551·510·535

Spatial, Seasonal and Solar Cycle characteristics of the Loss Coefficient in Ionospheric *F*-region

B. N. BHARGAVA and R. V. SUBRAHMANYAN

Astrophysical Observatory, Kodaikanal

(*Received 9 April 1963*)

ABSTRACT. Evaluations of loss coefficient at several fixed heights of ionospheric *F*-region by a comparatively simple method have been shown to yield reasonable values. Seasonal characteristics of the loss coefficient have been obtained and discussed in relation to concentration of atomic and molecular gases. From analysis at average heights of maximum electron densities, seasonal and solar cycle variations in the loss coefficient at temperate and sub-tropical latitudes have also been obtained. It has been found that the technique is not applicable to stations in low latitudes because of the large magnitudes of vertical drifts of ionization which have not been taken into account in the present method.

1. Introduction

The electron loss coefficient in the ionosphere has generally been determined from the diurnal variation of maximum electron densities and from the decay of ionization during solar eclipses. Neither of these methods is entirely satisfactory; the electron density variation method does not take into account movements of ionization and, at equatorial stations, often yields negative values of the loss coefficient. The method based on the decay of ionization also cannot be expected to yield satisfactory results, particularly for the higher regions in the ionosphere because of the large relaxation times and irregular distribution of sources of ionizing radiation on the solar disc and corona. Many of the earlier observations have, therefore, yielded loss coefficient values which differ considerably between themselves. For instance, the loss coefficient obtained by Van Zandt, Norton and Stonehocker (1960) is two to five times larger than that obtained from earlier eclipse analyses (Savitt 1950, Minnis 1956). In a comprehensive analysis, Ratcliffe *et al.* (1956) have determined the loss coefficient at several heights in the *F*-region by using true height profiles of electron density at Huancayo, Watheroo and Slough and by using methods of analysis which minimized the effect of drifts. They concluded that the loss of electrons between 250 and 350 km takes place by an attachment-like process and that the loss coefficient at any height between 250 and 350 km is given by—

$$\beta_{(h)} = 10^{-4} \exp.\left[\frac{300-h}{50}\right] \sec^{-1}$$

Schmerling, quoted by Mitra (1959), has indicated that a more detailed analysis results in an effective scale height nearer to 35 km. Van Zandt *et al.* (1960) observed that at the equatorial station of Danger Island (Geo. Lat. 10°48′ S; Geomag. Lat. 11°S) the value of the linear loss coefficient between 290 and 400 km is given by—

$$\beta_{(h)} = 6{\cdot}8 \times 10^{-4} \exp.\left[\frac{300-h}{103}\right] \sec^{-1}$$

For a height of 300 km the loss coefficient, obtained from this relation, is about seven times larger than that deduced by Ratcliffe *et al.* (1956). For the same height Bowhill (1961) has obtained loss coefficient which is also about five times that obtained from the Ratcliffe *et al.* expression. From a comparison of the photochemical theory of the *F*1-layer and its observed morphology, Hirsh (1959) has shown that the day time values of β must be an order of magnitude larger than the night time values of Ratcliffe *et al.* It appears, therefore, that no single expression for β as a function of height appears to be

TABLE 1

Values of $\beta \times 10^5$ sec^{-1} at fixed true heights deduced from (N, h) profiles, Puerto Rico, April 1959 through July 1961

	True height (km)									
	220	240	260	280	300	320	340	360	380	400
1959										
April				44·8	30·6	19·2	13·9	10·6	9·6	8·7
May				39·8	23·6	16·7	12·4	10·0	8·6	7·7
June					33·4	14·5	9·4	7·6	6·4	6·0
July		37·4	28·0	20·8	15·8	12·0	9·1	7·5	6·6	6·0
August			45·6	31·2	23·6	17·6	13·5	10·9	9·1	8·4
September			36·4	24·8	17·9	13·6	11·2	9·8	9·0	8·7
October		33·4	23·8	17·8	14·6	12·7	11·6	11·0	10·6	10·4
November		33·4	22·4	17·2	14·7	13·6	12·8	12·5	12·3	12·2
December	40·8	26·4	20·2	16·4	14·8	13·9	13·4	13·0	12·7	12·6
1960										
January	37·8	27·6	21·4	16·8	14·0	12·3	11·4	10·9	10·7	10·6
February	29·8	20·4	15·4	12·5	11·1	10·2	9·7	9·4	9·1	9·0
March	34·4	20·6	15·0	12·0	10·2	9·3	8·6	8·3	8·1	8·0
April		34·2	24·0	19·6	15·2	12·2	10·4	9·4	8·8	8·4
May		29·2	20·5	15·0	11·4	9·2	8·0	7·2	6·8	6·5
June		36·5	25·0	16·2	10·3	7·6	6·1	5·4	5·0	4·8
July		25·0	17·8	13·2	10·2	8·4	7·3	6·7	6·3	6·1
August		23·9	16·6	12·4	9·5	8·5	7·6	7·0	6·7	6·5
September		33·4	20·4	15·1	12·4	10·7	9·7	9·2	8·8	8·4
October		23·9	18·2	15·2	13·6	12·5	12·0	11·6	11·3	11·2
November		25·4	20·6	17·8	16·5	15·7	15·1	14·6	14·3	14·1
December		21·4	18·0	16·6	15·6	15·0	14·4	14·0	13·7	13·4
1961										
January		20·4	17·8	16·6	15·8	15·2	14·9	14·7	14·4	14·4
February		14·6	13·1	12·1	11·5	11·1	10·8	10·5	10·4	10·3
March	23·8	16·7	13·8	12·2	11·2	10·6	10·1	9·8	9·5	9·4
April	22·6	16·1	12·8	10·9	9·8	9·1	8·6	8·3	8·1	8·0
May	21·9	15·8	11·8	9·4	7·7	6·9	6·2	6·0	5·8	5·6
June		13·2	9·8	8·1	7·0	6·7	6·5	6·4	6·3	6·2
July		15·2	10·4	8·2	7·0	6·3	5·8	5·7	5·3	5·2

satisfactory for all latitudes and that in addition to height gradient and latitude variation, seasonal and solar cycle variations of β should be computed individually for different locations. Such determination will require a large number of evaluations with a comparatively simple method utilising easily obtainable data such as vertical incidence data from the individual stations.

It is known that in almost all parts of the world except in the vicinity of the magnetic equator the magnitude of the vertical transport velocity and its height gradient are small and any loss of ionization by vertical divergence is, to a large extent, neutralized by downward plasma diffusion under gravity (Martyn 1955). Rishbeth and Barron (1960) have shown that the movement of ionization does not significantly alter the layer shape. It is, therefore, reasonable to assume that at temperate and sub-tropical latitudes, the electrons, at the peak of the $F2$-layer and below it, are in quasi-equilibrium condition around noon. Under this assumption it should be possible to obtain reasonable values of the relaxation time from profile data as well as from mean $F2$-layer peak electron densities in these latitudes. It has been shown by Appleton and Lyon (1955) that the response of the layer lags behind the ionizing radiation by an interval of time because of the finite value of the loss coefficient and that the value of this time interval depends upon the height considered. The relaxation time τ is related to the linear loss coefficient β by the relation: $\tau=1/\beta$.

We have attempted to establish that the reciprocal of the time displacement from local noon of maximum phase of the diurnal harmonic of electron densities at a given height yields a reasonable value of β. The accuracy of the results will naturally depend on the extent to which the diurnal harmonic is related to the solar ionizing radiation. From an analysis of Washington N_mF2 data from 1948 through 1957 we find that the correlation coefficient between the amplitude of the diurnal harmonic and Zürich sunspot number is 0·918. The solar radio radiation flux at λ10·7cm is also a fairly good index of solar ionizing radiation (Minnis and Bazzard 1959); the correlation coefficient between the diurnal harmonic amplitude and λ10·7 cm solar noise flux is 0·942. Besides, the values of the loss coefficient yielded by the reciprocal of τ are very consistent with earlier determinations. A plot showing the height variations of β at Puerto Rico (Geo. Lat. 18·5°N; Geomag. Lat. 30·0°N; Dip. 52·0°N) for April 1960 which has been obtained from harmonic analysis of 24 hourly values of electron densities at several fixed heights is shown in Fig. 1.

The simplicity of the method has enabled us to make a large number of determinations of the loss coefficient at different heights from electron density profile data. We have also extended the method for determining seasonal and solar cycle variations in the loss coefficient at an average height of maximum electron density at a few low and moderate latitude stations.

2. Analysis of electron densities at fixed heights

The average monthly electron densities at Puerto Rico for quiet ionosphere ($K_p \leqslant 4^+$) published in the CRPL-F (Part A)

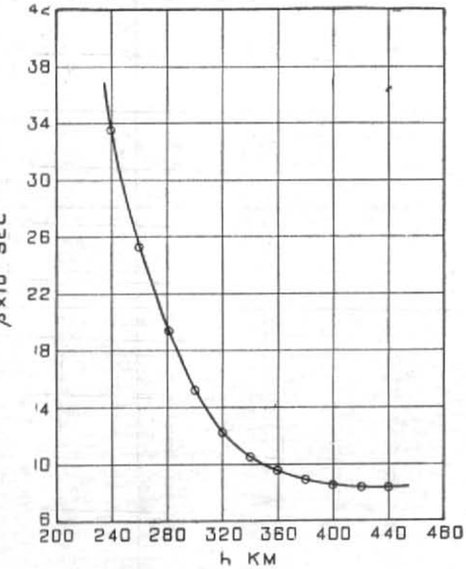

Fig. 1. Height variation of loss coefficient β at Puerto Rico, April 1960

Bulletins were harmonically analysed. The amplitudes and phases of the diurnal term were obtained every 20 km between 220 and 400 km. The process was carried out from the data covering 28 months from April 1959 through July 1961. The values of the loss coefficient at each one of the heights obtained from relaxation time monthwise are listed in Table 1. The magnitude and height variation characteristics are at once seen. The values of the loss coefficient are consistent with earlier determinations and show a systematic and smooth height variation as well as seasonal and solar cycle characteristics. The height gradient of the loss coefficient obtained by averaging the numerical values for each one of the heights for 28 months is shown in Fig. 2. It indicates that between 220 and 350 km the loss coefficient decreases exponentially with height and is approximately given by—

$$\beta_{(h)}=1{\cdot}46\times10^{-4}\ \exp.\left[\frac{300-h\,(\mathrm{km})}{110}\right]\mathrm{sec}^{-1}$$

A curve obtained from this expression is also plotted in Fig. 2 along with the values of β as a function of true height. The agreement

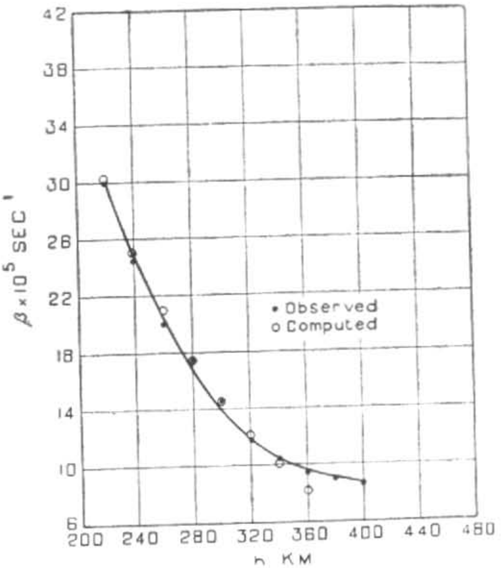

Fig. 2. Variation of loss coefficient β with height at Puerto Rico averaged over 28 months along with values computed from $\beta = 1 \cdot 46 \times 10^{-4}$ exp.[(300—h)/110] sec $^{-1}$

between the two is very close between 220 and 340 km. It is also observed that the scale height of β is considerably larger than Ratcliffe *et al.* value but it is close to that observed by Van Zandt *et al.* for Danger Island and by Mahajan and Mitra (1961) for Delhi (Geo. Lat. 28°38′N; Geomag. Lat. 19°11′N).

3. Seasonal characteristics in Loss Coefficient

It is known that one of the main causes for seasonal anomaly in the $F2$-region ionization is the seasonal change in the rate of electron loss. From nocturnal electron density variations, taking into account temperature changes, Yonezawa (1950) noticed marked seasonal change in the $F2$-layer recombination coefficient with a maximum ($\alpha = 3 \cdot 3 \times 10^{-10}$ cm^3 sec^{-1}) in December and minimum ($\alpha = 0 \cdot 3 \times 10^{-10}$ cm^3 sec^{-1}) in June and July. Several other investigations (Mitra 1951, Weiss 1953) also indicated larger values of the recombination coefficient for winter than for summer. The magnitudes of the winter to summer ratios were, however, widely different, from about 11 (Yonezawa) to 3 (Mitra). Analysis of extensive data from 25 stations (Weiss 1953) yielded a winter to summer ratio of over 4 for α. For decay by attachment process, Weiss found that the attachment coefficient β was independent of latitudes in all seasons and the value of β averaged over all latitudes varied from $2 \cdot 4 \times 10^{-5}$ sec^{-1} in winter to $1 \cdot 1 \times 10^{-5}$ sec^{-1} in summer. More recent investigations (Rishbeth and Setty 1961) suggest a seasonal variation of atmospheric composition which affects the rates of both electron loss and observable production and that these may result in different rates of electron decay in winter and summer. However, according to their suggestions, mixing of O and N_2 during the summer months and increased relative concentration of N_2 result in an increased loss rate. In winter, diffusive separation of O and N_2 results in a smaller loss coefficient partly from decreased relative N_2 concentration and partly due to increased rate of production because of increased O concentration. It will, therefore, be noticed that while earlier investigators concluded that seasonal changes consisted of increased β in winter as compared to summer, the suggestion of Rishbeth and Setty implies just the reverse. It has recently been suggested by King (1961) that in summer the overlapping effects of many travelling disturbances maintain a high value of β resulting in low $F2$ critical frequency and marked bifurcation of the layer and that β is low in winter when these disturbances are absent.

TABLE 2

True height (km)	Winter (Dec 1959) $R_z = 125$ Layer peak at 325 km	Summer (Jun 1960) $R_z = 122$ Layer peak at 370 km
	βx 10^5 cm^3 sec^{-1}	βx 10^5 cm^3 sec^{-1}
240	25·3	36·5
260	20·0	25·0
280	16·5	16·2
300	14·7	10·3
320	13·8	7·6
340	13·3	6·1

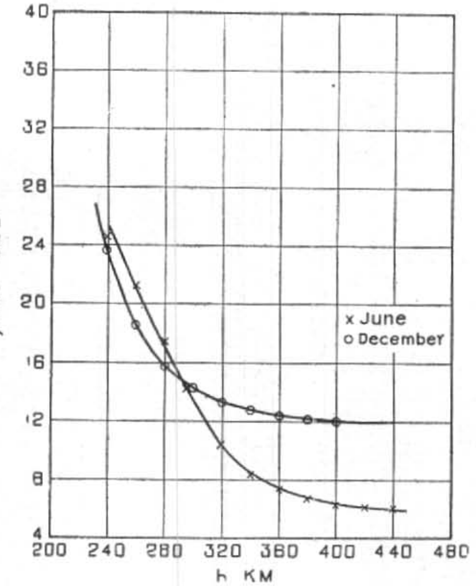

Fig. 3. Height variation of β at Puerto Rico during summer and winter

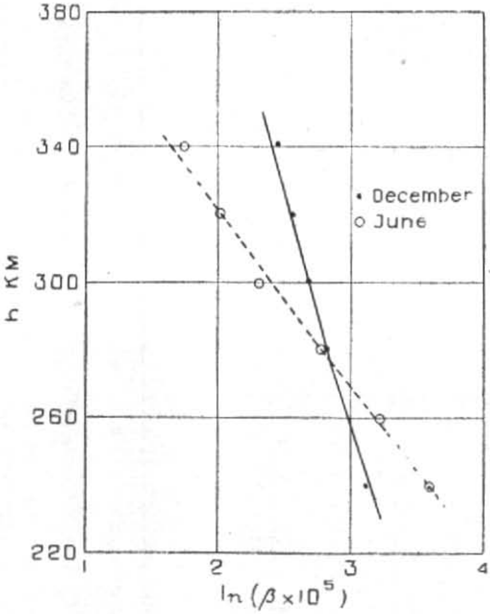

Fig. 4. Plot of ($\ln \beta$—h) for winter and summer conditions at Puerto Rico

The values of loss coefficient at Puerto Rico for 28 months listed in Table 1 indicate that considerable seasonal variations exist in the loss coefficient. For lower *F*2-region in the vicinity of about 250 km, the loss coefficient is slightly lower in winter than in summer; for higher heights, however, its magnitude is considerably higher in winter. Consequently the range in the variation of β for winter is much smaller than in summer. To illustrate the relative height changes in β in summer and winter, the values of the loss coefficient β for December 1959 and for June 1960, when the relative Zürich sunspot number was almost the same, are given in Table 2. From Fig. 3 where the averaged values of β are plotted as a function of height separately for winter and summer it will be noticed that while in winter the ratio $\beta_{(250)}/\beta_{(400)}$ is less than 2, in summer $\beta_{(250)}/\beta_{(400)}$ is almost 4.

It is obvious that β decreases with increase of height much more rapidly in summer compared to winter.

From plots of the loss coefficient for typical winter and summer months (December 1959 and June 1960) when R_z was almost the same we find that for heights over 300 km summer values of β are smaller than in winter and the height gradient of β in summer is larger than its gradient in winter. A plot of the height against $\ln \beta$ (Fig. 4) for these 2 months yields higher scale heights of β in winter than in summer.

It is known that in the *F*-region atomic oxygen is the ionizable constituent; molecular nitrogen is also ionized but does not contribute appreciably to the ionization as N_2^+ ions recombine rapidly by dissociative recombination. It is, therefore, the concentration of molecular gases that determines the value of β (Rishbeth and Barron 1960, Van Zandt *et al.* 1960). If the logarithmic decrement of β is the scale height of molecular gases O_2 or N_2, it means that the scale height of the molecular gases is larger in winter than in summer. The height gradients of β in summer and in winter shown in Fig. 4 indicate that the relative concentration of molecular gases decreases more rapidly with height in summer than in winter. However, in the lower

TABLE 3

$\beta \times 10^5$ sec^{-1}(Washington D.C., 1948 through 1960)

	1948	1949	1950	1951	1952	1953	1954	1955	1956	1957	1958	1959	1960
Jan	21·4	16·0	18·2	20·1	20·8	34·5	38·0	32·3	17·4	14·9	15·4	23·1	17·4
Feb	14·9	14·4	8·7	14·9	16·1	22·0	24·6	24·6	14·9	15·4	14·4	14·4	13·0
Mar	19·8	13·9	12·2	12·6	12·6	13·9	20·9	23·1	15·0	13·9	13·9	16·7	11·9
Apr	13·0	11·9	10·2	8·3	10·2	9·5	12·6	11·6	10·4	10·2	12·6	12·3	7·7
May	6·8	6·9	7·1	6·7	8·0	9·9	8·5	8·3	6·9	6·9	7·2	6·1	6·2
Jun	6·0	5·8	6·9	6·3	1·5	8·8	9·5	6·9	5·0	5·5	5·0	5·9	4·7
Jul	6·3	6·0	5·4	6·4	7·6	6·9	10·7	7·3	6·2	6·3	5·0	4·9	5·6
Aug	8·2	8·3	7·7	7·5	8·5	9·9	12·3	8·2	8·2	8·7	11·6	6·7	7·2
Sep	13·9	16·8	10·9	11·0	11·9	14·9	16·1	14·4	14·9	16·7	7·5	11·0	19·0
Oct	17·4	15·4	19·0	16·7	23·2	27·7	24·6	26·2	27·8	26·2	20·9	17·4	16·7
Nov	18·1	19·8	25·0	23·1	45·4	42·2	60·6	27·8	16·8	23·1	23·1	17·4	15·4
Dec	18·1	19·0	19·8	22·1	41·6	68·0	92·6	26·0	18·1	18·2	17·3	17·4	17·4

$F2$-region around the vicinity of 250 km there appears to be no substantial difference in the relative N_2 concentration during summer and winter.

4. Loss Coefficient and Solar Activity

The variation of the recombination coefficient for Washington D.C. (Geo. Lat. 38° 44′ N; Geomag. Lat. 50°·0 N; Dip. 71°·5N) over more than a solar cycle was considered by Yonezawa (1952) who found that the values were larger by a factor of about 10 in the years of sunspot minimum than those of sunspot maximum. Yonezawa mentioned that such a large change in the magnitude of α was difficult to understand on the basis of the recombination theory. Since the loss coefficient depends upon the scale height of some constituent of the atmosphere, any increase in the scale height with increase in solar activity such as envisaged by Bullen (1961) will reduce the magnitude of the loss coefficient. In order to examine if such variations could be obtained, Washington monthly median N_mF2 values were subjected to harmonic analysis for a period of 13 years from 1948 through 1960. The values of the loss coefficient β deduced from the relaxation times are given in Table 3.

A plot of yearly mean value of the loss coefficient for a mean height of maximum electron density against Zürich mean annual sunspot number R_z is given in Fig. 5. The essential features of the variation are the following : for a very low value of sunspot number (1954, $R_z \approx 4$) to a sunspot number of about 90, the decrease is exponential ; from a sunspot number of 90 there is practically no change in the magnitude of the loss coefficient. It is also found that the values are larger at sunspot minimum by a factor of only about 2. The nature of the changes in β obtained here can be visualized in terms of increased scale height with solar activity (Setty 1961) and, perhaps, of even different spectral composition of the ionizing radiation at sunspot maximum from that at minimum.

5. Seasonal and Solar Cycle changes of β in Equatorial Region

In order to examine how the loss coefficient in equatorial region and its seasonal and solar cycle characteristics compare with those at middle latitude stations like Puerto

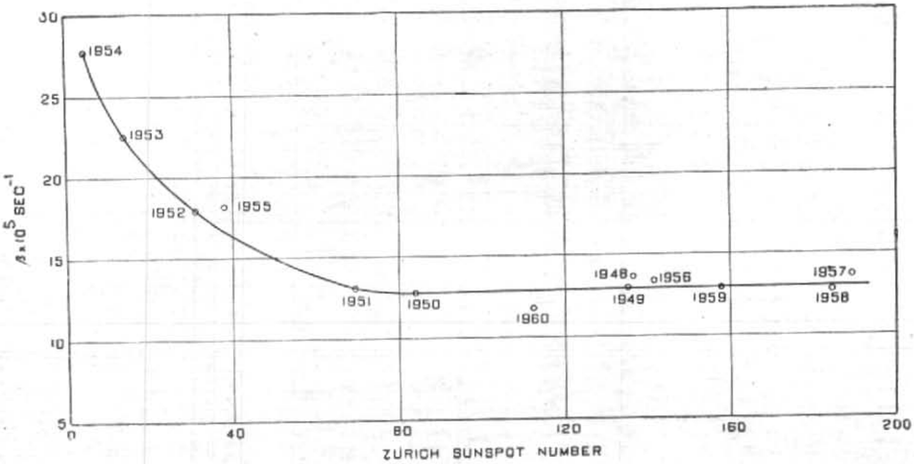

Fig.5. Variation of loss coefficient β with annual mean sunspot number at Washington.

Rico and Washington D.C., the relaxation times and the loss coefficient values for every month were obtained from Kodaikanal N_mF2 data for 6 years from 1956 through 1961. It was found that both the seasonal variation of β as well as the solar cycle changes were different for this station. In contrast to medium latitude stations, seasonal variation in the noon values of the virtual height of the $F2$-layer is known to have two maxima, one in the summer and the other in winter (Appleton 1950). The height of maximum electron densities has a similar variation and the scale heights can also, therefore, be assumed to vary in a manner similar to $h_{\max}F2$. At Kokaikanal β is found to vary inversely with cos χ as expected (Appleton 1952) but it has an equinoctial minimum and varies practically in the same way as $h_{\max}$ $F2$ contrary to the behaviour at temperate latitudes.

It is known that the magnitude of the height gradient of drift velocity increases with decrease in magnetic latitude and becomes appreciable near the dip equator (Martyn 1955). While in moderate latitudes the increased magnitude of the effective recombination coefficient or the attachment coefficient due to spreading of the layer from vertical divergence is offset by downward plasma diffusion, at low latitudes vertical diffusion is almost absent and, therefore, a high and thick layer results; the effective loss coefficient is, therefore, comparatively large in general due to contribution from the vertical divergence of ionization represented by $N\,(\partial v/\partial h)$. Some values of the loss coefficient deduced are given in Table 4 to illustrate this.

The seasonal variation, however, indicates that the loss coefficient is reduced during equinoxes at low latitudes and this appears to have a connection with increased geomagnetic activity at equinoxes. During these periods the magnitude of the drift velocity v and its height gradient $\partial v/\partial h$ are both reduced and the loss term of the continuity equation, *viz.*, $N\,(\beta+\partial v/\partial h)$ also undergoes a reduction. Our results thus agree with those of Skinner and Wright (1956) who found that on disturbed days the magnitude of the loss term at Ibadan was lower than on quiet days. As regards the solar activity influence on the loss coefficient at equatorial stations, it has been found from our analysis of Kodaikanal data, that, contrary to what is observed at middle latitude stations like Washington

TABLE 4

Station	Month and season	$h_{max}F2$ (km)	$\beta \times 10^5$ sec^{-1}	Period of data analysed
Washington, D.C.	Jun (Summer)	400	6·0	1955 Sep-1961 Aug (6 yrs)
Watheroo	Dec (Summer)	400	5·0	1955 Sep-1958 Aug (3 yrs)
Kodaikanal	Jun (Summer)	450	18·0	1955 Sep-1961 Aug (6 yrs)
Huancayo	Dec (Summer)	480	22·0	1955 Sep-1958 Aug (3 yrs)

D.C., at low latitudes the magnitude of effective β increases with an increase in solar activity. These anomalous seasonal and solar cycle variations in β suggest that the technique used for evaluation of the loss coefficient is not satisfactory for low latitudes because of the large magnitude of vertical divergence.

6. Acknowledgements

The authors wish to express their thanks to Shri S. N. Mitra, Research Engineer, All India Radio for his helpful criticism and Shri K. Thanikachalam for his help in computations.

REFERENCES

Appleton, E. V.	1950	*J. atmos. terr. Phys.*, **1**, p. 106.
	1952	Proc. 3rd Meeting, Mixed Commission on the Ionosphere, p. 29. I.C.S.U., Canberra.
Appleton, E. V. and Lyon, A. J.	1955	*The Physics of the Ionosphere*, p. 20, Phys. Soc., London.
Bowhill, S. A.	1961	*J. atmos. terr. Phys.*, **21**, p. 272.
Bullen, Jean M.	1961	*Nature*, **192**, p. 542.
Hirsh, A. J.	1959	*J. atmos. terr. Phys.*, **17**, p. 86.
King, G. A. M.	1961	*J. geophys. Res.*, **66**, p. 4149.
Mahajan, K. K. and Mitra, A. P.	1963	Proc. I.G.Y. Symposium, Vol. I, p. 57, C.S.I.R., New Delhi.
Martyn, D. F.	1955	*The Physics of the Ionosphere*, p. 254, Phys. Soc., London.
Minnis, C. M.	1956	*Solar Eclipses and the Ionosphere*, p. 81, Pergamon Press, London.
Minnis, C. M. and Bazzard, G. H.	1959	*J. atmos. terr. Phys.*, **14**, p. 213.
Mitra, A. P.	1951	*Ibid.*, **1**, p. 286.
	1959	*J. geophys. Res.*, **64**, p. 733.
Ratcliffe, J. A., Schmerling, E. R., Setty, C. S. G. K. and Thomas, J. O.	1956	*Phil. Trans.*, A. **248**, p. 621.
Rishbeth, H. and Barron, D. W.	1960	*J. atmos. terr. Phys.*, **18**, p. 234.
Rishbeth, H. and Setty, C. S. G. K.	1961	*Ibid.*, **20**, p. 263.
Savitt, J.	1950	*J. geophys., Res.*, **55**, p. 385.
Setty, C. S. G. K.	1961	*J. Inst. Telecom. Engrs.*, **7**, p. 65.
Skinner, N. J. and Wright, R. W.	1956	*J. atmos. terr. Phys.*, **7**, p. 105.
Van Zandt, T. E., Norton, R. B. and Stonehocker, G. H.	1960	*J. geophys. Res.*, **65**, p. 2003.
Weiss, A. A.	1953	*J. atmos. terr. Phys.*, **4**, p. 175.
Yonezawa, T.	1950	Proc. 2nd Meeting, Mixed Commission on the Ionosphere, p. 206, I.C.S.U., Brussels.
	1952	Proc. 3rd Meeting, Mixed Commission on the Ionosphere, p. 29, I.C.S.U., Canberra.