Effect of radiation on the equilibrium of the higher layers of the troposphere and the nature of the transition from troposphere to stratosphere.

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With 8 figures.

Introduction.

The low temperatures and sharp inversions characteristic of the tropical tropopause have not yet received a satisfactory explanation. An associated striking characteristic of the distribution of temperature in the tropical atmosphere is the frequent occurrence of high lapse-rates, often approaching and sometimes exceeding that of dry air adiabatic, between about 8 and 14 km. This feature is very pronounced in the results of sounding balloon ascents over north India ⁽¹⁾ during the months June to October and over Java ⁽²⁾ and the peninsular portion ⁽³⁾ of India in all parts of the year. These high lapse-rates occur both in clear and cloudy weather. In temperate latitudes also, a layer of maximum lapse-rate is generally met with in the upper part of the troposphere, at heights of 6 to 8 km ⁽⁴⁾, the height decreasing with decrease in height of the tropopause ⁽⁵⁾. In a recent interesting paper, *F. Albrecht* ⁽⁶⁾ has pointed out that the loss of heat by radiation in the sub-stratosphere would result in the setting up of high lapse-rates there.

In this paper, the loss or gain of heat by each kilometre layer of the upper half of the troposphere due to absorption and emission of both long-wave and solar radiation has been evaluated assuming different values of moisture-content in these atmospheric layers. The moisture is assumed to be entirely in the form of vapour. The calculations show that as a necessary consequence of the very low temperatures at the base of the stratosphere over the tropics, there should be a region of active convection between about 8 and 12 km and weaker convection for two or three kilometres further above, both during day and night. It is suggested that by a cycle of cause and effect the radiation helps to maintain the high levels and low temperatures of the tropical tropopause.

Data used and the method of calculation.

From *Hettner*'s ⁽⁷⁾ and *Paschen*'s ⁽⁸⁾ experimental results on the absorption of water-vapour, curves showing the variation of transmission with wave-length were prepared for the following eleven ranges of wave-lengths: 1.08—1.2 μ, 1.25—1.50 μ, 1.7—2.0 μ, 2.3—3.0 μ, 5—6 μ, 6—7 μ, 7—8 μ, 13—16 μ, 16—20 μ, 20—30 μ and 30—34 μ corresponding to known quantities of water-vapour in the

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path of the radiation. Other ranges of wave-length for which either the energy of radiation or the absorption in the higher layers of the troposphere was comparatively small, were neglected. From these curves, derived curves of transmission for five other quantities of water-vapour were prepared — assuming *Beer*'s law to hold[1]. The integrated percentage transmissions for each of these ranges of wave-lengths were then plotted against the logarithms of precipitable water (see Fig. 2) as has been done by *Abbot* (9) in his paper on "the Radiation of the planet Earth to space". From these curves, the transmissions for any quantity of moisture in the path of the radiation could be read off.

For the regions 2.3—3.0 μ and 13—16 μ where absorption by carbon dioxide is strong, similar transmission curves were prepared for varying quantities of carbon dioxide using the experimental results of *C. Schaefer* and *B. Philipps* (10) for the former region and those of *Rubens* and *Aschkinass* (11) for the latter. For computing the absorption produced by any layer, the transmissions through the water-vapour and carbon dioxide contained in the layer were compounded by multiplication. The transmission curves for wave-lengths shorter than 3 μ are used for the calculation of the absorption of solar radiation in the atmosphere and those for 5 μ and above for calculating the emission and absorption of atmospheric radiation. The *net long-wave* atmospheric radiation becomes negligibly small for wave-length shorter than 6 μ.

The emission of radiation from any layer within each of the above-mentioned ranges of wave-lengths was obtained by multiplying the black-body radiation at the mean temperature of the layer by the fraction representing the absorption in that layer. A convenient table of black body radiation in different ranges of wave-lengths is given by *Abbot* in his paper referred to above. Equal amounts of energy are supposed to be radiated from the upper and lower boundaries of the layer. Owing to the diffuseness of the radiation, the length of path in any layer is on the average greater than the thickness of the layer and in order to allow for this, the amount of absorbing material in the layer was supposed to be twice the actual amount. This does not, however, apply to the absorption of solar radiation, where the actual amount of absorbing material along the ray in the layer, depending on the altitude of the sun, was taken. The energy of solar radiation just outside the earth's atmosphere was calculated with the help of data given in the International Critical Tables on the assumption that the sun radiates as a black body at 6000° A.

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1) The assumption of *Beer*'s law regarding the absorption of water-vapour and carbon dioxide in the atmosphere is open to criticism. *Hettner*'s experiments were made with water-vapour at atmospheric pressure and temperatures 81° C or 127° C. *Rubens* and *Aschkinass*' observations were also at atmospheric pressure and temperature. It is known from the experiments of *Eva Bahr*, *Hertz*, *Schaefer* and others that the character and intensity of the absorption bands of a gas depend to some extent on its pressure and temperature and if mixed with other gases, on the nature and presence of the added gases (cf. *Schaefer* and *F. Matossi*: *Das Infrared Spectrum*, J. Springer, Berlin). Insufficient resolution of the absorption lines in the experiments would also invalidate the application of *Beer*'s Law. But with the quantitative data available at present, it seems impossible to introduce greater precision. Accurate data regarding absorption of both water-vapour and carbon dioxide under conditions such as occur in the neighbourhood of the tropopause are very much needed.

The numbering of the layers was done according to their temperatures in the manner adopted by *Simpson* in his famous papers on Atmospheric Radiation (12) — the layer + 1 having a mean temperature of 223° A, + 2 having 229° A and so on, the mean temperature increasing by 6° A for each increase of number of layer by 1. Proceeding in the opposite direction, the layer — 1 has a mean temperature of 217° A, — 2 a temperature of 211° and so on. The calculations were carried out for two values of humidity 100% and 25%. It must not, however, be overlooked that considerable supersaturation is possible in the higher layers of the troposphere. From an admirable series of experiments, *C. F. Powell* (13) obtained the following values of supersaturation at the cloud limit in rapidly-expanded dust-free air.

Initial Temp. °C	Expansion rates v_2/v_1	Final Temp. °C	Super-Saturation
77	1.252	47	2.87
50	1.286	10.1	3.74
35	1.314	3.2	5.07
18	1.370	— 16.4	7.80
7	1.375	— 26.4	8.95

The super-saturation increases with decreasing temperatures; it would therefore not be surprising if the upper layers of the tropsphere are highly supersaturated at times when there is an accumulation of moist air in the lower levels.

Moisture in the Stratosphere.

A question of some difficulty is that of the moisture-content of the stratosphere. Assuming that the temperature at the base of the tropopause is 220° A, that the air there is saturated and that the distribution of moisture in the stratosphere is given by *Dalton*'s law, *Simpson* adopted 0.03 cm of precipitable water as the probable moisture-content of the stratosphere. *Abbot* states that there are many occasions when the *total* amount of water-vapour over Mount Montezuma in Chile, altitude 2710 m, as determined from the spectro-bolograms obtained there approached the value adopted by *Simpson* and that it is therefore likely that the moisture-content of the stratosphere is considerably smaller than the value assumed by *Simpson.* Over the tropics at any rate, the moisture-content in the stratosphere must be much less than 0.03 cm as the temperature of the tropical tropopause is almost invariably lower than 200° A. The saturation vapour pressure of water at 220° A is 0.027 mb, while at 200° A, it is only 0.0016 mb (14). This difference while not in any way affecting *Simpson*'s results regarding the atmosphere as a whole, will be of great influence in determining the radiation equilibrium of the higher layers of the troposphere. In the present paper, the values of radiation of different atmospheric layers have been calculated assuming the effective temperature of the stratosphere to be 220° A for the following values of precipitable water in the atmosphere above layer — 3; .03 cm, .01 cm and .0015 cm. The last of these values corresponds approximately to the water-content in the stratosphere on the assumption that the temperature at its base is 200° A and that the distribution of moisture therein is governed by *Dalton*'s law.

Results.

The method of calculation will be clear from the following example. In Fig. 1 are drawn the absorption curves of water-vapour in the wave-length range 16—20 μ for the following seven values of moisture-content: — .1 cm, .05 cm, .0177 cm, .01 cm, .005 cm, .001 cm and .0002 cm. Of these, the third is *Hettner*'s experimental curve and the others are derived from this using *Beer*'s law. The corresponding percentages of absorption are 98, 88, 59, 41, 23, 5 and 1. These are plotted against the logarithms of moisture-content in Fig. 2 (curves for other regions of the long-wave spectrum are also drawn in the same figure). Using this

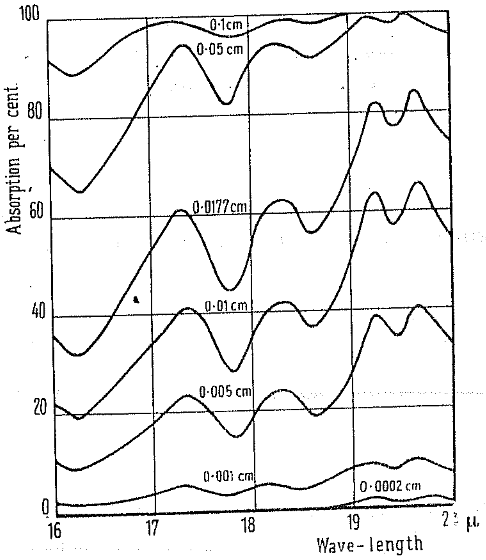

Fig. 1.

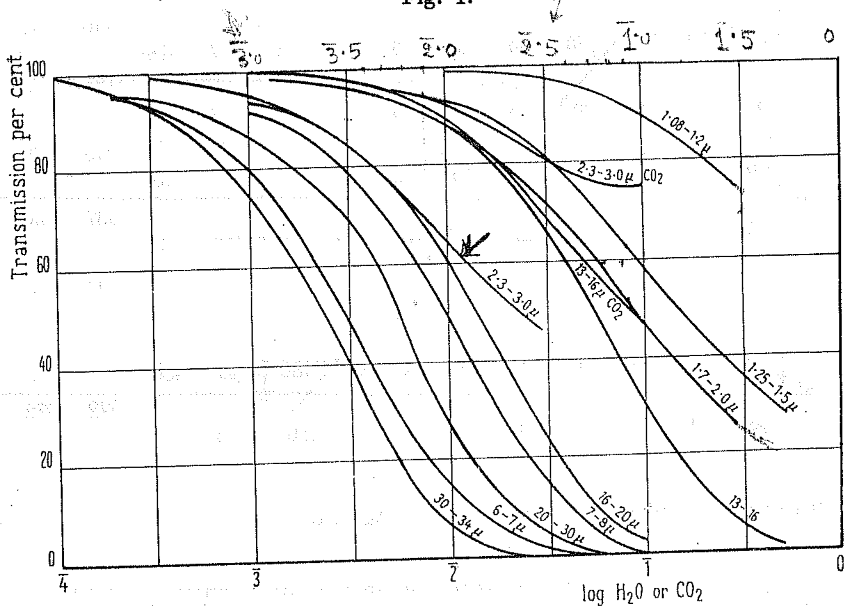

Fig. 2.

curve, tables similar to Table 1 for calculating the net radiation loss from each layer under different assumptions regarding moisture-content of the atmosphere were prepared.

Table 1.

Atmospheric Radiation 16—20 μ; unit = cal. cm^{-2} min^{-1} $\times$ 10^{-4}.

(Assumed relative humidity in troposphere = 100%; moisture-content of stratosphere = .0015 cm of precipitable water.)

Atmospheric layer	Assumed mean temperature of layer	Percentage absorption in layer	Downward and upward radiation from each layer									Total downward radiation ↓	↑ Total upward radiation	Total radiation entering layer ↓+↑	Absorption in layer	Emission from layer	Net radiation from layer
1	2	3	4									5	6	7	8	9	10
	220	16															
			54	10	31	56	80	98	65	21	2	54	363				
−3	205	4												422	17	20	3
			52	10	32	58	84	102	68	22	2	62	368				
−2	211	11												440	48	64	16
			46	9	32	65	95	115	76	25	2	87	378				
−1	217	20												478	96	130	34
			37	7	26	65	118	144	95	31	3	135	391				
1	223	33												543	179	236	57
			25	5	17	43	118	215	142	46	5	208	408				
2	229	55												637	350	430	80
			12	2	8	20	53	215	315	102	12	310	429				
3	235	75												768	576	630	54
			3	1	2	5	13	54	315	409	49	393	458				
4	241	90												879	790	818	28
			0	0	0	1	1	5	31	409	486	447	486				
5	247	100												972	972	972	0
			0	0	0	0	0	0	0	0	486	525					

The numbers in column 4 below the diagonal line represent the downward and those above the line the upward radiation. For example, 118×10^{-4} cal/cm^2 is radiated by layer 1 every minute both upwards and downwards. Of the downward radiation, 53 units pass through layer 2, 13 units through layers 2 and 3, 1 unit through layers 2, 3 and 4, and so on. The values in column 5 give the total atmospheric radiation passing through the boundaries between different layers and represent the sum of the figures in column 4 to the left of the diagonal line. Similarly, the figure in column 6 represents the total upward atmospheric radiation. The sum of the downward radiation passing through the upper boundary of the layer and the upward radiation passing through the lower boundary are given in column 7. This, multiplied by the fraction representing absorption in the layer gives the energy absorbed in it. The other columns are self-explanatory.

In Tables 2 (a) und (b) are collected together the values of gain of energy by long wave radiation of the layers + 6 to — 3 assuming the moisture-content of the stratosphere to be .0015 cm and the relative humidity in the tropospheric layers to be 100 and 25 per cent, corresponding to cases (a) and (b) respectively. In the same tables, the rate of absorption of energy per sq. cm from solar radiation (1) when the rays are incident normally and (2) when the secant of the sun's zenith distance is 2, are also given. All the absorption bands due to water-vapour in the region 0.9—3.0 μ and the carbon dioxide band near 2.7 μ have been taken into account. The carbon dioxide is supposed to be present in a constant proportion of 0.03 per cent in all layers.

Table 2.

Net radiation from different layers of the atmosphere. Unit = cal cm^{-2} min^{-1} $\times 10^{-4}$.

Precipitable water in stratosphere assumed to be 0.0015 cm.

(a) Relative humidity in troposphere assumed to be 100 per cent.

Atmospheric layer (1)	Rate of absorption of solar radiation										Rate of absorption due to long-wave radiation (4)							Total rate of gain of energy due to radiation 5 = (3) + (4)
	Normal incidence (2)					Sec Z = 2 (3)												
	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	6—7μ	7—8μ	13—16μ	16—20μ	20—30μ	30—34μ	Total	
str.	0	13		84		0	13				3	1	9	— 3	— 23	— 3	— 16	5
— 3	0	0	0	28	28	0	0	0	21	21	3	0	3	— 16	— 51	— 9	— 70	— 25
— 2	0	13	6	38	57	0	13	6	26	45	2	1	— 6	— 35	— 72	— 12	— 122	— 61
— 1	0	26	12	49	87	0	19	12	30	61	0	— 3	— 14	— 57	— 75	— 7	— 156	— 70
1	0	38	24	57	119	0	30	25	31	86	— 1	— 6	— 20	— 81	— 48	— 2	— 158	— 18
2	10	66	51	65	192	10	63	41	26	140	0	— 8	— 32	— 54	— 18	0	— 112	71
3	20	116	83	55	274	15	93	60	15	183	0	0	— 35	— 27	0		— 62	144
4	30	167	109	33	339	25	114	60	7	206		1	— 26	0			— 25	196
5	50	230	125	17	422	40	127	51	3	221			— 9				— 9	187
6	70	250	102	6	428	65	102	29	—	196								

(b) Relative humidity in troposphere assumed to be 25 per cent.

Atmospheric layer (1)	Rate of absorption of solar radiation										Rate of absorption due to long-wave radiation (4)							Total rate of gain of energy due to radiation 5 = (3) + (4)
	Normal incidence (2)					Sec Z = 2 (3)												
	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	6—7μ	7—8μ	13—16μ	16—20μ	20—30μ	30—34μ	Total	
str.	0	13		84		0	13				1	0	10	0	— 3	0	8	25
— 3	0	0	0	25	25	0	0	0	17	17	2	0	5	1	— 9	— 4	— 5	13
— 2	0	0	0	29	29	0	0	0	18	18	2	0	— 1	— 7	— 27	— 7	— 40	— 12
— 1	0	0	6	33	39	0	6	3	19	28	0	0	— 5	— 22	— 55	— 11	— 93	— 53
1	0	13	6	35	54	0	13	6	21	40	— 4	— 3	— 12	— 45	— 78	— 13	— 155	— 102
2	0	26	12	42	80	0	19	12	22	53	— 1	— 10	— 20	— 66	— 84	— 16	— 197	— 121
3	0	38	24	49	111	0	30	23	23	76	— 3	— 11	— 22	— 71	— 49	— 10	— 166	— 56
4	10	61	45	51	167	10	52	36	18	116	3	— 11	— 25	— 70	— 16	— 2	— 121	43
5	20	104	73	42	239	15	85	53	11	164	.	— 8	— 25	— 43	— 8	.	— 84	108
6	30	148	162	28	368	20	108	59	5	192	.	.	— 26	— 9	.		— 35	

A summary of data for the cases when the precipitable water above — 3 is 0.01 cm and 0.03 cm is given in Tables 3 and 4.

Table 3.

Net radiation from different layers of the atmosphere. Unit = cal cm^{-2} $min^{-1} \times 10^{-4}$. Precipitable water in stratosphere assumed to be 0.01 cm. stratosphere above — 3.

(a) Relative humidity in troposphere assumed to be 100 per cent.

Atmospheric layer (1)	Rate of absorption of solar radiation										Rate of absorption due to long-wave radiation (4)							Total rate of gain of energy due to radiation 5 = (3) + (4)
	Normal incidence (2)					Sec Z = 2 (3)												
	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	6—7μ	7—8μ	13—16μ	16—20μ	20—30μ	30—34μ	Total	
str.	15	92	70	187	364	15	85	65	120	285								
—3	0	0	0	17	17	0	0	.	11	11	4	3	17	3	+ 4	5	36	47
—2	0	12	5	21	38	0	11	5	15	31	6	3	9	0	+ 7	6	31	62
—1	0	24	11	36	71	0	17	10	17	44	5	6	0	— 6	10	2	17	61
1	0	35	21	52	108	0	27	21	20	68	0	3	— 8	—18	— 8	1	— 30	38
2	10	57	46	43	156	10	51	33	17	111	—2	—1	—23	—32	—16	— 3	— 77	34
3	20	110	75	38	243	15	83	50	10	158		—6	—29	—32	— 9		— 76	82
4	30	158	98	25	311	25	100	48	5	178		0	—32	—21			— 53	125
5	50	215	110	29	404	40	113	41	2	196			—25	— 1			— 26	170
6	70	235	92	5	402	65	91	23	.	179			— 9				— 9	170
(b) Relative humidity in troposphere assumed to be 25 per cent.																		
str.	15	92	70	187	364	15	85	65	120	285								
—3	0	0	0	15	15	0	0	0	10	10	1	1	15	0	5	2	24	34
—2	0	0	0	19	19	0	0	0	11	11	2	2	9	4	10	3	30	41
—1	0	6	5	21	32	0	5	3	12	20	6	1	4	0	5	2	18	38
1	0	12	5	23	40	0	11	5	13	29	7	3	— 1	— 5	— 4	0	0	29
2	0	24	9	26	59	0	17	9	14	40	2	0	— 8	—19	— 7	— 2	— 34	6
3	0	35	21	35	91	0	27	20	14	61	2	—4	—17	—31	—26	— 2	— 78	— 17
4	10	57	41	37	145	10	41	29	11	91		—6	—18	—37	—23	— 7	— 91	0
5	20	98	66	33	217	15	75	42	7	139		—9	—27	—44	—10		— 90	49
6	30	139	145	32	346	20	94	47	3	164		—7	—23	—23	— 8		— 61	103

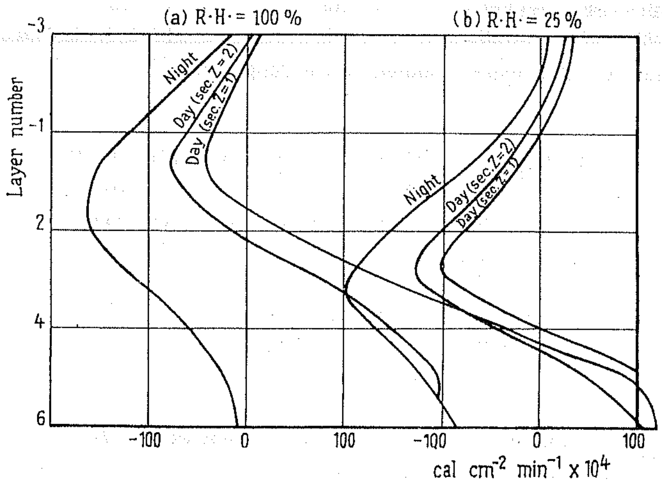

Fig. 3. Net rate of gain of energy by radiation by different atmospheric layers. Moisture in stratosphere assumed to be .0015 cm.

Table 4.

Net radiation from different layers of the atmosphere. Unit = cal cm^{-2} min^{-1} × 10^{-4}.
Precipitable water in stratosphere assumed to be 0.03 cm.

(a) Relative humidity in troposphere assumed to be 100 per cent.

Atmospheric layer (1)	Rate of absorption of solar radiation										Rate of absorption due to long-wave radiation (4)							Total rate of gain of energy due to radiation 5 = (3) + (4)
	Normal incidence (2)					Sec Z = 2 (3)												
	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	6—7μ	7—8μ	13—16μ	16—20μ	20—30μ	30—34μ	Total	
str.	30	236	174	258	698	30	197	127	153									
−3	0	0	0	13	13	0	.	2	8	10	5	3	22	7	9	5	51	61
−2	0	11	4	15	30	0	10	6	11	27	6	4	16	12	19	8	65	92
−1	0	22	9	26	57	0	13	7	11	31	5	8	7	12	26	2	60	91
1	0	31	17	28	76	0	22	15	13	50	0	6	1	3	6	1	17	67
2	10	51	37	34	132	10	47	25	11	93	−2	1	−11	−14	−10	−3	−39	54
3	20	96	60	20	196	15	66	37	7	125		−6	−22	−16	−13		−57	68
4	30	138	80	18	266	23	81	37	3	144		1	−28	−16	1		−42	102
5	50	184	91	10	335	40	90	31	1	162			−22	2			−20	142
6	70	205	74	4	353	65	68	17	0	150			−9				−9	141

(b) Relative humidity in troposphere assumed to be 25 per cent.

Atmospheric layer (1)	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	1.08—1.2μ	1.25—1.5μ	1.7—2.0μ	2.3—3.0μ	Total	6—7μ	7—8μ	13—16μ	16—20μ	20—30μ	30—34μ	Total	5 = (3) + (4)
str.	30	236	174	258	698	30	197	127	153	507								
−3	0	0	0	10	10	0	0	0	7	7	2	1	11	1	7	2	24	31
−2	0	0	0	14	14	0	0	0	7	7	3	3	18	3	14	3	44	51
−1	0	0	4	15	19	0	5	0	9	14	5	2	12	5	11	4	39	53
1	0	11	4	17	32	0	9	3	9	21	3	4	9	5	6	3	30	51
2	0	22	9	18	49	0	13	7	9	29	−2	3	−1	−2	7	0	5	34
3	0	31	17	23	71	0	22	13	9	44	−1	−3	−10	−9	−14	−2	−39	5
4	10	50	33	24	117	10	39	21	7	77	−3	−5	−16	−16	−18	−7	−65	12
5	20	86	53	21	180	15	60	32	5	112		−8	−18	−17	−9		−52	60
6	30	121	117	14	282	20	75	36	3	134		−7	−20	−16	−8		−51	83
													−25	−7			−32	

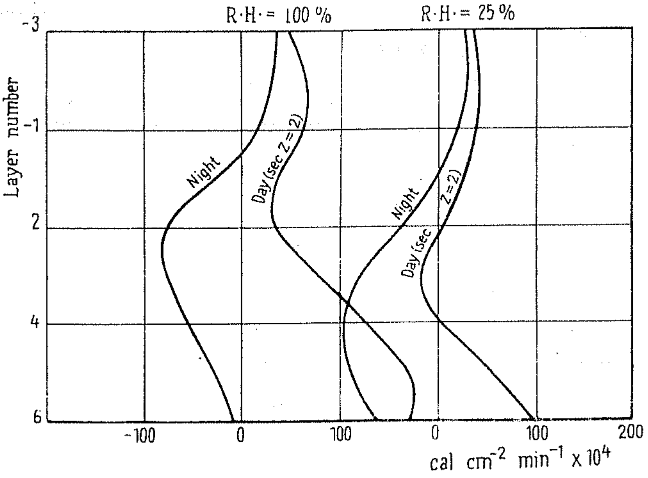

Fig. 4. Net rate of gain of energy by radiation by different atmospheric layers. Moisture in stratosphere assumed to be .01 cm.

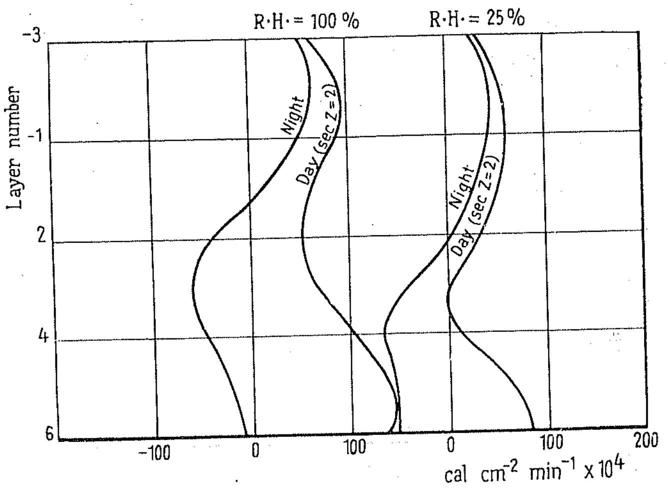

Fig. 5. Net rate of gain of energy by radiation by different atmospheric layers. Moisture in stratosphere assumed to be .03 cm.

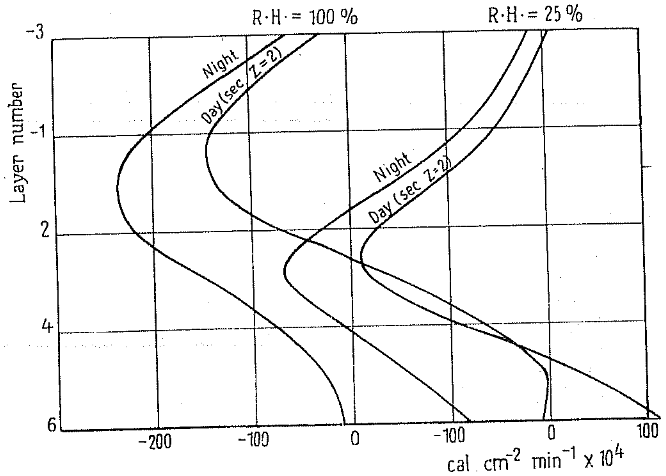

Fig. 6. Net rate of gain of energy by radiation by different atmospheric layers. No moisture in stratosphere.

The net radiation from each layer at night and also during day when the secant of the Sun's Zenith distance is 2 are given in Figs. 3, 4 and 5. In Fig. 3 alone, the case when sunlight is incident normally has also been included. As a limiting example, the values of radiation have also been calculated when the amount of moisture in the stratosphere is zero (see Fig. 6). In all the above cases, the stratosphere is supposed to lie above — 3, i. e. at a height of about 16 km as in the tropics.

Fig. 7 refers to the case when the temperature is uniformly 220° A above layer + 1; as usual, two values of humidity 100% and 25% are assumed for the troposphere and 25% for layers — 1 to — 3. .01 cm has been assumed to be the total moisture-content of the layers above — 3. Fig. 8 refers to temperature conditions such as exist on the average over Agra (India) during September and October (1).

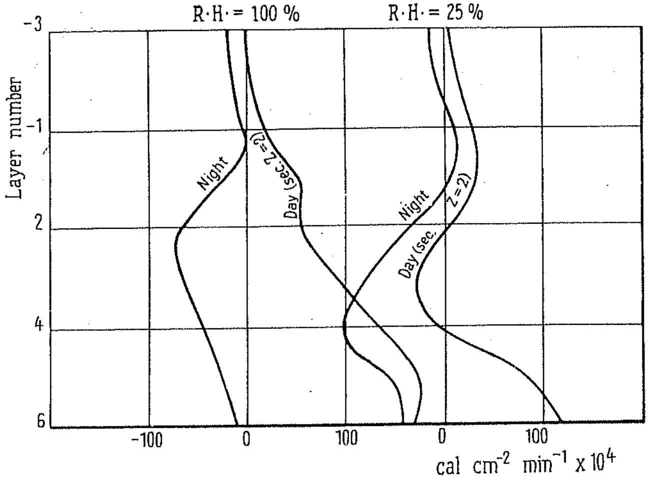

Fig. 7. Net rate of gain of energy by radiation by different atmospheric layers. Isothermal 220^0 A above + 1. Moisture above — 3. Assumed to be .01 cm.

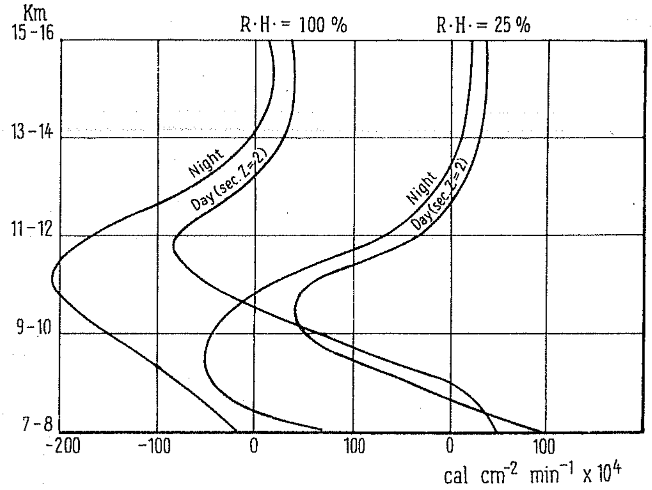

Fig. 8. Net rate of gain of energy by radiation by different layers of the atmosphere under mean N. India conditions in September and October — Moisture above — 3 assumed to be .0015 cm.

Discussion of results.

A very striking feature of the curve of atmospheric radiation which corresponds to 100% humidity in the troposphere in Fig. 3 is the large increase in the rate of loss energy as we go up from layer 6 to layer 2 or 1 and its rapid decrease at greater heights. The net loss of energy in layers 6 and — 3 are small. In the tropics, layers 6, + 1 and — 3 correspond to heights of about 8, 12 and 15 km respectively; in temperate latitudes, the corresponding heights are 2 to 3 km lower. When the effect of solar heating (with sun's rays incident at an angle of 60°) is added, there is a net *gain* of energy up to layer + 2 and maximum net loss of energy in layers + 1 and — 1. When the angle of incidence is smaller, the heating is greater. Whether we consider day or night conditions they are favourable for the increase of lapse-rate and eventually the development of instability and active convection in layers 5 to 1 — the intensity of instability and the maximum height of its extension being greater during day than during night. The convection may be expected to extend beyond — 1 and may go up even to — 3, i. e. practically to the upper limit of the region where there is a net loss of energy by radiation but the lapserate above 1 would be much less than in the layers immediately underneath. Fig. 3 (b) shows that a decrease of humidity in the troposphere from 100 to 25 per cent lowers the height of maximum loss of energy by about 2 km, but does not cause any other serious alteration in the character of the curves. Any super-saturation in the higher layers of the troposphere will tend to raise the level of maximum cooling by radiation.

A comparison of Figures 3, 4, 5 and 6 shows that an increase of moisture-content in the stratosphere produces the following effects:

(1) It lowers the height of the layer of maximum energy loss.

(2) It reduces the actual and differential cooling of the layers immediately below the above mentioned layer thus diminishing the tendency for the setting up of high lapse-rates in the lower layers, and

(3) When the moisture-content of the stratosphere is 0.01 cm or more, the effect of radiation is to increase the energy in the higher layers even during night.

Comparison of Figures 4 and 7 shows that when decrease of temperature with height is assumed to stop above layer + 1, the rate of increase of energy in the layers — 1 to — 3 becomes smaller. The effect of substituting Agra conditions of temperature in the troposphere in September and October (Fig. 8) for those in Fig. 3 is to lower the layer of maximum energy loss but steepen the differential cooling of the layers.

We may also consider the actual rates of cooling. In Fig. 8 (a) the layer 10 to 11 km cools at the rate of 210×10^{-4} cal/cm^2/min when the sun is not shining. Taking the specific heat of air at constant pressure to be 0.242 and the mass of air over a square centimetre between 10 and 11 km to be 39 gms [Table 20 of (1)], the cooling of the layer in one hour of the night comes out to be 0.13° C or 1.6° C in 12 hrs. During day, with the sun at an altitude of 30°, the differential cooling of layer 11—12 km compared to 8—9 km is 0.16° C per hour. The fact that the maximum cooling during night hours is at a greater height than the maximum heating during day is of great importance in promoting increase of lapse-rate and not merely producing a slow settling down of the higher layers. Naturally, whenever extra warm air is introduced at heights of 7 to 9 km, it would lead to increased instability.

If we compare the curves of radiation given above with the height temperature curves in different parts of the world [Fig. 9 of (16)], the resemblance of the two sets of curves becomes very striking and strongly suggests that the difference between the nature of the transitions from troposphere to stratosphere in the tropics and temperate latitudes is due to the different amounts of moisture present in the stratosphere. We have already seen that there are reasons to believe that there is less moisture present in the stratosphere in the tropics than in temperate latitudes.

Accepting as a hypothesis for the present that there is a moisture content of about 0.0015 cm in the tropical stratosphere, it is easy to see what will happen if the atmosphere is left to itself under the sole action of radiation. Let us take the case when the humidity in the troposphere is 100 %. Gradual increase of lapse-rate from 7 to 12 km will lead to mixing of air at different levels in the sub-stratosphere and condensation in levels like 3, 2 and 1. The result will be that some moisture will drop out as precipitation and by mixture with air at higher levels the air will get gradually drier and slowly tend to condition represented by 3 (b). The top part of sub-stratosphere will begin to get a net gain of heat instead of a loss and get absorbed into the stratosphere. The layer of maximum loss of heat due to radiation will also get lower and the minimum temperature reached correspondingly higher. More moisture will thus be able to pass into the stratosphere. We shall thus progress towards conditions represented by Fig. 4 (b) or even those of Fig. 7 (b) — where the differential cooling of layers is smaller and changes are generally more gradual. It seems therefore likely that the tropical stratosphere and sub-stratosphere when left to themselves under the sole action of radiation without further supplies of moisture from below would tend to get to the conditions of those of temperate latitudes.

Summary of reasons for the high level and low temperature of the tropical tropopause.

In the lower levels of the troposphere up to about 8 km the potential temperature over equatorial regions is higher than over temperate regions, and more important still, the moisture-content is greater. Owing to the increasing rate at which the vapour pressure of water increases with temperature, a lapse-rate of the environment which is stable for rising moist air at a lower temperature becomes unstable when the temperature is higher. At times when instability occurs, a greater quantity of moisture can be pushed upwards in tropical latitudes and convection extend to levels of lower temperatures than in temperate latitudes. Near the top of the water-vapour column, radiation produces another layer of large lapse-rate, which is higher the smaller the amount of moisture in the stratosphere. The cooling of the upper layers of the troposphere itself is the effective cause of reducing the amount of moisture penetrating into the stratosphere. These two circumstances therefore act and react on each other. A limit is reached to the possible cooling owing to the radiative action of carbon dioxide. Carbon dioxide being a less efficient absorber of atmospheric radiation than water-vapour, but possessing significant absorption for solar radiation, would have a higher equilibrium temperature than water-vapour. (Cf. Absorption of solar radiation between 2 and 3 μ and net loss of radiation between 13 and 16 μ in Tables 2, 3 and 4.) The continued increase of temperature from the tropopause to at least 25 km over the tropics may be attributed to this cause. But this point requires further investigation.

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