Shortwave radiation climatonomy

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ABSTRACT

Functional relations are discussed which express absorbing and scattering processes of shortwave radiation in the atmosphere. A simplified model is derived in which only overall intensities (integrated over the spectral band of insolation) are considered. Balance requirements are established which relate "top albedo" to effective absorption in the air and by the ground, both as fractions of extra-atmospheric irradiance. Measurements of diffuse radiation from the sky, in addition to global radiation, are used to specify scattering versus absorption components of aerosol attenuation for clear-sky conditions with known liquid water content and Rayleigh scattering. Data from three continental locations are selected which are representative of clear-sky attenuation over a desert (La Joya, Peru), an open prairie (O'Neill, Nebraska), and an industrialized urban area (Kew, England). The ratio of absorption to scattering shows significant systematic variation, city aerosol being more efficient as absorber, desert aerosol more as a scatterer. Effects are calculated which result from a systematic variation of (1) surface albedo, (2) liquid water content, (3) aerosol, and (4) cloud cover. Among other results it is demonstrated how top albedo (measurable nowadays from orbiting or synchronous satellites) will react to changes in surface albedo and aerosol scattering. It is also shown numerically, how cloud effects on global radiation are modified by the existing turbidity conditions in the cloudless air.

1. Introduction

As far as we know the word "climatonomy"[2] was first used by H. Lettau at the meeting of the American Geophysical Union in Washington D.C. in 1954. It was suggested in logical expansion of thoughts expressed in a technical study entitled "Synthetische Klimatologie" (1952). In "Review of Climatology, 1951–1955" H. Landsberg gave recognition to the term and suggested that "climatonomy" could encompass theoretical, energetic, and circulation climatology. During the last decade a new definition has evolved. Now, "climatonomy" shall be used for the mathematical explanation of the basic elements which determine the physical environment at any planetary surface. This includes the specific climate represented by mean-value and temporal-spatial variations of temperature in the lower atmosphere as well as the upper strata of the underlying lithosphere or hydrosphere.

Mathematical explanation implies the use of numerical models which yield theoretical solutions in the form of a precise "response function", or output, in physical relationship to a precise "forcing function", or input. For any planet the intensity of the forcing function is determined by absolute distance to the central sun, and its time structure by orbital elements and spin or the varying relative positioning of the considered planetary surface section with respect to the sun's energy flux-density. The climatonomic response function, governed by fundamental physical laws of radiation, heat conduction, and universal conservation principles, will be investigated in another paper. The following discussion will be restricted to problems of the forcing function in terrestrial climatonomy which require the

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[2] The suffix "nomy", related to the Greek word "nomos", has among other meanings that of "law"; while "logy", derived from "logos", has also several meanings, one of which is "word". Von Ficker argued more than forty years ago that the time has come to proceed from meteorology to "meteoronomy" in order to express increasing reliance on numerical methods at the expense of descriptive methods. More than a decade ago, similar reasoning by S. Chapman resulted in the acceptance of the term "aeronomy" for the mathematical study of physics of the upper atmosphere, restricting "aerology" to a more descriptive approach.

parameterization of shortwave radiation flux-density between the top and the bottom of the atmosphere, and its diurnal, seasonal, and multi-annual variation at a given locality.

One of the objectives of climatonomical parameterization is to investigate the possibilities of climate modification and control. It is important to know, for example, how the insolation at a given locality will be affected if one specific climatic parameter were doubled or halved, or changed otherwise in a physically realistic way; such parameter could be cloudiness, amount of precipitable water, aerosol content, surface albedo, etc. This problem can only be solved by budgetary considerations of the complex attenuation processes in the atmospheric column above the region.

Inspection of the international literature shows that the problems of measurement, data interpretation, and climatological analysis of shortwave energy continues to attract attention of investigators in many countries. During the last decade, several world-wide studies and atlasses were published, such as the monographs by Budyko (1956), Bernhardt & Philipps (1958), Ångström (1962), and several others. Interest in these problems was stimulated when radiation data (including measurements of "earthshine" in very satisfactory resolution both geographically and temporally) became available from orbiting satellite stations; see Suomi (1958), Fritz, Rao & Weinstein (1964), Hanson (1967), Vonder Haar (1968), and others.

Why do we want to add another to the numerous existing and valuable discussions? Because we feel that it is necessary to emphasize the usefulness of a straightforward and reasonably complete yet still elementary budgetary-type method. The quantitative appraisal of attenuation processes in an atmospheric column within the framework set by upper and lower boundary conditions, appears to be most promising for fuller understanding and prediction of how the local radiation-climate will be affected by natural or man-made changes of the environment. Indeed, manipulation of factors which are an integral part of the local shortwave-radiation budget, such as albedo changes by either brightening or darkening of the ground, by shading, smoke-screening, etc., are among the oldest techniques for climate and weather control. In climatonomy, total energies are more important than spectral distributions. Therefore we restrict the discussions to values integrated with respect to wavelength over the band-width of the solar spectrum.

2. Mathematical relations

It is convenient to make the mathematical relationships dimensionless by dividing any energy-flux density by the extra-atmospheric irradiance, $I_0 \cos \vartheta$, (ϑ = zenith angle of the sun). To guarantee a reasonable degree of completeness and detail, nine variables must be considered. These can be divided into three groups:

2.1. *Non-dimensional values of shortwave fluxes*

A = top albedo, or fraction returned to space,
G^* = global radiation, or fraction received at ground level from sun and sky,
D^* = diffuse radiation, or fraction received at ground level from sky alone,
H^* = solar heating in the air, or fraction absorbed in the atmosphere,
a = energy albedo of the lower boundary; or, $(1-a)G^*$ = fraction absorbed by the lower boundary.

2.2. *Contributions to the attenuation of non-dimensional beam radiation*

σ = part of attenuation due to scatterers in the air,
α = part of attenuation due to absorbers in the air.

2.3. *Coefficients of the scattering process*

μ = fraction describing effective outward scattering to space; or $(1-\mu)$ = fraction of effective downward scattering to the lower boundary,
$\varkappa$ = fraction describing backward scattering of that part of shortwave radiation which has been reflected at least once by the lower boundary.

The physical relationships between the nine dimensionless variables result in five budget equations. The first one deals with total shortwave radiation energy. The fraction not reflected to space must be equal to the sum of absorption by the atmosphere (H^*) and by the ground $(1-a)G^*$,

$$1 - A = H^* + G^*(1-a). \tag{1}$$

Set of observed values	Variables to be prescribed	Variables to be calculated	Remarks and Examples
G^*, D^*, a	μ, $\varkappa$	α, σ, A, H^*	Conventional case of ground-based measurements: especially applicable to a uniform lower boundary with uniform albedo; example: desert La Joya, or prairie country O'Neill.
G^*, A, a	μ, $\varkappa$	D^*, H^*, α, σ	Satellite-based measurements of A, with G^* from network of surface stations, (a) from regional climatological survey flights; example: Hanson's USA data.
G^*, D^*, α, σ	μ	a, H^*, A, $\varkappa$	G^* and D^* from recordings in complex surroundings; example: Robinson's data for Kew.
α, σ, μ, $\varkappa$	a	G^*, D^*, A, H^*	Three examples of shortwave radiation-climatonomy for a systematic variation of only one of the prescribed variables at a time.
a, σ, μ, $\varkappa$	α	G^*, D^*, A, H^*	
a, α, μ, $\varkappa$	σ	G^*, D^*, A, H^*	
G^*, D^*, a, A, α	—	σ, $\varkappa$, μ, H^*	Cases of detailed radiation measurements including airborne and satellite data; example: Center of Climatic Research study in Rajasthan desert.
G^*, D^*, a, A, σ	—	α, $\varkappa$, μ, H^*	

The second equation considers beam radiation $(G^* - D^*)$. Its attenuation $(1 - G^* + D^*)$ must equal the depletion by absorbers and scatterers,

$$1 - G^* + D^* = \alpha + \sigma. \tag{2}$$

The third equation deals with radiation reflected back to space, the "earth-shine". The top albedo (A) is the sum of the outward directed part of primary scattering ($\mu\sigma$) plus that part of ground-reflected global radiation which is neither absorbed in the air nor scattered back to the ground,

$$A = \mu\sigma + (1 - \alpha)aG^*(1 - \varkappa\sigma). \tag{3}$$

The fourth equation expresses the diffuse radiation arriving at the ground. It consists of the downward directed part of primary scattering $(1 - \mu)\sigma$ and of that part of ground-reflected radiation which is scattered back towards the lower boundary,

$$D^* = (1 - \mu)\sigma + (1 - \alpha)aG^*\varkappa\sigma. \tag{4}$$

The fifth equation is not independent. The overall balance of the preceding equations requires that the sum of the left-hand sides and the right-hand sides of equations (1) and (4) must be equal, or,

$$H^* = \alpha(1 + aG^*). \tag{5}$$

Equation (5) gives the heating function (H^*) as the sum of absorption of: beam radiation plus ground-reflected radiation.

With four independent equations and nine unknowns, five variables must be either observed independently, or partly observed and partly prescribed by model assumptions, or externally given in order to calculate the remaining four variables. In the following, several practically important combinations of variables are listed which can be observed or prescribed.

The fundamental equations (1) to (4) can be transformed by eliminating certain variables. The elimination of D^* in (2) with the aid of (4) produces

$$G^* = (1 - \alpha - \mu\sigma)/[1 - a\varkappa\sigma(1 - \alpha)]. \tag{6a}$$

In the special case of $a = 0$, an ideally black ground surface, (6a) reduces to $G_0^* = (1 - \alpha - \mu\sigma)$. With $d \equiv \varkappa\sigma(1 - \alpha)$, equation (6a) can be restated as

$$G^* = G_0^*/(1 - ad) \tag{6b}$$

which is Ångström's formula as quoted from Möller (1965) where d represents the "back-

Table 1. *Basic information on location and time of measurements used for case studies of shortwave radiation budgets during near-to-noon conditions under cloudless sky*

Data Sources: Kew (Robinson, 1963), La Joya (Stearns, 1966), O'Neill (Lettau and Davidson, 1957)

	Kew	La Joya	O'Neill
Location			
Geogr. Latitude	51°30′ N	16°41′ S	42°28′ N
Geogr. Longitude	0°00′ W	71°41′ W	98°32′ W
Elevation (m)	5	1,260	600
Surface Pressure (mb)	×[a]	880	949
Date of Observation			
Month, Year	May 1948	July 64	Sept. 53
Day (or Days)	15 thru 19	9 thru 15	7
Extra-Atmospheric Insolation			
Solar Constant (I_{00}, ly/min)	2.00	1.980	1.980
Reduced[b] Value (I_0)	×[a]	1.914	1.950
Sun's Zenith Distance (ϑ)	36.9°	38.7°	36.5°
Extra-Atmospheric Irradiance ($I \equiv I_0 \cos \vartheta$, ly/min)	×[a]	1.494	1.568
Observed Shortwave Radiation			
Global Radiation (G, ly/min)	0.728[c]	1.206	1.260
Diffuse Radiation (D, ly/min)	0.110[c]	0.282	0.210
Reflected Radiation (aG)	0.109[c]	0.217	0.277

[a] Not reported by Robinson (1963).
[b] Reduced to actual distance sun–earth.
[c] Reported only as fractions of extra-atmospheric irradiance, i.e. $G^* = G/I$, $D^* = D/I$, and $aG^* = aG/I$.

scatterance" of the sky. Our equation (6a) shows how backscatterance depends on the efficiency of scatterers and absorbers in the air.

3. Case studies

From information on geographical coordinates, time, surface pressure, and solar constant, the optical air mass and the extra-atmospheric irradiance can be derived with the aid of standard tables as included in the IGY manual (1958). These basic numerical data are listed in Table 1 for Kew (England), La Joya (Peru), and O'Neill (Nebraska). The three locations were selected for study because of an interesting contrast in aerosol type. At Kew, a city with presumably high industrial pollution, aerosol absorbs more than it scatters, and at La Joya, where the particulate matter in the desert air consists mostly of mineral dust, the aerosol scatters more than it absorbs. Calculated attenuation coefficients are listed in Table 2 which also shows that optical air mass and total attenuation are similar at the three stations.

The important problem is to separate attenuation by absorbing gases (H_2O, CO_2, O_3, and minor constituents) from attenuation by aerosol. Robinson (1963) has shown that in the presence of appreciable quantities of water vapor the absorption by carbon dioxide and other minor constituents will be negligibly small.

To determine the absorptivity of water vapor we used data summarized and reported by Yamamoto (1962) and derived from them the following semi-empirical formula with the aid of a least-square fit,

$$\alpha_w = 0.102\ (wM)^{0.276}, \tag{7}$$

where w = precipitable water in cm, and M = optical air mass. The actual absorption is obtained as the product of α_w times incident extra-atmospheric irradiance.

The total amount of atmospheric ozone may be estimated from climatic data. Ozone absorption is a complex process which is known to depend on total amount as well as vertical distribution of this constituent gas. Dave & Sekera (1959) have parameterized it with the aid of a simplified model of vertical O_3-distribution. Because ozone absorption is normally

Table 2. *Contributing factors and calculated coefficients of attenuation of beam radiation* ($G^* - D^*$); *case studies for near-to-noon conditions under cloudless sky at Kew, England, La Joya Peru, and O'Neill, Nebraska*

	Kew	La Joya	O'Neill
Basic information for calculation			
Coefficient of total attenuation ($\alpha + \sigma$)	0.382	0.381	0.330
Optical air mass (M)	1.250	1.112	1.163
Precipitable water (w, cm H_2O)	2.0	0.84[a]	2.10
Total ozone (equivalent cm of O_3 at NTP)	0.25	0.23	0.26
Aerosol type	City	Desert	Prairie
Attenuation coefficients describing absorption of beam radiation			
Water vapor absorptivity (α_w)	0.102	0.100	0.130
Ozone plus other gases (α_g)	0.020	0.016	0.018
Aerosol absorptivity (α_a)	0.103	0.027	0.012
All absorbers (α)	0.225	0.143	0.160
Attenuation coefficients describing scattering of beam radiation			
Neutral molecules (Rayleigh scattering σ_R)	0.095	0.103	0.107
Aerosol scattering (σ_a)	0.062	0.135	0.063
All scatterers (σ)	0.157	0.238	0.170

[a] Calculated as averages of radiosoundings at Lima (Peru) and Antofagasta (Chile).

small in comparison with water vapor effects, it is suggested here to carry the simplification one convenient step farther by assuming that this absorption is directly related to the total amount of O_3 as well as proportional to the readily calculated intensity of Rayleigh scattering. The tabulations by Dave & Sekera (1959) suggest that, for example, for a total of 0.23 cm, the O_3 absorption will amount to approximately 0.16 of the calculated Rayleigh scattering. Even though this estimate may be wrong by 40%, the effect on the relative error of total attenuation will remain less than 5%.

The atmospheric parameters necessary for the calculation of absorption quantities are listed in Table 2 for the three locations described in Table 1. An important problem is the separation of aerosol attenuation into absorbing and scattering fractions (α_a and σ_a). After determination of α_w and α_g this separation is possible with the aid of diffuse radiation data and albedo values, provided the coefficients of the scattering process μ and $\varkappa$ are prescribed. A tentative value of $\mu = \frac{1}{3}$ was used while $\varkappa$ was taken to be $\frac{3}{8}$ for Kew, $\frac{3}{3}$ for La Joya, and $\frac{3}{4}$ for O'Neill. Note that the data in Table 2 refer to beam radiation and its depletion under cloudless conditions. It is interesting that the total attenuation ($\alpha + \sigma$) turns out to be similar for "city" and "desert" conditions while the ratio α/σ is about $\frac{4}{3}$ for the city and only $\frac{2}{3}$ for the desert, and about $\frac{3}{3}$ for prairie land.

For the atmosphere above the continental United States, covering the period from March through May of 1962, Hanson (1967) derived an average top albedo value of 0.40 from TIROS IV low-resolution measurements of shortwave radiation reflected to space by earth, air, and clouds. The Peruvian desert belt normally appears as a rather dark area on satellite pictures. This suggests that local albedo values of the upper boundary should be substantially lower (possibly smaller by more than one half) than in normally cloudier regions. Thus, under cloudless conditions, it will be assumed that the top albedo is smaller than 0.20 above each of the locations considered here. Estimates of actual diffuse radiation fluxes are listed in the shortwave radiation budgets of Table 3, which also specifies seven of the nine variables defined in Section 2. The remaining two (μ and $\varkappa$) had been prescribed by model assumptions.

A more secure basis for partitioning aerosol attenuation into absorbing and scattering fractions would be desirable. The numbers presented in Table 2 are the result of trial and

Table 3. *Calculated shortwave radiation budgets (expressed as fractions of extra-atmospheric radiation) for near-to-noon conditions under cloudless sky at Kew, England, La Joya, Peru, and O'Neill, Nebraska*

	Kew	La Joya	O'Neill
Beam radiation (on horizontal surface)			
Arriving from space at outer boundary	1.000	1.000	1.000
Transmitted to lower boundary ($G^* - D^*$)	0.618	0.619	0.670
Attenuation ($\alpha + \sigma = 1 - G^* + D^*$)	0.382	0.381	0.330
Part of attenuation due to absorbers (α)	0.225	0.143	0.160
Part of attenuation due to scatterers (σ)	0.157	0.238	0.170
Diffuse, global, and reflected radiation			
Diffuse radiation arriving at lower boundary (D^*)	0.110	0.189	0.134
Global radiation at lower boundary (G^*)	0.728	0.808	0.804
Reflected at lower boundary (aG^*)	0.109	0.145	0.177
Energy albedo of the lower boundary (a)	0.150	0.180	0.220
Effective albedo of the outer boundary, planetary top-albedo (A)	0.129	0.175	0.187
Shortwave radiation budget			
Outgoing at outer boundary (A)	0.129	0.175	0.187
Absorbed in the atmosphere (H^*)	0.252	0.162	0.186
Absorbed by the ground, or transfered into heat and other energy ($(1-a)G^*$)	0.619	0.663	0.627
Balance, or incoming at outer boundary ($A + H^* + (1-a)G^*$)	1.000	1.000	1.000

error. Obviously, the ratio between aerosol scattering and absorbing is a major factor contributing to climatic differences. Robinson (1963) finds aerosol absorption to be more than twice as large as scatter but reports also that other investigators have suggested various ratios from about equal parts to hardly any absorption.

The question of absolute accuracy of the numerical values reported in Table 3 can be left open if we want to use these data exclusively as background information for the purpose of studying the effects caused by a systematic variation of a parameter, such as surface albedo, or precipitable water, or aerosol content. Controlled parameter variations under cloudless conditions will be discussed in Section 4, and with clouds present in Section 5.

Knowing the parameters α, σ, and $\varkappa$, the "backscatterance" d as defined in equation (6b) can be computed. The result is 0.05 for Kew, 0.11 for O'Neill, and 0.20 for La Joya. This sequence illustrates again differences in aerosol effects between city air and desert air. The range of d-values lies within the extremes of 0.04 and 0.30 for the backscatterance under cloudless sky as quoted by Möller (1965).

4. Systematic variation of attenuation parameters under cloudless conditions

The surface albedo (a) undergoes naturally a considerable change from summer to winter in the temperate zone. It is also most readily modified by artificial means. Hence, as a first example of shortwave radiation climatonomy, let us consider a stepwise albedo increase from $a = 0$ to $a = 1.00$, with other parameters remaining unchanged. The resulting values of A, H^*, $(1-a)G^*$, and D^* for Kew, La Joya and O'Neill are listed in Table 4. It must be emphasized that for these calculations the coefficients of scattering (σ, including μ and $\varkappa$), and absorbing (α) were kept unchanged and equal to the respective local values specified in Table 3. Most strongly affected by surface albedo changes is the fraction absorbed by the ground because of its direct proportionality to $(1-a)$. However, even in the unlikely extreme case that the earth's surface were perfectly black (with $a = 0$), only 72 % to 78 % of the extra-atmospheric irradiance would be absorbed; the remainder is partly trapped in the atmosphere by absorbers and partly scattered to space causing a significant minimum

Table 4. *Example I of Shortwave Radiation Climatonomy*

Calculated top albedo (A) and non-dimensional values of: Total absorption in the air (H^*), absorption by the ground ($(1-a)G^*$), and diffuse radiation (D^*). Assumed for the calculation is that the absorptivity (α) and the coefficient of scattering (σ) remain unchanged (and equal to the values specified in Table 3 for Kew, La Joya, and O'Neill), while the albedo (a) of the lower boundary is varied systematically, as indicated below

Albedo	Kew ($\alpha=0.225$; $\sigma=0.157$)				La Joya ($\alpha=0.143$; $\sigma=0.238$)				O'Neill ($\alpha=0.160$; $\sigma=0.170$)			
a	A	H^*	$(1-a)G^*$	D^*	A	H^*	$(1-a)G^*$	D^*	A	H^*	$(1-a)G^*$	D^*
0.00	.052	.225	.723	.105	.080	.143	.777	.158	.057	.160	.783	.113
0.04	.073	.232	.695	.106	.100	.148	.752	.164	.080	.165	.755	.116
0.12	.115	.245	.640	.109	.143	.156	.701	.178	.127	.175	.698	.123
0.24	.180	.264	.556	.113	.208	.171	.621	.198	.198	.191	.611	.134
0.36	.245	.285	.470	.117	.277	.188	.535	.216	.272	.206	.522	.145
0.52	.333	.311	.356	.123	.375	.208	.417	.250	.372	.230	.398	.159
0.76	.467	.353	.180	.131	.536	.243	.221	.301	.531	.265	.204	.182
0.88	.536	.373	.091	.136	.625	.261	.114	.329	.613	.283	.104	.194
1.00	.605	.395	.000	.140	.717	.283	.000	.357	.699	.301	.000	.207

"earth-shine" of 5% to 8%. The top albedo increases with the surface albedo although at reduced rate because part of the reflected radiation is re-reflected and absorbed in the air. This causes the intensification of relative heating H^* and diffuse radiation D^* with increasing surface albedo.

It can be concluded from Table 4 that top albedo and diffuse radiation respond strongest to changes in surface albedo if the amount of scatterers in the air is relatively large; this was the case in La Joya. The heating function reacts significantly to the amount of absorbing agents present in the air; this was the case in Kew. Ground absorption shows the strongest response to albedo changes in O'Neill where scatterers and absorbers are almost equally divided. For the maximum possible surface-albedo change (zero to unity), the range of top albedo for the city-type aerosol at Kew is 0.553, that is significantly less than 0.637 for the desert, and 0.641 for the prairie. These findings suggest that measurements of top albedoes from satellites over cloudfree areas of known surface albedoes may permit conclusions on the amount of scatterers in the air.

The data in Table 4 confirm that under otherwise unchanged conditions the global radiation increases with increasing albedo-values of the earth-air interface surrounding a pyranometer station. In nature this can happen as the result of a snowcover build-up. According to Möller (1965) this fact (first demonstrated by Ångström) was confirmed by a comparison of pyranometer data on days with, versus without, snowcover on the ground at the Canadian Station Moosonee. With beam radiation unchanged, the increase of G is caused by a larger D due to backscatter. Equations (6b) and (2) or (4) show that

$$D^* = D_0^* + adG_0^*/(1-ad), \tag{8}$$

where the subscripts denote values which would be observed when the albedo of the surroundings were exactly zero. Numerical values of backscatterances d were quoted at the end of Section 3 for the three stations. Obviously, the climatic forcing function, or insolation absorbed by the ground, must decrease with increasing albedo, because $(1-a)/(1-a\cdot d)$ is always smaller than unity with a and d, both being positive fractions. This can be proven with the aid of equation (6b).

The second example of parameter modification deals with absorption as affected by systematic variation of precipitable water content. Equation (7) was employed to calculate α_w. Although this absorptivity is dependent on optical air mass, the M-values at the three stations are comparable due to near-to-noon conditions (see Table 2). Yamamoto's equation contains only the 0.276 th power of M, a relative weak dependency, which was used by Loewe (1963) in his argument that multiple absorption is a factor of minor importance in

Table 5. *Example II of Shortwave Radiation Climatonomy*

Calculated top albedo (A) and non-dimensional values of total absorption in the air (H^*), absorption by the ground ($(1-a)G^*$), and diffuse radiation (D^*). Assumed for the calculations are that surface albeco (a) and the coefficient of atmospheric scattering remain unchanged and equal to the values specified in Table 3, while the coefficient of absorption (α) is varied as indicated below corresponding to a systematic variation of precipitable water (w)

	Precipitable water (w, cm)	Absorptivities α_w	α_{total}	A	H^*	$(1-a)G^*$	D^*
Kew							
$a = 0.15$	0.00	.000	.123	.155	.139	.706	.111
$\sigma = 0.157$	0.42	.085	.208	.135	.232	.633	.110
$\alpha - \alpha_w = 0.123$	0.84	.103	.226	.132	.250	.618	.110
	1.70	.126	.249	.127	.275	.598	.110
	2.55	.140	.263	.125	.289	.586	.109
	3.40	.152	.275	.121	.304	.575	.109
La Joya							
$a = 0.18$	0.00	.000	.043	.200	.050	.750	.196
$\sigma = 0.238$	0.42	.083	.126	.179	.144	.677	.189
$\alpha - \alpha_w = 0.043$	0.84	.100	.143	.175	.163	.662	.188
	1.70	.122	.165	.170	.188	.642	.186
	2.55	.136	.179	.166	.204	.630	.185
	3.40	.147	.190	.164	.216	.620	.184
O'Neill							
$a = 0.22$	0.00	.000	.030	.232	.036	.732	.139
$\sigma = 0.170$	0.42	.084	.114	.202	.101	.697	.134
$\alpha - \alpha_w = 0.030$	0.84	.101	.131	.196	.122	.682	.133
	1.70	.123	.153	.189	.148	.663	.132
	2.55	.138	.168	.184	.166	.650	.131
	3.40	.149	.179	.180	.179	.641	.131

comparison to multiple scattering. However, it should be emphasized that our budget-type considerations in Section 2 include multiple scattering as well as multiple absorption.

With α_w obtained from equation (7), the total absorptivity was found by adding $(\alpha_g + \alpha_a)$ which remained unchanged at values listed in Table 2. The computational results are summarized in Table 5. Least affected is the diffuse sky radiation, which changes hardly at all, whereas H^* reacts most strongly. In the case of Kew, for example, H^* increases by 0.165 units at the expense of 0.131 units in ground absorption and 0.034 units in top albedo, while D^* decreases only by 0.02 units. Considering the different ratios of α/σ, and the original ratios $\alpha_w/(\alpha_g + \alpha_a)$, the results for the two other stations are self-explanatory.

The third example is one of special importance at a time when more and more particulate matter is released into the atmosphere due to rapid industrialization all over the world. For partial answers to questions of air-pollution effects on climate, the quantitative, budget-type calculations, summarized in Table 6, are basic.

The specific parameter modified in the third example is the attenuation fraction of aerosol scattering σ_a. Keeping the ratio of aerosol absorption to scattering (α_a/σ_a) at the values listed in Table 6, the absorbing fraction α_a was calculated. With attenuation by water vapor and other gases unchanged, $(\sigma - \sigma_a)$ and $(\alpha - \alpha_a)$ were assumed to be local constants; varying total values of σ and α were calculated, and also, with the aid of equations (1) through (5), the values of A, H^*, $(1-a)G^*$, and D^*. The results are listed in Table 6.

The importance of the ratio α/σ became apparent in the calculations summarized in Table 5. In the third example, the systematic change of aerosol scattering emphasizes the significance of α_a/σ_a. This ratio is affected by the type of particles, their size, structure, and chemical composition. It can be assumed that particulate matter over La Joya consists mostly of mineral dust from the desert soil, whereas at Kew almost all comes probably

Table 6. *Example III of Shortwave Radiation Climatonomy*

Calculated top albedo (A) and non-dimensional values of total absorption in the air (H^*), absorption by the ground ($(1-a)G^*$), and diffuse radiation (D^*). Assumed for the calculations is that surface albedo (a) and the coefficients of Rayleigh scattering (σ_R) and absorption by gases (water vapor plus ozone, α_g) remain unchanged and equal to the values specified in Table 3, while coefficients of scattering and absorption by aerosol (σ_a and α_a) are systematically varied as indicated below (keeping the ratio α_a/σ_a constant)

	Attenuation by Aerosol		Total Att.						
	Scatt. (σ_a)	Absorp. (α_a)	σ	α	A	H^*	$(1-a)G^*$	D^*	(G^*-D^*)
Kew									
$a=0.15$	.000	.000	.095	.122	.140	.137	.723	.067	.01
$\sigma-\sigma_a=0.095$	.070	.116	.165	.238	.131	.264	.605	.115	.19
$\alpha-\alpha_a=0.122$	.135	.224	.230	.346	.129	.376	.495	.158	.37
$\alpha_a/\sigma_a=1.66$	.200	.332	.295	.454	.131	.485	.384	.201	.80
	.270	.448	.365	.570	.139	.597	.264	.246	3.78
La Joya									
$a=0.18$	.000	.000	.103	.116	.158	.134	.708	.083	.01
$\sigma-\sigma_a=0.103$	.070	.014	.173	.130	.166	.150	.684	.137	.20
$\alpha-\alpha_a=0.116$	.135	.027	.238	.143	.174	.163	.663	.189	.31
$\alpha_a/\sigma_a=0.200$	.200	.040	.303	.156	.183	.179	.638	.237	.44
	.270	.054	.373	.170	.194	.193	.613	.291	.64
O'Neill									
$a=0.22$	.000	.000	.107	.148	.179	.175	.646	.083	.01
$\sigma-\sigma_a=0.107$	.070	.013	.177	.161	.187	.190	.623	.137	.21
$\alpha-\alpha_a=0.148$	.135	.026	.242	.174	.195	.204	.601	.186	.32
$\alpha_a/\sigma_a=0.190$	.200	.038	.307	.186	.204	.217	.579	.235	.46
	.270	.051	.377	.199	.216	.229	.555	.287	.68

from stack effluents. At La Joya and O'Neill, the ratios α_a/σ_a are nearly equal and approximately 0.2, and addition of aerosol increases diffuse radiation, top albedo, and atmospheric heating, while ground absorption decreases. This contrasts markedly with results for Kew where α_a/σ_a equals about $\frac{5}{3}$. According to Table 2 observations at Kew and O'Neill yielded the same value of σ_a, about 0.06. However, if σ_a is raised to 0.27 (see Table 6), the beam radiation ($G^*-D^*=1-\alpha-\sigma$) decreases over the city from 0.597 to an extremely low value of 0.065, whereas the drop over the open prairie is moderate, from 0.662 to 0.424. Accordingly, the ratio of diffuse to beam radiation, $D^*/(G^*-D^*)$ varies from 0.21 to 0.68 at O'Neill and from 0.19 to 3.78 at Kew. Values larger than unity mean that diffuse radiation exceeds beam radiation. When $D^*/(G^*-D^*)$ passes unity, it will become difficult or even impossible for an observer to see his own shadow. Data in Table 6 suggest that such a state of haziness under "clear" sky may be reached at Kew when the initial level of aerosol scattering (as given in Table 2) would be about tripled. Further increase of turbidity would lead to $G^*=D^*$, which means disappearance of beam radiation; or $(\alpha+\sigma)$ approaches unity. Then the haze would cease to be translucent. Attenuation, sufficiently dense to obscure the sun, occurs normally with opaque clouds present. However, the above calculations show that in city air (with strongly absorbing aerosol) a relative modest pollution increase may extinguish the direct sunlight.

The results summarized in Table 6 are calculated for a gradual increase of aerosol scattering (σ_a) by equal increments, with a proportional increase in α_a. In a modified version of this third example, the local value of aerosol attenuation for both, σ_a and α_a, was first doubled and then increased to the fivefold of the initial level for Kew and La Joya. Fig. 1 shows the initial conditions in city versus desert air, Fig. 2 and Fig. 3 the results computed with the modified values. The graphical scheme selected is the simplest possible which shows for the incoming radiation the relative magnitudes of

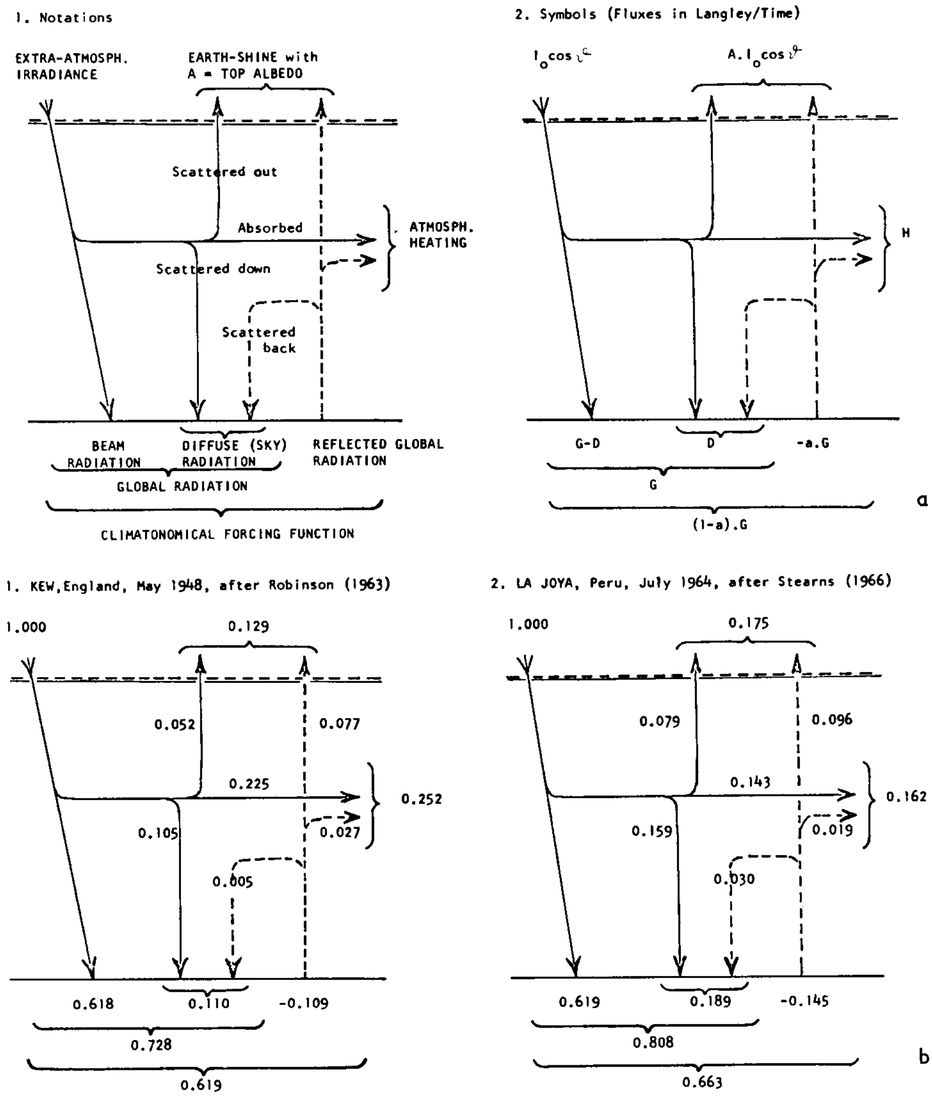

Fig. 1. Schematic illustration of the atmospheric short-wave radiation budget. (*a*) Notation used for the description of radiation flux-densities, (*b*) example of short-wave radiation budgets for air over a city in comparison with air over a desert. Consult Tables 1 to 3 for information on localities and external conditions.

attenuation (separately for scattered out, scattered down, and absorbed), the resulting beam radiation, and the reflected global radiation (separately for backscattered and absorbed). The scheme is completed by indicating the top albedo, the diffuse radiation at ground level, the total global radiation, and the absorption by the ground. Budget considerations, such as in Section 2, can be readily verified.

It can be seen that a fivefold increase in aerosol content blocks out beam radiation in both atmospheres. In the desert air, the loss of beam radiation results in a considerable increase in diffuse radiation due to the high scattering quality of desert aerosol, whereas in the city air where considerable more energy is absorbed than scattered, the heating function increases accordingly.

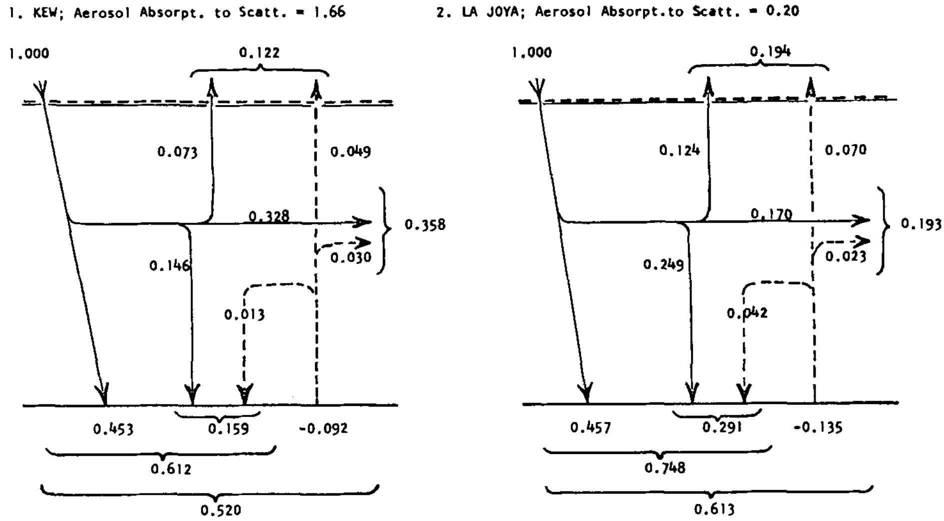

Fig. 2. Short-wave-radiation budget for air over a city in comparison with air over a desert, calculated under the assumption that aerosol amount is increased by a factor of two while other conditions remain unchanged and as illustrated on the lower half of Fig. 1. Unchanged was also the characteristic ratio which expresses scattering efficiency relative to absorbing efficiency of the local aerosol type.

5. Estimates of cloud effects on shortwave radiation fluxes

If a fraction (c) of the sky is covered by clouds the flux densities of the cloudless area have a weight factor of $(1-c)$. Let σ_c denote the effective coefficient of scattering and α_c the absorptivity of air in the cloud and, in analogy to μ in Section 2, let μ' be a fraction so that $\mu'\sigma_c = A'$ will be scattered back to space by the upper cloud surface and $(1-\mu')\,\sigma_c$ will reach the ground. The effective albedo of the cloud base will be assumed to equal A'. The employment of "effective" coefficients represents a

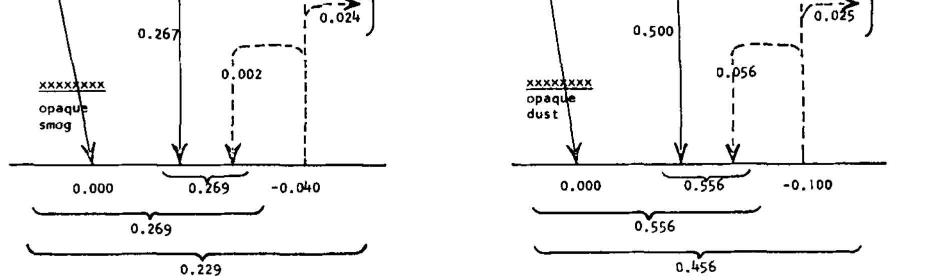

Fig. 3. Same as Fig. 2 but calculation based on the assumption that the aerosol content at both localities is increased by a factor of five. This happens to be just enough to block out direct beam radiation at both localities. Note the significant differences in diffuse radiation, atmospheric heating, and top radiation, in comparison with the initial conditions shown on the lower half of Fig. 1.

considerable model simplification of the highly complicated scattering processes which possess pronounced dependencies on particle sizes and their spectral distribution, liquid water content of the cloud, and solar elevation angle. Compressing these characteristics into a few single parameters can be tolerable only for the study of climatonomic gross-effects and overall energy partitions.

In the following energy budget equations, let the subscripts o and c indicate values for cloudless sky (with prescribed conditions for clear-air attenuation and ground albedo), and for cloudy sky, respectively. For the calculation of effective coefficients in partly cloudy air, we assume that the contributions by absorbers are additive, while contributions by scatterers are distributive (or must be prorated). Thus, in air with cloudiness (c)

$$\alpha = (1-c)\,\alpha_0 + c(\alpha_0 + \alpha_c) = \alpha_0 + c\alpha_c, \tag{9}$$

$$\sigma = (1-c)\,\sigma_0 + c\sigma_c = \sigma_0 + c(\sigma_c - \sigma_0). \tag{10}$$

From continuity principles developed in Section 2, it follows that the top albedo is composed of a prorated contribution from the clear area and direct reflection from the cloud surface, plus diffuse radiation reflected from the ground, and reduced by what is either reflected back from the cloud base or absorbed in the air. Hence,

$$A = (1-c)A_0 + c[A' + a\cdot G^*(1-\alpha)(1-A')]. \tag{11}$$

A corresponding development for the heating function H^* yields

$$H^* = (1-c)H_0^* + c\cdot\alpha(1 + a\cdot G^*) \tag{12}$$

The sum of equations (11) and (12) is used to eliminate the term $(A + H^*)$ in equation (1). The resulting relation can be solved for G^*,

$$G^* = [1 - A_0 - H_0^* + c(A_0 + H_0^* - A' - \alpha)] \\ /[1 - a + c\cdot a(1 - A' + \alpha A')]. \tag{13}$$

After G^* has been determined for a given cloudiness, A follows from (11), and H^* either from (5) or (12).

For overcast conditions we must distinguish between opaque and translucent clouds. Equations (9) and (10) apply to both cases and yield $\alpha_1 = \alpha_0 + \alpha_c$, and $\sigma_1 = \sigma_c$. For opaque clouds there is no beam radiation, so that in equation (2) $G^* = D^*$, or $\alpha_1 + \sigma_1 = 1$. Translucent clouds permit the sun to be seen, thus $\alpha_1 + \sigma_1 < 1$.

It follows from equations (9) and (2) that for opaque clouds (with $G_1^* = D_1^*$) D_1^* must be smaller than D_0^*. For translucent clouds D_1^* can be larger than D_0^*. Direct downscatter will be exceptionally large from a partly cloudy sky when the sun shines on lateral boundaries of clouds. Under not unlikely conditions such downscatter may cause $G_c^* > 1$, or to surpass temporarily even the value of the extra-atmospheric irradiance. This may occur when at noon in summer a brilliantly white cumulus tower approaches the pyranometer station from the direction opposite to an unobscured sun. Alternating sunlight and cloud shadows under broken cumulus-cover can cause interesting variations in reflected versus downcoming insolation. Examples of this effect are flight-measurements of ground albedo reported by Bauer & Dutton (1962). For climatonomical purposes, however, these special cases are not as important as statistical relations for global radiation under overcast sky.

Equation (13) is compatible with equation (6a). It is seen more easily from the latter that global radiation responds to albedo changes of the surrounding area under cloudy to overcast sky stronger than under clear sky (other conditions unchanged), because backscatterance from a cloudbase is always larger than from clear sky. Analysing radiation data from Moosonee (Canada), Möller (1965) finds a backscatterance of 0.25 from clear sky, but 0.58 to 0.69 from overcast. Similar values are quoted by Linke (1942).

The problem remains to determine the three parameters μ', α_c, and σ_c. Even if all three would be independent of the degree of cloudiness, equations (10), (11), (12), and (13) show that the change of G_c^*/G_0^* with cloudiness will depend on the turbidity in the cloudfree atmosphere. On page 440 of "Smithsonian Tables" (1963), earlier works by Kimball, Ångström, Mosby and Haurwitz are quoted, all suggesting a simplified universal relationship between insolation and cloudiness. Other authors preferred to relate global radiation linearly to the fraction of actual over possible sunshine-hours per day. These statistical relationships are discussed by Budyko (1956). Gräfe (1956) considers a nonlinear relationship between G_c/G_0 and cloudiness.

Plots of actual G_c/G_0 data versus c (as on page 440 of Smithsonian Tables) usually show a wide scatter of points. At the location of special interest for our study, the pampa de La Joya, clouds cover not more than $\frac{2}{10}$ of the sky on the daily average, and are mostly of the stratiform type. During the period of radiation measurements in July of 1964, no clouds intercepted the beam radiation. However, it became apparent that employment of an average statistical relationship between insolation and cloudcover (as taken from the literature) and disregarding the special nature of the desert aerosol, would not produce reliable results in the computation of G_c for La Joya. A new approach to the general problem of calculating cloudcover effects on global radiation became necessary under due consideration of the shortwave-radiation budget relationships derived in Section 2. Useful for this purpose are results obtained by Haurwitz (1934) concerning the relationship between global radiation and cloud type during overcast, relative to global radiation under cloudless conditions. Haurwitz included in his statistical analysis the dependency of G_1/G_0 on optical air mass. Our abbreviated summary refers to an average air mass of $M = 1.2$, but the observed variations of G_1/G_0 with M are relatively slight.

Overcast type:	Cirro-Strat.	Alto-Cum.	Alto-Strat.	Strato-Cum.	Stratus	Nimbus
G_1/G_0	0.81	0.51	0.41	0.34	0.25	0.17

One of the few systematic evaluations of attenuation effects of clearly defined cloud masses, or thicknesses, is the study by Neiburger (1949), based on aerological measurements of shortwave radiation fluxes above and below a coastal stratus in California. He finds that absorption is generally small, rarely exceeding 10%, so that transmission is mainly determined by reflection at the upper cloud surface (A') which increases significantly with increasing cloud thickness. For climatonomical calculations, it would be desirable to know the statistical relationship between attenuation parameters and degree of cloudiness at all tropospheric levels, including overcasts of different thicknesses. Neiburger's data refer only to a low stratus. Lacking detailed information, let us assume that separation by cloud types is a substitute for cloud thickness. Table 7 summarizes tentative results of attenuation parameters under overcast ranked according to probable cloud thicknesses. α_c and σ_c were obtained with the aid of equations (9) and (10). The three parameters (albedo, absorption, scattering) were determined so that a monotonic dependency on cloud rank reproduced Haurwitz's empirical data on global radiation ratios. Note that $G_1^* > D_1^*$ indicates overcast of the "translucidus" type, and $G_1^* = D_1^*$ cloud forms of the "opacus" type.

It was assumed that $\alpha_0 = 0.18$ and $G_1^* = 0.766$ (which corresponds approximately to the average clear-air conditions at O'Neill and Kew), would apply also at Blue Hill Observatory near the city of Boston. Not surprisingly, the G_1^*/G_0^*-ratios calculated for the desert atmos-

Table 7. *Tentative parameterization of shortwave attenuation in cloud overcast of various type*

α_c = absorptivity; σ_c = scattering coefficient; μ' = fraction of upward scattering; $\mu'\sigma_c = A'$ = cloud albedo; $G_1^* - D_1^*$ = beam radiation computed as $1 - \alpha_0 - \alpha_c - \sigma_c$ (for $\alpha_0 = 0.180$); G_1^* = global radiation computed as $G_1^* - D_1^* + (1 - \mu')\sigma_c$; ($G_1^*/G_0^*$ was computed assuming that in clear air $G_0^* = 0.766$)

	Cirro-Strat.	Alto-Cum.	Alto-Strat.	Strato-Cum.	Stratus	Nimbus
α_c	0.00	0.01	0.03	0.05	0.07	0.10
σ_c	.68	.80	.79	.77	.75	.72
μ'	.30	.52	.60	.66	.74	.82
A'	.200	.416	.474	.508	.555	.590
$G_1^* - D_1^*$	0.140	0.010	0.000	0.000	0.000	0.000
G_1^*	.620	.394	.316	.262	.195	.130
G_1^*/G_0^*	.81	.51	.41	.34	.25	.17

Table. 8. *Example IV of shortwave radiation climatonomy*

Calculated non-dimensional values of global radiation G^*, diffuse radiation D^*, absorption in air H^*, absorption by the ground $(1-a)G^*$, and top albedo A. Assumed for calculation is that the "clear-air" attenuation parameters remain unchanged (and equal to the values specified in Table 2 for Kew, La Joya, and O'Neill) while cloudiness c varies systematically with cloud attenuation parameters for cirrostratus and stratus (as specified in Table 7)

	Kew					La Joya					O'Neill				
Cloud c	G^*	D^*	H^*	$(1-a)G^*$	A	G^*	D^*	H^*	$(1-a)G^*$	A	G^*	D^*	H^*	$(1-a)G^*$	A
Cs 0	.728	.110	.252	.619	.129	.808	.189	.162	.663	.175	.804	.134	.186	.627	.187
$\frac{1}{3}$	.677	.234	.251	.575	.184	.760	.289	.162	.623	.215	.750	.250	.186	.585	.229
$\frac{2}{3}$	.632	.362	.248	.537	.215	.717	.392	.162	.588	.250	.706	.376	.185	.551	.264
1	.589	.494	.245	.501	.254	.678	.501	.160	.556	.284	.665	.505	.183	.519	.298
St 0	.728	.110	.252	.619	.129	.808	.189	.162	.663	.175	.804	.134	.186	.627	.187
$\frac{1}{3}$	.544	.132	.257	.462	.281	.628	.215	.170	.515	.315	.616	.169	.193	.480	.327
$\frac{2}{3}$	.354	.148	.275	.293	.432	.441	.235	.191	.362	.447	.427	.204	.213	.333	.454
1	.159	.159	.302	.135	.563	.252	.252	.226	.207	.567	.237	.237	.242	.189	.569

phere at La Joya, turn out to be systematically larger than for the two other locations, obviously due to the difference in aerosol scattering. Note that in Table 7 cloud albedo ($A' = \mu'\sigma_c$) has a trend opposite to that of downward scattering, $(\mu' - 1)\sigma_c$. With the exception of cirrostratus, the coefficient of scattering tends to remain nearly constant, between 0.80 and 0.72.

In a tentative and exploratory manner let us discuss now the fourth and last example of shortwave radiation climatonomy. We formally calculated the change in radiation fluxes which occur if cloudiness is increased stepwise from zero to unity ($c = 0, \frac{1}{3}, \frac{2}{3}, 1$) for cirrostratus and stratus at Kew, La Joya, and O'Neill. The results summarized in Table 8 are self-explanatory. Interesting is the difference between translucent and opaque cloud conditions. The top albedo for overcast shows still a dependency on ground albedo, rather significantly so for translucent, and rather weakly for opaque clouds. Atmospheric heating (H^*) decreases slightly with increasing c for translucent clouds, but increases considerably for opaque clouds.

The differences due to clear-air attenuation between the three locations are maintained; global radiation at Kew, for example, is always relatively low. This demonstrates again that one can hardly expect a universal relationship between type of overcast and global radiation.

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КЛИМАТОНОМИЯ КОРОТКОВОЛНОВОЙ РАДИАЦИИ

Обсуждаются функциональные соотношения, описывающие процессы поглощения и рассеяния коротковолновой радиации в атмосфере. Выводится упрощенная модель, в которой рассматриваются только цолные интенсивности (проинтегрированные по всему спектру инсоляции). Устанавливаются соотношения баланса, связывающие альбедо на верхней границе атмосферы с эффективным поглощением в воздухе и в почве. Измерения диффузной радиации неба, в дополнение к глобальной радиации используются для определения рассеянной и иоглощенной компонент аэрозольного ослабления в условиях ясного неба при известных водности и рэлеевском рассеянии. Для трех континентальных мест выбраны данные, представляющие ослабление при ясном небе над пустыней (Ла Хойя, Перу), над открытой степью (О'Нейл, Небраска) и над индустриальной городской областью (Кью, Англия). Отношение поглощения к рассеянию проявляет значительные систематические вариации, цричем городской аэроэоль более эффективен как поглотитель, а аэроэолр пустыни — как рассеиватель. Рассчитываются эффекты, вызываемые систематическими вариациями 1) альбедо поверхности, 2) водности, 3) аэрозоля и 4) облачности. Кроме других результатов, показывается, как альбедо на верхней границе атмосферы (измеряемое в настоящее время с обращающихся или синхронных спутников) будет реагировать на изменения альбедо поверхности и азрозольного рассеяния. Численно также показано, как влияние облаков на глобальную радиацию модифицируется существующими условиями замутненности в безоблачном воздухе.