well-known linear correlation coefficient has a value of 0.87. This correlation would be higher if the obvious curvilinear relation between the two quantities had been taken into account.

The large scatter of the points is of course due to errors in both quantities. With regard to the brightness levels, difficulties still remain in controlling their absolute values, but no doubt most of the scatter is due to the severe approximations involved in formulas (1) and (2), particularly with regard to the treatment of clouds. This is made clear in an excellent review of the pragmatic approach to short-wave radiation "climatonomy" by Lettau and Lettau (1969). Although as in the development of (1), many judicious compromises and assumptions were made in selecting critical coefficients, the treatment by the Lettaus is very thorough, and will be used in this report as part of the testing of (1).

Their study shows, for example, that the reflectivity and transmissivity of the clouds are highly variable, and depend critically on cloud type and thickness, while in (1) these factors are assumed to be almost constant over an area the size of the U. S. No doubt the surface albedo is also more variable than assumed here. It is the author's opinion that the random scatter in Fig. 2 is due mainly to these restrictive assumptions.

The separation of data points by magnitude of surface albedo (Fig. 2) together with reference to the individual computations (not shown), reveal that the formula correctly computes the brightest appearance of the earth-atmosphere system to occur with a combination of large cloud amount and extensive snow cover, while the darkest appearance occurs with low surface albedo (forested areas with no snow on the ground) and small cloud cover. However, when curves of "best fit"[1] are drawn through the data points, separately for each of the three surface-albedo categories (upper three curves AA, BB, CC in Fig. 3), it is seen that, for the same brightness level, there is a slight tendency for the formula to indicate increasingly higher top albedo as the surface albedo increases.

The reason for this separation is not clear, although it may be related to the way the formula handles multiple reflection between the ground and the clouds. It was decided to use the lower of the three curves (curve AA in Fig. 3) in the remainder of this study, because it represents about 80% of the area of the Northern Hemisphere, where low surface albedoes prevail. This curve has been extended beyond the data points (dashed portion) by drawing it more or less parallel to the other two curves.

Dependence of transmissivity on cloudiness

Let us next turn to a consideration of the assumed relationship between brightness level and the true top albedo, which has been obtained by Winston (1969) from a comparison between brightness levels and selected albedoes

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[1]Each curve of best fit was drawn by "eye" along the major axis of the "correlation ellipse" defined by each of the three sets of data points. It is not a regression curve.

measured with radiation sensors on aircraft and the NIMBUS II satellite (lower line EE in Fig. 3). This suggests that the top albedo from (1) is much too high, especially for small brightness levels (low cloudiness).

The report by Lettau and Lettau (1969) suggests that the large difference of 13% for low cloud amounts cannot easily be accounted for by systematic errors in computed absorption in the atmosphere or in radiation reaching the ground with clear sky. Formula (1) uses values of atmospheric absorption somewhat lower than those listed by the Lettaus, but this is partly compensated by higher formula values of radiation reaching the ground. The net result is the suggestion that the formula top albedo for clear sky may be about 5% too high, for the same surface albedo.

It is unlikely that an explanation can be sought in systematic errors of surface albedo; because, if anything, the values used in this study are too low, especially for snow-free surfaces. For example, the surface albedoes of Posey and Clapp (1964) at low latitudes in summer over the U.S. are systematically lower by (2 to 8 percent) than comparable estimates of Kung et al (1964). This suggests that the explanation must be sought in the radiative properties of clouds.

Significant errors in cloud absorptivity can immediately be ruled out, because, considering all cloud types and their relative frequencies, total cloud absorptivity averages out to be only 3 or 4 percent, even for a complete overcast.

Therefore it seems necessary to turn to possible errors in the cloud transmissivity (k) for an explanation of the systematic discrepancies. The average value of this quantity used in (1) over the U. S. is 33% (range, 32 to 34%), as compared to an average value for all cloud types of around 45% (Lettau and Lettau, 1969). If we accept the Lettaus' mean value of transmissivity, as well as the probability that the reflectivity of the clouds implied by the formula (about 60%) is too high (average value, 43% given by the Lettaus), this means that the computed top albedo is too high.

In light of these considerations, new values of k have been computed from (1) to "force" the top albedo to be lower, so as to agree with the assumed correct top albedo given by Winston's curve EE in Fig. 3. This has been done, using as given variables the cloudiness and surface albedo, for 18 of the 325 cases over the U. S., selected at random from among all cases having a surface albedo of 15 percent or less.

The most interesting result is the strikingly close relationship between the new values of k and cloudiness itself. This is shown by a plot of k versus cloudiness (large dots in Fig. 4). The curve of best fit for these points (curve AA) suggests that the transmissivity can exceed 1 for low cloud amounts; meaning that more solar and sky radiation reaches the ground under partly-cloudy conditions than with a clear sky. Such an event is extremely unlikely for a mean state averaged for a month and over large areas. Therefore, the relationship between k and cloudiness has been modified by simply extending the straight line (indicated by the data points for cloudiness greater than 45%)

down to low values of cloudiness; in such a way as to obey the constraints that k may not exceed 1 nor be less than 0. The equation for that straight line is shown as formula (3).

$$k = 1 - 0.568\,\mathcal{E} \tag{3}$$

This modification forces a corresponding change in line EE of Fig. 3, as shown by the dot-dashed extension, E'. It must be stressed that this does not imply that the extended curve E' gives a more correct relation between brightness level and true top albedo, but is used here to obtain a logical relation between cloudiness and transmissivity. In fact, taking account of the usual finding that surface-observed cloudiness is too high for small cloud amounts, it can be shown that lowering the cloud amounts by ten hundredths or less will cause the last 3 of the 18 data points to fall near the line AA' of Fig. 4, and therefore will leave unchanged the line EE of Fig. 3.

This inverse relation between transmissivity and cloudiness has been known in principle for many years, and is revealed by a non-linear relationship between total cloudiness and the ratio of incoming radiation reaching the ground with cloudy as compared to clear skies (e.g. see Fig. 14 from Fritz, 1955). It is due mainly to the observed fact that (at least in middle latitudes) both the frequency and the thickness of the more opaque types of clouds increase (in a statistical sense) as cloudiness increases; i.e., thin cirrus and stratus clouds tend to give way to thickening nimbo-stratus and/or towering cumulus types as cloudiness increases. This systematic dependence of some of the important properties of the vertical structure of clouds on a single parameter (total cloudiness) is of obvious utility in developing numerical models, where practical considerations make it essential to maintain optimum simplicity of the governing equations.

Fig. 14 of Fritz (1955) shows the transmissivity ratio as a function of cloudiness for 3 different localities. This ratio is essentially an expression of the factor $\left[1 - \mathcal{E}(1-K)\right]$ in the last term of formula (1). Therefore, the transmissivity was re-computed by equating this factor to the average of the 3 values of the ratio for a large range of cloudiness. The resulting smoothed relation between cloudiness and transmissivity is shown by the solid curve in Fig. 5. The straight line AA' (dashed) has been transposed from Fig. 4 for comparison. It is interesting that the lowest transmissivity (41%) indicated by the new curve at 100% cloudiness is almost exactly the same as that for the line AA' (43%). Furthermore, the transmissivity also increases as cloudiness decreases, but at a much more rapid rate, and the relationship is strongly non-linear. The reason for these differences is probably related to the fact that Fritz' data are based on individual days and locations, whereas the data of this study deal with monthly averages ranging over a large and climatologically inhomogeneous continent.

Independent test of formulas for Top albedo

In order to make an independent test of the modified radiation formulas, formula (3) was used to define k as a function of cloudiness; curve EE' (Fig. 3), to determine top albedo from brightness level; and these were combined with (1) to compute cloudiness for 26 months and locations outside the U. S. (where the dependent sample was chosen). These areas were selected mainly over the oceans, where surface albedo is known within reasonable limits; although 3 continental points and 1 island location were also selected.

The computed cloudiness is listed in column 4 of Table 1. The corresponding brightness levels, read from the charts of Taylor and Winston, are listed in column 7.

The observed cloudiness (column 5) was obtained from unpublished data over the Northern Hemisphere, north of about 15°N, provided by the Air Force Environmental Technical Applications Center (ETAC). The ETAC cloudiness represents the mean of twice-daily (00 and 12 GMT) values, computed at each intersection of a cartesian grid of 1977 points on a polar -stereographic projection. Each grid-point value is the average of all available cloud observations made at surface ships or land stations within a single grid area of about 3° latitude on a side at 40°N; a somewhat smaller area than that used for the mean brightness levels. When no observations were available within a reasonable distance of a gridpoint, the mean cloudiness was omitted for that synoptic time, so that a monthly average may be made up from less than 60 values (for a 30-day month).

With the few exceptions discussed below, no attempt has been made to question the probable reliability of the ETAC cloudiness. The values listed in Table 1 were selected from the gridpoints closest to the indicated geographical locations. The algebraic error in computed cloudiness is listed in column 6.

It can be seen from a comparison of columns 4 and 5, or from column 6, that there is a good relation between computed and observed cloudiness. If an error of more than 10% is considered large compared to the limits of accuracy of observed monthly-mean cloudiness, then only 7 of the 26 cases are grossly in error. The largest error is for a point located over the Sahara Desert at 25°N, 25°E. This illustrates the extreme sensitivity of formulas (1) and (3) to surface albedo when cloudiness is small. The top albedo obtained from the formulas with the observed cloudiness of 2% and a surface albedo of 30% (Posey and Clapp, 1964) is 28%, while the top albedo from the indicated brightness level and curve EE of Fig. 3 is 32%. To make up this difference of only 4% in albedo it is necessary to increase the cloudiness to 44%!

The point at 21°N, 157°W was chosen to coincide with the average position of 4 weather stations in the Hawaiian Islands, while that at 52 3/4°N, 35 1/2°W coincides with Ocean Station Vessel "C" (Ship Charlie). It is interesting to note that while the mean-monthly cloudiness at ship Charlie and the other U. S. weather ships in both oceans (unpublished data furnished by ESSA's National Climatic Center) agree very well with the ETAC cloudiness, the cloudiness at 4 Hawaiian stations (ESSA, 1968) average 12% higher than the ETAC values. This agrees with the well-known finding that island weather stations tend to have unrepresentatively high cloudiness as compared to the surrounding oceans. For this reason, plans were abandoned to test the formulas using available cloudiness from island stations near the equator.

The results have been summarized quantitatively in Table 2, where the mean error of the computed cloudiness has been obtained for each of three class intervals of the observed values. The mean absolute error (column 4) shows that the computed cloudiness is correct within ± 15%, with the error decreasing sharply with increasing cloudiness. The mean algebraic error (column 3) shows that there is a systematic error (bias), so that the computed cloudiness tends to be too large for low and too small for intermediate and high values, although the bias remains within 15%.

Comparison with recently-computed ~~planetary~~ top albedoes from NIMBUS II

Raschke and Bandeen (1970) have recently re-determined ~~planetary~~ top albedoes from the short-wave radiation sensors of NIMBUS II. Their calculations include careful corrections for the complex anisotropic dependence of reflected radiation on the zenith angle of the sun and on the zenith and azimuth angles of the recording instrument. Their estimates should therefore be reasonably accurate. Their values of zonally-averaged ~~planetary~~ top albedo for each 5° latitude in the Northern Hemisphere, and for the month of July 1966, are shown in the last column of Table 3. These were obtained from an unpublished table furnished by them, which was used to construct their Fig. 14.

In order to compare these with computations using formulas (1) and (3), it was at first assumed that the non-linear term in (1) can be evaluated using zonally-averaged values of cloudiness, cloud transmissivity and surface albedo.

The surface albedoes are again set equal to their climatological values as determined by Posey and Clapp (1964); and the zonally-averaged values from their Northern Hemisphere charts are listed in column 2 of Table 3. There is no reason to expect large anomalies of zonally-averaged surface albedo in Summer, except possibly north of 60°N.

Mean cloudiness for July 1966 was available only for latitudes 5, 15, and 25°N (Sadler, 1969, shown in column 4 of the table). Therefore, in order to obtain a more complete comparison with the top ~~planetary~~ albedoes of Raschke and Bandeen, it was decided to use cloudiness at these and other latitudes for the summer of 1962 (Clapp, 1964, shown in column 3). Both sets of cloudiness values are based on satellite nephanalyses.

Because of year-to-year changes in the general circulation, it is quite possible that the mean cloudiness for the summer of 1962 may have differed significantly from that of July 1966, leading to large errors in the computed albedoes. However, it will be noted that the cloudiness values at the three lower latitudes agree within a few percent.

Top ~~Planetary~~ albedoes based on formulas (1) and (3) using the summer 1962 and July 1966 cloudiness values are listed in Table 3, columns 5 and 6 respectively. It will be noted that there is good agreement in the two sets of computed albedoes for the three lowest latitudes. Comparison with the satellite-observed top albedo shows that the computed values are quite good, although they tend to be somewhat too low, with maximum errors of -10 to -12 percent at latitudes 25, 35 and 45°N.

The larger under-estimates at low latitudes are due to the presence of a sharp minimum in computed albedo, corresponding to a weak minimum (at 15°N) in Raschke and Bandeen's albedoes. This lack of a distinct minimum is somewhat unexpected in view of the fact that both Sadler's and Clapp's cloudiness values show a sharp minimum around latitude 25°N (see Fig. 12 of Clapp, 1964; and Fig. 4 of Sadler, 1969). Since most of the hemisphere at low latitudes consists of open water, it seems that the planetary albedo should be very responsive to changes in cloudiness.

It is likely that this discrepancy can be accounted for by the non-linearity of the last term in formula (1). Special calculations at latitude 25°N show that the insertion in this term of zonally-averaged values of the parameters results in too low top albedoes due to the presence of a negative correlation between cloudiness and surface albedo, and especially because of the inverse relation between cloudiness and transmissivity given by formula (3). These factors will produce a maximum discrepancy along 25°N where cloudiness has its maximum variability and tends to be much higher over the dark oceans than over the bright deserts. Indeed, the top albedoes of Raschke and Bandeen over desert areas (maximum, 43% shown in their Fig. 8) suggest that the surface albedo there is considerably higher than that given by Posey and Clapp (maximum, 30%), because in regions with few clouds the top albedo must be close to its surface value.

The conclusion to be drawn from this and the preceeding section is that the modified formulas (1) and (3), combined with curve EE' (Fig. 3), give a reasonably good representation of the short-wave radiation budget of the atmosphere.

Conclusions

A single example has been presented here of the value of satellite data in modifying equations used in numerical weather prediction; in this case an empirical formula for simulating the short-wave radiation budget of the earth-atmosphere system. Even though the principle of the modification (an inverse relationship between the transmissivity of the vertical cloud structure and total cloud amount) has been known in principle for a long time, the ease with which it may be applied in numerical models, and the profound effect it has on the heat budget, are perhaps not so well known.

To give some idea of the importance of this correction, it may be pointed out that it leads at all values of cloudiness, to an increase in the calculated absorption of solar energy at the earth's surface, reaching a maximum increase of 35% at 8/10 cloudiness. This substantial increase is available for direct heating of the atmosphere through turbulent transfer or long-wave radiative exchange, and/or for evaporation of moisture.

Vonder Haar and Suomi (1969) have suggested that the net absorptivity of the earth-atmosphere system in the tropics may be considerably higher than previously calculated. They attribute this to earlier overestimates of opaque cloudiness. It is suggested here that a lower than average cloud albedo in the tropics, where total cloudiness is low over the oceans, may also be a contributing factor.

There is clearly room for improvement in the formulas and graphs presented here. For example, the too low absorptivity in the atmosphere may be due to the omission of ozone absorption, which may be corrected in the manner suggested by Adem (1967). Also, a correction can be made to allow for the increase of transmissivity with the surface albedo, which is due to multiple reflections between cloud and ground, as pointed out by Fritz (1955). Finally, it is necessary to point to the need for further improvement in the accuracy of the satellite albedo measurements. The latter seem to have had a disturbing tendency to increase with each new study: e.g. the albedoes of Raschke and Bandeen (1970) are locally considerably higher than those of Winston (1969).

Other important aspects of the energy budget can also be inferred from the satellite data, such as the distribution of the heat of condensation, through precipitation, (Lethbridge, 1967) which is perhaps the most important component of the heat budget of the atmosphere. A study of the relation between precipitation and top albedo was made as part of the project discussed here, and will be reported elsewhere.

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Legends to Figures

Fig. 1. Five-degree latitude-longitude "squares" used in computing mean environmental paramenters.

Fig. 2. Plot of top albedo (percent) computed from formula (1) against brightness level (times 10) for 1/3 of all 325 cases. Open circles are for cases with surface albedo 15 percent or less; dots, 16 to 29; crosses, 30 or more. Dashed closed curve is envelope enclosing 97 percent of all cases. Horizontal and vertical lines are drawn through averages of top albedo and brightness level.

Fig. 3. Composite of several curves relating top albedo and brightness level: AA, BB and CC are curves of best fit corresponding to computed albedoes of test sample shown in Fig. 2; drawn respectively through cases with surface albedo 15 percent or less, 16 to 29 percent and 30 percent or greater. DD is line of perfect agreement, shown only for convenience in orientation; EE, assumed relationship between brightness level and true top albedo (after Winston, 1969), with adjusted extension, E'.

Fig. 4. Adjusted cloud transmissivity related to cloudiness (both in percent). AA is curve of best fit corresponding to curve EE of Fig. 3 and drawn through selected data sample (dots). AA' is straight-line extension of AA and corresponds to EE' of Fig. 3.

Fig. 5. Cloud transmissivity related to cloudiness, both in percent. Solid curve is smoothed fit to data (dots) derived from Fritz (1955). Dashed line is line AA' transposed from Fig. 4.

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Tables

Table 1. Cloudiness computed by formulas (1) and (3) (col. 4) compared to observed cloudiness (col. 5) for 26 cases for the months February to October, 1967. Computed and observed cloudiness, and algebraic error (col. 6=col. 4 minus col. 5) are in percent. Brightness levels (col. 7) are 10 times values from Taylor and Winston (1968).

Table 2. Error in cloudiness computed by formulas (1) and (3) for 3 class intervals of observed cloudiness. Data from Table 1. Average observed cloudiness and its range in each class interval (cols. 1 and 2) are in percent. Average algebraic (computed minus observed) and absolute (sign disregarded) errors also in percent.

Table 3. Zonally-averaged top albedoes computed from formulas (1) and (3) compared to those from data of NIMBUS II, July 1966. Surface albedoes from Posey and Clapp (1964) in col. 2. Observed cloudiness in col. 3(for summer 1962 from Clapp, 1964) and in col. 4 for July 1966 (from Sadler, 1969) were used in computations of top albadoes shown in cols. 5 and 6 respectively. Observed top albedoes from NIMBUS II (Raschke and Bandeen, 1970) are in col. 7. All cloudiness and albedo values in percent.

TABLE 1

1 North Lat.	2 Long.	3 1967 Month	4 Comp. Cldns.	5 Obs. Cldns.	6 Alg. Error	7 Bright Level
15	110W	Feb.	45	54	-9	10
45	180	"	77	75	2	60
45	40W	"	89	84	5	73
20	180	"	48	66	-18	20
50	40W	Mar.	89	86	3	71
15	165E	Apr.	44	50	-6	18
45	180	"	85	83	2	63
15	20W	"	46	19	27	3
55	30W	"	80	84	-4	61
25	160W	July	55	55	0	9
25	120W	"	78	88	-10	42
25	60W	"	56	39	17	5
25	25E	"	44	2	42	44
25	100E	"	83	88	-5	51
45	0	Sept.	59	59	0	36
55	180	Oct.	62	71	-9	43
25	120W	"	52	58	-6	18
55	30W	"	46	81	-35	32
21	157W	Feb.	49	54	-5	21
"	"	Apr.	57	50	7	28
"	"	July	54	50	4	12
"	"	Oct.	47	46	1	17
52 3/4	35 1/2W	Feb.	76	83	-7	67
"	"	Apr.	82	94	-12	62
"	"	July	81	90	-9	54
"	"	Oct.	56	78	-22	38

TABLE 2

1	2	3	4	5
OBS. CLDNS.		ERROR IN Computed Cldns.		NUMBER CASES
AVE.	RANGE	ALGEB.	ABSOL.	
39	2 - 53	+13	15	7
68	54 - 82	-10	11	10
87	83 - 94	-4	6	9
65	ALL	-2	10	26

TABLE 3

1	2	3	4	5	6	7
Latitude	α	ε_c	ε_s	α_{TC}	α_{TS}	α_{TR}
5°N	7	64	59	27	25	27
15	8	50	48	19	19	26
25	11	41	44	17	18	27
35	9	45		17		29
45	12	58		26		36
55	13	68		34		38
65	19	74		40		41
75	30	80		50		52

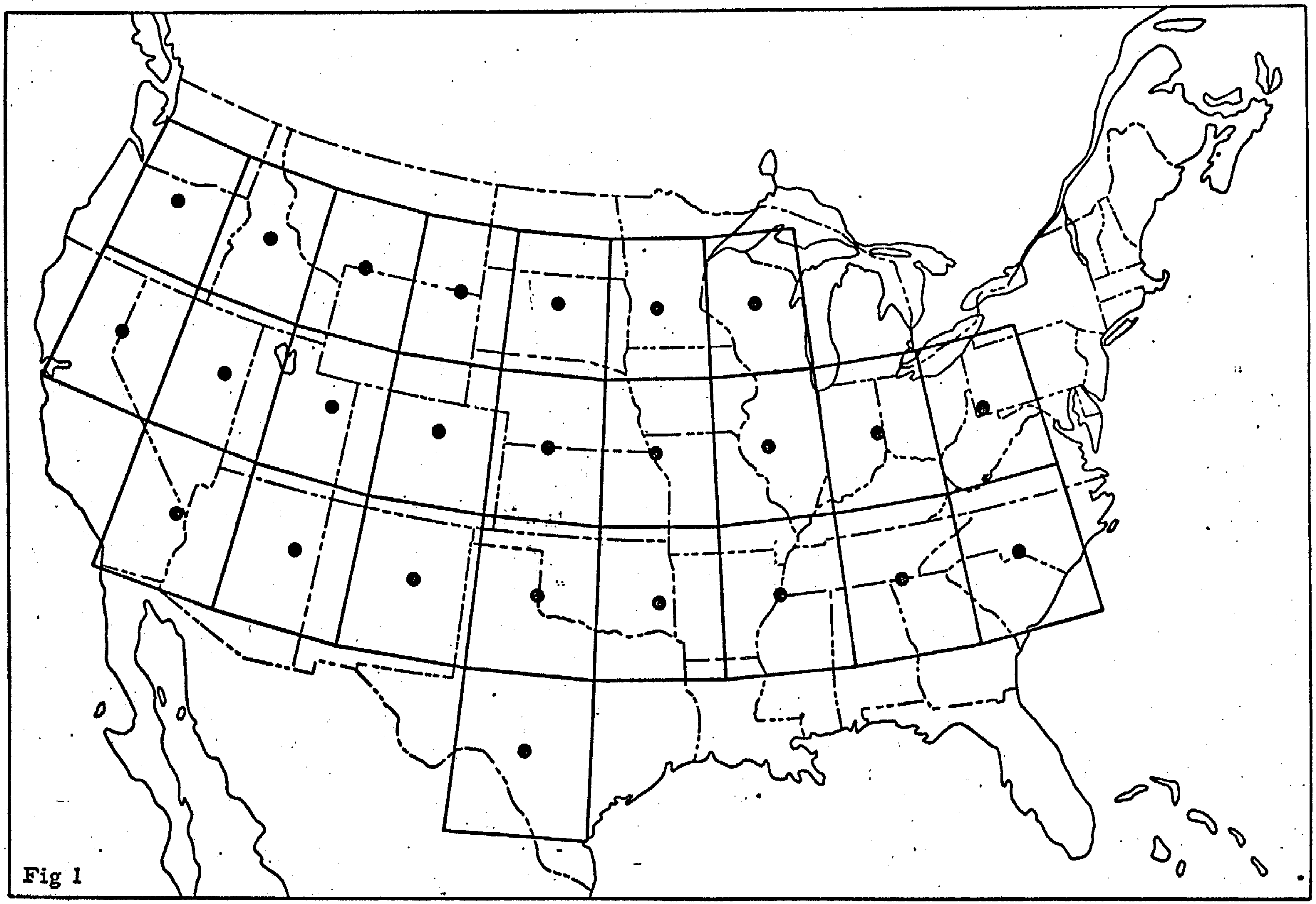
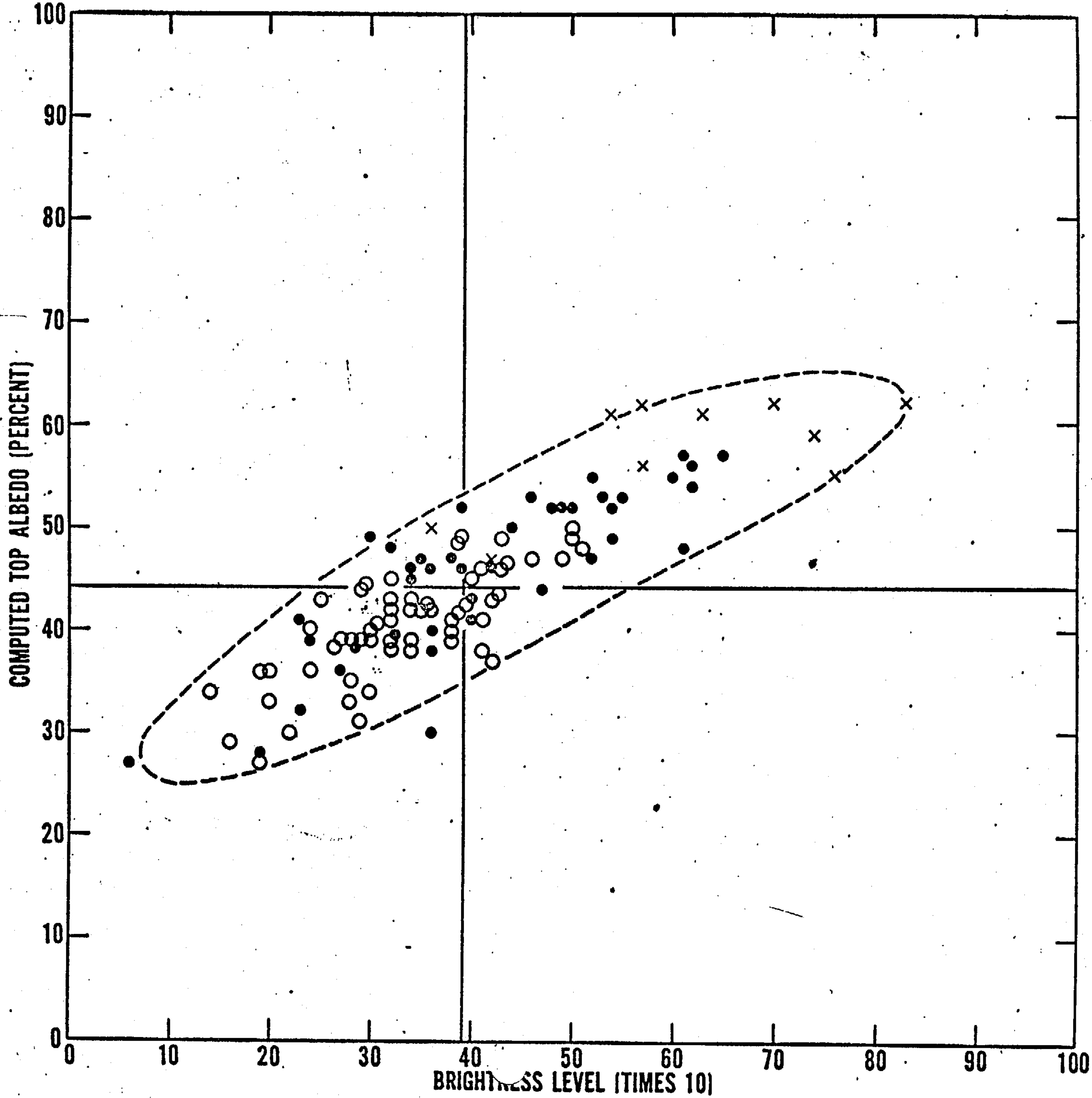
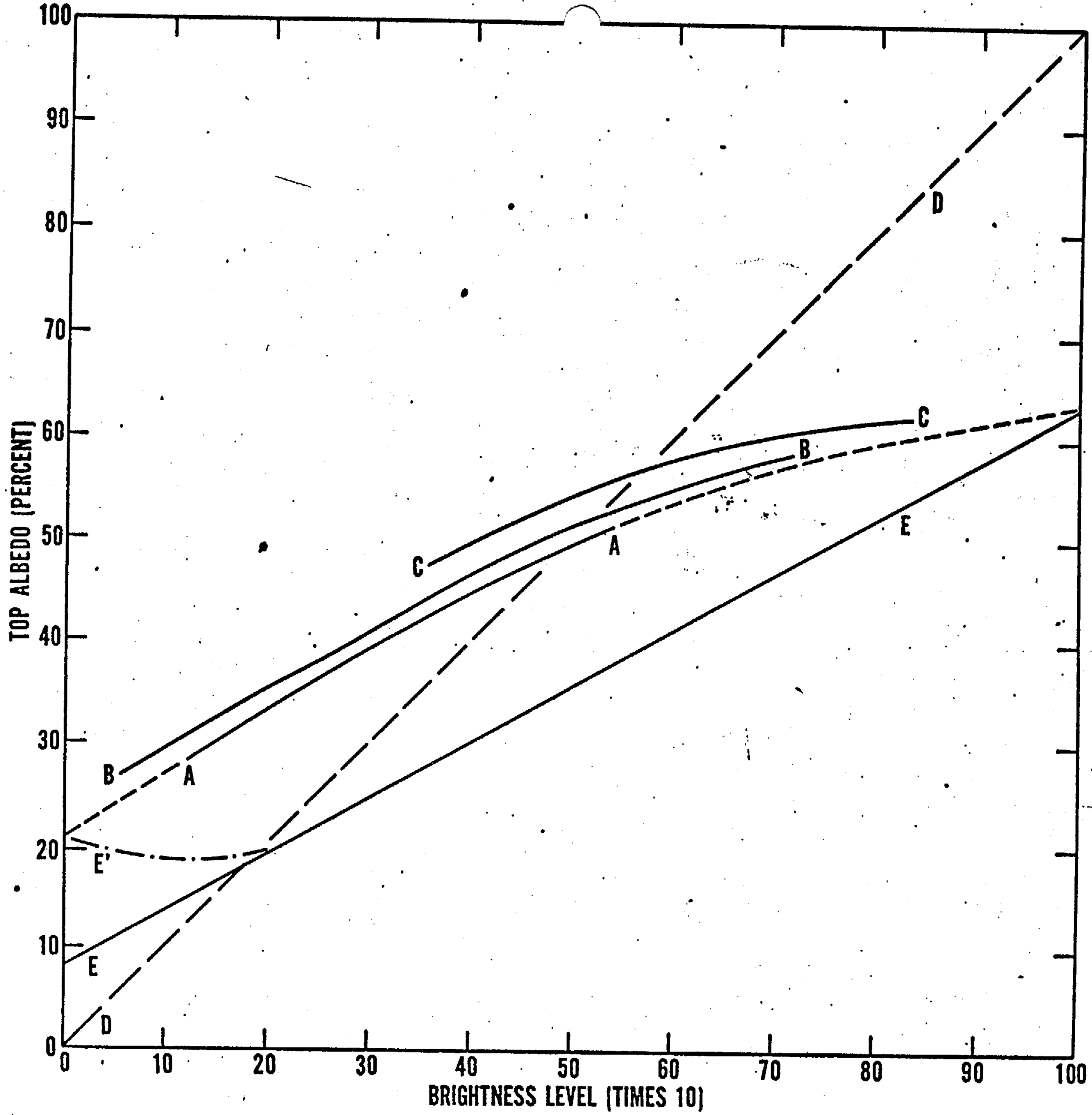


Fig 1

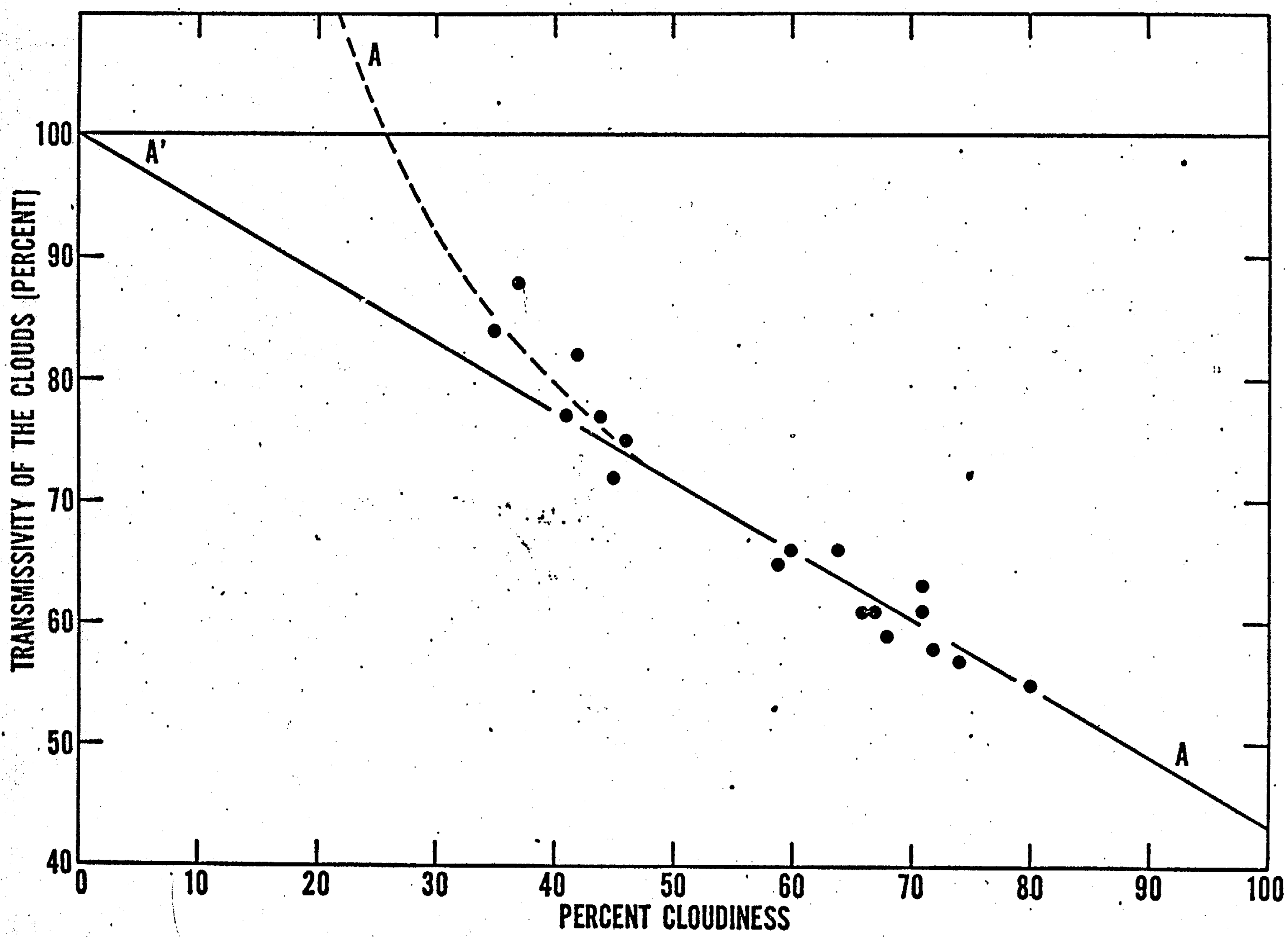

COMPUTED TOP ALBEDO (PERCENT)
BRIGHTNESS LEVEL (TIMES 10)
0
10
20
30
40
50
60
70
80
90
100

Fi



TOP ALBEDO (PERCENT)
BRIGHTNESS LEVEL (TIMES 10)
0
10
20
30
40
50
60
70
80
90
100
A
B
C
D
E
E'

Fig 3



A
A'
A
100
90
80
70
60
50
40
0
10
20
30
40
50
60
70
80
90
100
TRANSMISSIVITY OF THE CLOUDS (PERCENT)
PERCENT CLOUDINESS

Fig 4

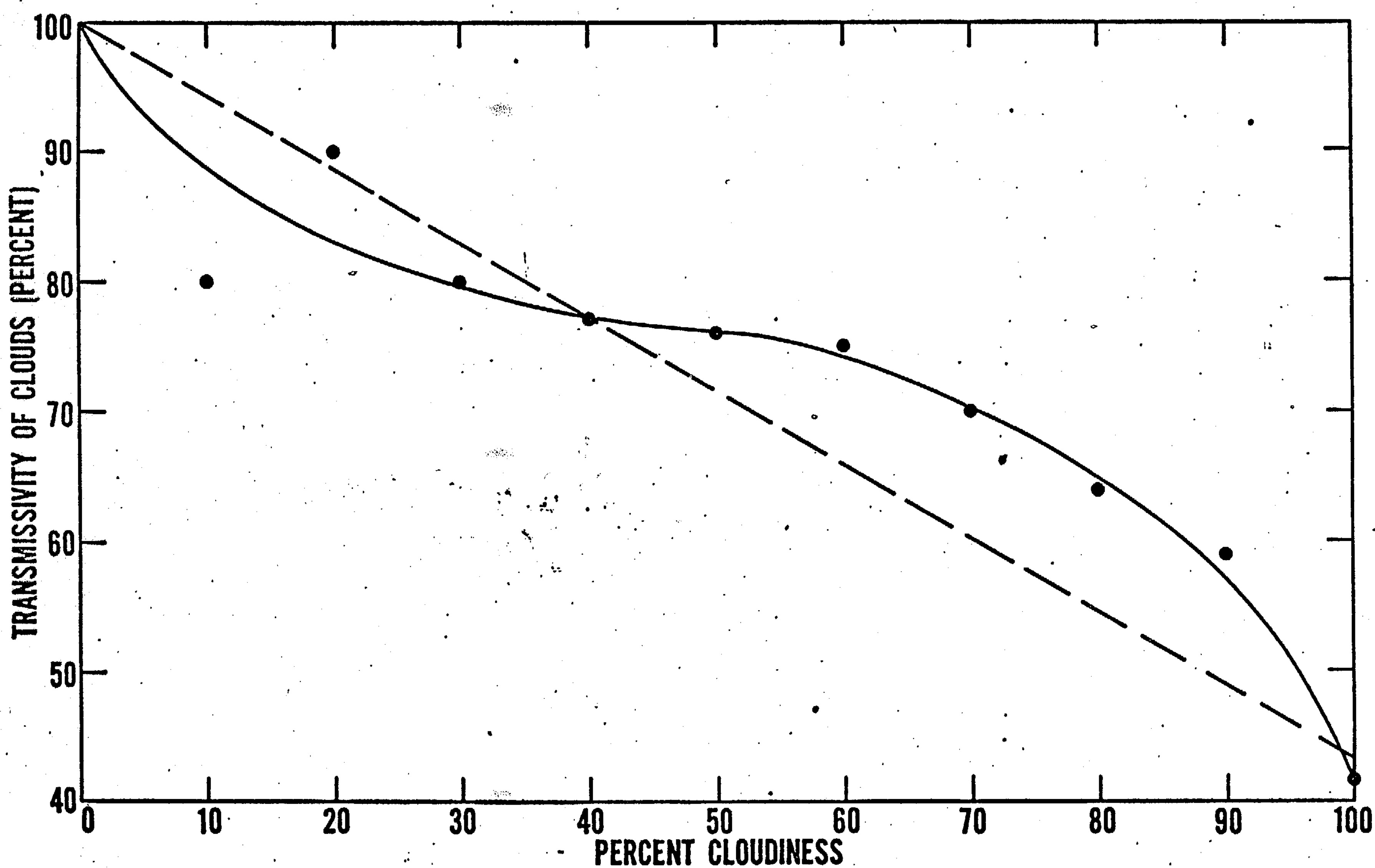

100
90
80
70
60
50
40
TRANSMISSIVITY OF CLOUDS (PERCENT)
0
10
20
30
40
50
60
70
80
90
100
PERCENT CLOUDINESS

Fig 5